Chapter 5

Aridic Soils, Patterned Ground, and Desert Pavements

John C. Dixon

Introduction

Pedogenic and geomorphic processes operating in deserts are inextricably linked. These linkages are particularly well expressed in the development of patterned ground and desert pavement. In addition, the nature and efficacy of hydraulic, gravitational, and aeolian processes on desert surfaces are strongly influenced by the physical and chemical characteristics of the underlying soils. As a result, the evolution of a diversity of desert landforms is either directly or indirectly linked to pedogenic processes.

The significant role played by pedogenic processes in desert landscape evolution is also strongly reflected in the development of duricrusts – indurated accumulations of calcium carbonate, gypsum, and silica. These essentially pedogenic materials impart considerable relief in the form of tablelands, mesas, and buttes to otherwise low-relief landscapes. Erosion of these topographic eminences in turn provides the coarse debris for the formation of patterned ground and desert pavement.

In this chapter the nature and genesis of soils in arid and semi-arid environments are examined. This discussion is followed by an examination of the major types of patterned ground and the processes responsible for its formation. Finally, the nature and origin of desert pavements are examined. The nature and origin of duricrusts are discussed in Chapter 6.

J.C. Dixon (✉)
Department of Geosciences, University of Arkansas,
Fayetteville, AR 72701, USA
e-mail: jcdixon@uark.edu

Aridic Soils

Distribution

Desert soils occupy approximately 46.1 M km^2, or 31.5% of the Earth's surface (Dregne 1976). They are most widespread in Australia, occupying some 82% of the continental surface (Stephens 1962), and are also the dominant soils of Africa, covering some 59% of the continent. Arid soils cover substantially smaller areas of the remaining continents: 33% of Asia, 18% of North America, 16% of South America, and just under 7% of Europe (Dregne 1976). Desert soils are dominated by five soil orders (U.S. Comprehensive Soil Classification): Entisols (41% of arid zone), Aridisols (36%), Mollisols (12%), Alfisols (7%), and Vertisols (4%) (Dregne 1976). These soil orders are subdivided into thirteen suborders discussed in greater detail later in this chapter.

Characteristics

Aridic soils display a variety of distinctive morphological features. Among these are the presence of gravel -covered surfaces, the development of surface organic and inorganic crusts, and the widespread occurrence of vesicular A horizons. In addition they commonly display the development of ochric and mollic epipedons and are characterized by the formation of a variety of diagnostic subsurface horizons including cambic, argillic, calcic, petrocalcic, gypsic, petrogypsic, natric, salic, and duripan horizons (Buol et al. 1997, Southard 2000).

A.J. Parsons, A.D. Abrahams (eds.), *Geomorphology of Desert Environments*, 2nd ed.,
DOI 10.1007/978-1-4020-5719-9_5,

Surface Crusts

Desert soils commonly display the presence of thin (10–20 mm) crusts on surfaces largely devoid of vegetation and gravel cover. The crusts are typically massive but may possess a platy structure in their upper part. The lower portion of the crusts is commonly vesicular where it is in contact with the underlying vesicular horizon. Crusts typically display low permeability and accompanying enhanced runoff (Buol et al. 1997, Schaetzl and Anderson 2005), however some studies have suggested that some crusts (especially organic-rich crusts) may in fact enhance infiltration (Gifford 1972, Blackburn 1975, Dunkerley and Brown 1997). Crusting is widely believed to be the result of repeated wetting and drying of predominantly loamy soils (Buol et al. 1997). These authors believe that during wetting, soil plasma moves to contact points between skeletal grains where upon drying it acts as a weak reversable cement. The source of the binding agents (the plasma) has long been elusive (Sharon 1962, Schaetzl and Anderson 2005). For a long time it was widely believed that crusting was the result of raindrop impact (McIntyre 1958a, b, Schaetzl and Anderson 2005) which both compacted soil particles and produced and transported clay size particles into voids. However crusting probably includes both inorganic and organic processes and components. Inorganic components include clays (Frenkel et al. 1978, Benhur et al. 1985, Dunkerley and Brown 1997) which serve as binding agents, exchangeable sodium (Painuli and Abrol 1986) which results in soil dispersion, and the presence of small quantities of salts and calcium carbonate (Sharon 1962, Benhur et al. 1985).

Desert soil crusts are now generally attributed to biological processes resulting in the production of thin microphytic layers on the soil surfaces (Dunkerley and Brown 1997, Schaetzl and Anderson 2005). Microphytic crusts are dominated by assemblages of algae, mold, cyanobacteria, (Schaetzl and Anderson 2005) as well as mosses, liverwarts, lichen, bacteria, and fungi (Dunkerley and Brown 1997). These organisms grow in the upper few millimeters of the soil surface after rain events, and form biomantles in which their mycelia bind soil partices together. The soil mantles are commonly enriched in C and N as well as silt and clay (Fletcher and Martin 1948).

Vesicular Horizons

Immediately beneath the gravel surface layer of many aridic soils a vesicular Av horizon (Fig. 5.1) commonly occurs and is often associated with a surface crust. Formation of the vesicular horizon has been widely attributed to the saturation of the fine grained soil surface horizon (Miller 1971). Nettleton and Peterson (1983) argued that in the saturated state the soil plasma of this upper horizon is free to move and in so doing traps air. With repeated episodes of saturation and air entrapment, the vesicles in the surface horizon increase in size. Repeated destruction of the vesicular horizon is thought to be caused by wetting and drying cycles (Springer 1958, Miller 1971). Nettleton and Peterson (1983) alternatively suggested that vesicular horizon destruction is primarily the result of soil trampling by animals. Upon destruction, the formation of the vesicular horizon begins anew as soil saturation episodes begin and air entrapment resumes.

A substantially different explanation for the origin of the vesicular horizon has been proposed by Wells et al. (1985), McFadden et al. (1986, 1987, 1998) and Blank et al. (1996). These workers attributed the origin of the fine-grained surface horizon and accompanying vesicular structure to aeolian addition of fine grained materials and associated soluble salts, carbonates, and iron oxides. The development of vesicles is attributed to entrapment of air by aeolian infall with subsequent expansion due to heating following summer rainfall events. This model follows that of Evenari et al. (1974) for vesicular horizons developed in soils in Israel.

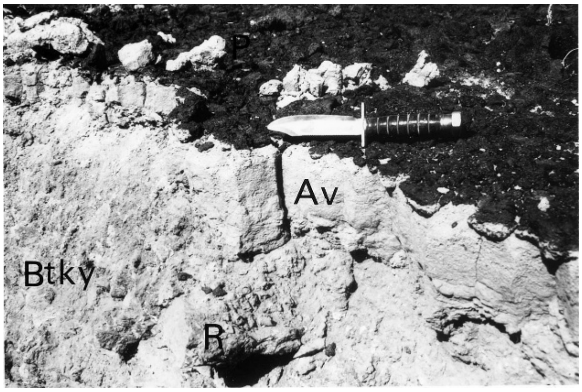

Fig. 5.1 Soil developed beneath desert pavement showing vesicular A horizon (Av) and carbonate and gypsum enriched B horizon (Btky). Bedrock rubble is designated (R) (photo courtesy of L.D. McFadden)

Fig. 5.2 Columnar structure of vesicular A horizon resulting from high shrink–swell capacity (photo courtesy of L.D. McFadden)

Supporting evidence for an aeolian sediment source includes the predominantly silt and clay dominated texture of the Av horizon (McFadden et al. 1987, 1998, Anderson et al. 2002), the presence of abundant amounts of salts and carbonate which exceed those in parent materials (Reheis et al. 1995), and the presence of mineralogies which differ from those of the soil parent materials (Blank et al. 1996). The subsequent stability of the vesicular horizon is attributed largely to the formation of thin $CaCO_3$ coatings (Evenari et al. 1974) and clay coatings (Sullivan and Koppi 1991). While aeolian infall is envisaged as the primary source of the fine grained surface layer, McFadden et al. (1987) also suggested that some of the sand and silt fraction may be the result of mechanical weathering of the surface gravels. The presence, and successive accumulation, of clay in the aeolian mantle leads to the development of a distinctive columnar structure in the Av horizon due to enhanced shrink–swell capacity (McFadden et al. 1986) (Fig. 5.2). Sullivan and Koppi (1991), working on desert loams in Australia, have also proposed that the fine grained materials in the vesicular horizon are externally derived. They suggested a combination of aeolian, colluvial, and overland flow sources.

Cambic Horizons

Intimately associated with the vesicular, ochric epipedons are cambic horizons. These horizons are found below ochric epipedons and are typically reddish or brownish in colour. There is evidence of alteration in the horizon in the form of obliteration of original structures due to mixing, development of new structure, accumulation of clay, and carbonate translocation through and into the horizon (Nettleton and Peterson 1983, Buol et al. 1997). The source of the fine material and the carbonates in this horizon is widely attributed to aeolian infall (Gile 1970, 1975, Gerson et al. 1985, McFadden et al. 1986, Gerson and Amit 1987, Reheis 1987a, Wells et al. 1987, Birkeland 1990). McFadden et al. (1986) suggested that as the vesicular horizon thickens due to continued aeolian addition, plasma migration near the base of the horizon results in destruction of vesicles and a slowing of the migration of infiltrating waters. This slowing of infiltration results in ferrous iron alteration as well as the accumulation of authigenic ferric iron oxyhydroxides. Soil reddening results, and a cambic horizon slowly develops. Clay and carbonate added by aeolian infall also migrate to the cambic horizon.

Argillic Horizons

Many aridic soils are characterized by the occurrence of strongly developed argillic (clay-rich) horizons. This horizon differs markedly from the aforedescribed cambic horizon in that it is substantially more abundant in clay and commonly has a noncalcareous matrix. Research on the origin of the argillic horizon has focused on two contrasting sources of clay. Smith and Buol (1968) and Nettleton et al. (1975) have shown that the degree of mineral weathering in A horizons of some argids in the south-western United States is comparable to, or greater than, that in underlying Bt horizons. This finding therefore suggests that chemical weathering in A horizons is a possible source of illuvial clay. Further, these authors have pointed to the occurrence of clay skins on mineral grains and pebbles in the argillic horizon as evidence for clay illuviation. The frequent lack of clay skins on ped faces is attributed by Nettleton and Peterson (1983) to destruction by swell–shrink mechanisms, bioturbation, and calcium carbonate crystal growth. Nettleton et al. (1975) also argued for clay illuviation on the basis of patterns of distribution of fine and coarse clay and total iron and aluminium down the profile. Finally, Nettleton and Peterson (1983) reasoned that the occurrence of clay skins on the walls of deep pipes or as downward extensions of the argillic horizon support an illuvial origin for the clay. Chartres (1982), on the other hand, has stressed the importance of the aeolian addition of fine debris in the formation of argillic horizons, citing a relative lack of weathered soil particles in the A horizon as evidence against an in-profile source of clay. Blank et al. (1996), suggest a multi-stage model of argillic horizon formation involving aeolian addition/weathering/translocation. Based on chemical, mineralogical and micromorphological evidence, they suggest the neoformation of secondary clay minerals resulting from the weathering of aeolian–derived sediment followed by subsequent translocation of the neoformed clays to depth within the soil.

Carbonates are commonly associated with the argillic horizon. In some cases they may dominate the soil matrix (Gile 1967), but more commonly they occur as nodules which have displaced the previously deposited clays (Gile and Grossman 1968, McFadden et al. 1986). Accumulation of carbonates in argillic horizons is widely interpreted to be indicative of significant climate change from humid to arid climatic regimes with an accompanying reduction in depth of leaching (Nettleton and Peterson 1983, McFadden and Tinsley 1985, McFadden et al. 1986, Reheis 1987b, Wells et al. 1987). Thick argillic horizons in Aridisols are therefore widely regarded as being largely relics of the Pleistocene.

Natric and Salic Horizons

The accumulation of soluble salts is a further characteristic of aridic soils. Accumulation results in the formation of either natric (sodic) or salic (saline) horizons. Natric horizons are argillic horizons characterized by high (≥15%) exchangeable sodium or a sodium adsorption ratio of 13 or more and by the development of prismatic or columnar soil structure (Soil Survey Staff 1975, Dregne 1976).Magnesium salts are also commonly present and on occasion may exceed the abundance of exchangeable sodium, but their presence is not mandatory for the classification of natric horizons (Dregne 1976). In Holocene-age soils these horizons are typically thin, while in Pleistocene soils they thicken considerably (Nettleton and Peterson 1983). These horizons typically occur above or in a zone of calcium carbonate accumulation (Nettleton and Peterson 1983). Although there is sodium accumulation, there is insufficient for these horizons to be classified as salic. In order for a horizon to be classified as salic it must be at least 15 cm thick, be enriched in salts that are more soluble than gypsum, contain at least 2% soluble salt and have a product of salt content times horizon thickness greater than 60 (Soil Survey Staff 1975, Dregne 1976).

Salts in natric and salic horizons appear to be derived from several sources. Airborne addition of salts has been suggested for these horizons in aridic soils in the United States by Alexander and Nettleton (1977), Peterson (1980), Nettleton and Peterson (1983), and McFadden et al. (1991). An airborne source of salts has also been suggested for saline soils in Israel (Yaalon 1964, Yaalon and Ganor 1973, Dan et al. 1982, Amit and Gerson 1986). The source of salts in sodic soils in Australia has also been widely attributed to airborne sources, particularly for those soils within about 200 km of the coastal environment (Hutton and Leslie 1958, Wetselaar and Hutton 1963, Chartres 1983, Isbell et al. 1983) as well as from the mobilization of salts from playa surfaces in arid and semi-arid continental interiors.

A second source of the sodium is from the incursion of very shallow ground waters. Nettleton and Peterson (1983) suggested such an origin for some of the natric horizons in soils in the United States. These workers also proposed that this process is particularly important in the formation of salic horizons. This process appears to be especially significant for soils occurring in the Riverine Plain of south-eastern Australia (Northcote and Skene 1972, Isbell et al. 1983). Detailed discussion of salinization of soils in Western Australia by saline ground waters is provided by Conacher (1975). Secondary salinization as a result of groundwater pumping for irrigation is a growing problem in some arid Australian soils. Similarly, poor irrigation-agriculture practices has led to the development of sodic solis due to groundwater infiltration in Argentina, Peru, and Iraq (Dregne 1976).

The third explanation for the formation of natric or salic horizons is the inheritance of salts from parent materials. Three parent-material sources have been suggested by workers from Australia. Chartres (1983) suggested that sodic horizons developed well below the depth of contemporary infiltration may well have originated in older sodic soils which have subsequently been eroded. A number of studies of sodic soils from western Queensland strongly suggest that many have derived their sodium from the underlying bedrock (Gunn and Richardson 1979, Isbell et al. 1983). In particular, soils developed on marine shales and gypsum or anhydrite beds are highly susceptible to salt accumulation as are obviously those developed directly on pre-existing salt deposits (Dregne 1976). In a related fashion, some sodic soils derive their salts from underlying deeply weathered parent materials which have in turn been derived from salt-rich bedrock. Sodic soils developed on laterite in Western Australia have been reported by Bettenay et al. (1964) and Dimmock et al. (1974). The significance of the deep sodic horizons as sources of secondary salinization have been investigated by Peck and Hurle (1973) and Peck (1978). Sodic soils developed in deeply weathered materials have also been reported from Queensland by Hubble and Isbell (1958) and Isbell (1962). More recently, Gunn (1967) has suggested that deep weathering profiles are also the source of salts in solodic and solodized–solonetz soils in central Queensland.

Salts may also be derived from surface water runoff, especially where those waters accumulate in depresssions in the landscape such as in salt pans, salinas, sebkhas etc. In such settings slow rates of infiltration coupled with high rates of evaporation result in salt accumulation. Salic and natric horizons may also develop in soils which are affected by interconnected surface and groundwater sources as is the case for soils in the Dasht-i-Kavir in Iran (Dregne 1976).

Finally, some saline horizons may form as a result of the accumulation of salts from decomposing vegetation. Such salts represent accumulations of recycled salts from deeper layers within the soil which pass through halophytes and accumulate in soil surface horizons. Such salt accumulations have been reported from the western United states by Wallace et al. (1973).

Calcic and Gypsic Horizons

Aridic soils are frequently dominated by calcium carbonate and/or gypsum. With progressive concentration of carbonate or gypsum and accompanying induration, petrocalcic and petrogypsic horizons develop. These horizons are collectively referred to as duricrusts, which are discussed fully in Chapter 6. Calcic and gypsic soils are characterized by the occurrence of non-indurated accumulations of calcite and/or dolomite in the former and gypsum in the latter. These materials occur in a variety of forms including powdery fillings, nodules, pendents, or crusts beneath pebbles and cobbles. Crusts in calcic and gypsic soils slake in water. In the United States, calcic soil horizons must be 15 cm or more thick, contain the equivalent of $\geq$15% $CaCO_3$, and have a carbonate content $\geq$5% greater than the C horizon. Gypsic soil horizons must also be 15 cm or more thick, contain at least 5% more gypsum than the underlying horizon, and have a product of horizon thickness in centimetres and percentage gypsum content not less than 150 (Soil Survey Staff 1994).

Descriptions of calcic and gypsic soils are common in the desert soils literature. It is important to point out that many aridic soils contain intergrades of calcic–gypsic horizons. A comprehensive discussion of gypsic soils (and intergrades) is provided by Boyadgiev and Verheye (1996) in which they identify the geographic extent, geomorphic and climatic settings, pedogenic characteristics and classification of gypsosols. Calcic soils in the United States have been extensively investigated by Bachman and Machette (1977), Shlemon (1978), Gile et al. (1981), Nettleton and Peterson (1983), Machette (1985), Weide (1985), Reheis et al. (1989), Harden et al. (1991), McFadden et al. (1991), Monger et al. (1991), Rabenhorst et al. (1991), and Nettleton et al. (1991). Studies of gypsic soils in the United States are notably fewer but include those ofNelson et al. (1978), Nettleton et al. (1982), and Harden et al. (1991). Detailed descriptions of gypsic soils in Wyoming have been provided by Reheis (1987a) Figs. 5.3, 5.4 and 5.5). Studies of calcic soils from Australia have been summarized by Stace et al. (1968) and Northcote and Skene (1972), Hutton and Dixon (1981), Milnes and Hutton (1983), Isbell et al. (1983), and Akpokodje (1984). Early descriptions of gypsic soils in Australia were provided by Prescott (1931), Jessop (1960) and Bettenay (1962). Calcic and gypsic soils have been reported from many other arid and semi-arid environments, some of which were summarized by Eswaran and Zi-Tong (1991). These authors reported the extensive occurrence of gypsic soils in China, India, Pakistan, the Middle East,

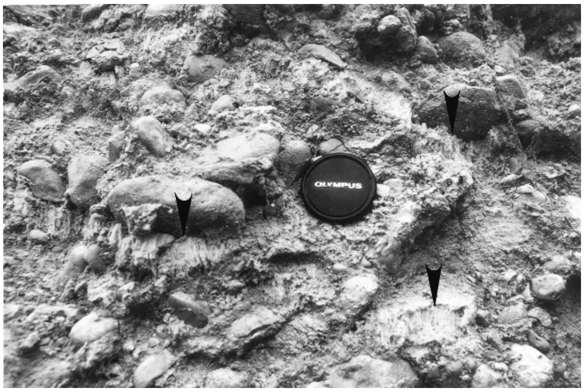

Fig. 5.3 Stage I gypsum (arrows) in soil developed on alluvial fan, Bighorn Basin, Wyoming (photo courtesy of M. Reheis (U.S. Geological Survey Bulletin 1590-C)

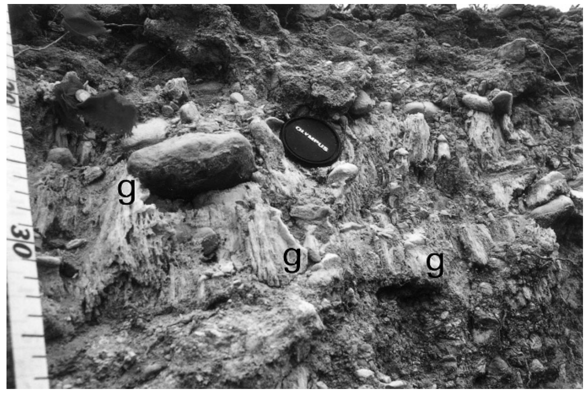

Fig. 5.5 Stage IV gypsum (g) in soil developed on alluvial fan, Bighorn Basin, Wyoming (photo courtesy of M. Reheis (U.S. Geological Survey Bulletin 1590-C))

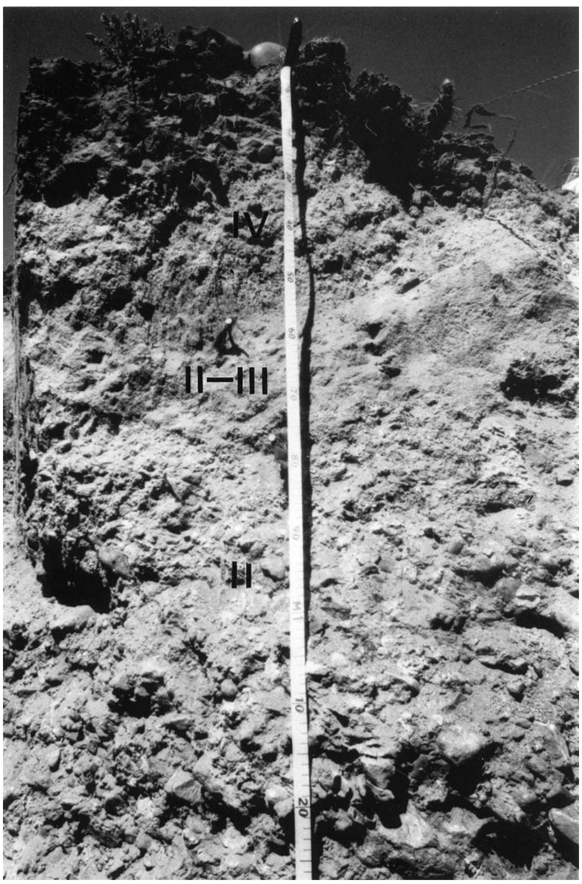

Fig. 5.4 Complex gypsic soil developed on alluvial fan, Bighorn Basin, Wyoming. Stage IV gypsum 20–50 cm, Stage II/III gypsum 50–75 cm, Stage II gypsum 75+ cm (photo courtesy of M. Reheis (U.S. Geological Survey Bulletin 1590-C))

and Africa. More limited occurrences of gypsic soils were recorded in southern Europe. Dan (1983) and Dan et al. (1982) described gypsic soils from Israel.

The source of the gypsum and carbonate in aridic soils is widely believed to be from atmospheric addition as either dust or carbonate dissolved in rainwater (Bull 1991, Dohrenwend et al. 1991, Gustavson et al. 1991). However, some studies such as those by Akpokodje (1984) suggest an *in situ* origin of the carbonate and gypsum. Boettinger and Southard (1991) have provided compelling evidence from Aridisols developed on pediments in the Mojave Desert of California that the source of carbonate deep within calcareous Haplargids is derived from the weathering of the granite and accompanying release of calcium from plagioclase. Hutton and Dixon (1981) provided evidence that at least some of the carbonates in southern south Australia were derived from pre-existing dolomite-rich laccustrine parent materials. Other sources include capilliary rise from shallow groundwater sources and from surface water runoff. Additionally, biological processes may play significant roles in calcic and gypsic horizon development. The mineralization of dead plant material has been suggested by Dregne (1976) to contribute significant quantities of carbonate and gypsum to soils. Amit and Harrison (1995) have shown that calcic horizons in Israeli soils are intimately associated with fungal hyphae.

Duripans

Many arid soils possess a duripan. These pans are subsurface indurations of predominantly silica, but may grade to petrocalcic (calcrete) horizons (Soil

Survey Staff 1975, Nettleton and Peterson 1983). In gross morphology, chemistry, and mineralogy they are analogous to silcretes described in detail in Chapter 5. Duripans are generally platy in structure with individual plates ranging in thickness from 1 to 15 cm. Pores and surfaces of the plates are commonly coated with opal, chalcedony, and/or microcrystalline silica. For a soil horizon to be classified as a duripan, at least one of three criteria must be met. The soil must display (a) some vertical coatings of opal, (b) siliceous nodule development, or (c) the development of silica pendents on the undersides of coarse fragments. In addition, opal or other forms of silica must partly fill interstices and form bridges between sand grains (Soil Survey Staff 1975).

Duripans in the United States appear to be most strongly developed in soils formed on volcanic ash or other pyroclastic parent materials containing abundant silica (Flach et al. 1969, 1973, Brasher et al. 1976, Nettleton and Peterson 1983). Boettinger and Southard (1991) examined duripans from a Durorthid pedon developed in grus on a pediment in the Mojave Desert. These workers suggested that the source of the silica is not volcanic glass, which is essentially absent from the profile, but the weathering of feldspars. In Australia, Stace et al. (1968) reported hardpan soils commonly occurring on strongly weathered alluvial and colluvial deposits as well as on eroded laterites. Duripans are most strongly developed on older landscape surfaces where there has been sufficient time for prolonged silica dissolution, translocation, and deposition. They form at the average depth of wetting, which progressively diminishes with increasing aridity. Duripans commonly form below or in the lower part of argillic or natric horizons and have also been observed to be interlayered with illuvial clay (Nettleton and Peterson 1983). In Australia, hardpan soils have been reported from Western Australia by Litchfield and Mabbutt (1962), Stace et al. (1968), and Brewer et al. (1972), from South Australia by Stace et al. (1968), Wright (1983), and Milnes et al. (1991), and from New South Wales by Chartres (1985).

Classification

As indicated previously, the dominant soil order of the arid regions of the world are entisols. Within this order three suborders are widely identified in arid environments: Fluvents, Orthents, and Psaments. Fluvents occur on floodplains, deltas, and alluvial fans where drainage is not impeded. Orthents occur where diagnostic surface horizons have been eroded or occur on gypsiferous surfaces with deep water tables. Typically they are associated with rocky slopes (Dregne 1976). Psaments are Entisols which are sandy in all subhorizons below an Ap horizon or below a depth of 25 cm to a depth of 1 m. They contain less than 35% gravel; gravelly psaments are classified as Orthents. Psaments are typically associated with stabilized and unstabilized sand dunes, cover sands and sand sheets. Entisols occur extensively on landsurfaces of Holocene age in the southwestern United States (McFadden 1988, Bull 1991). The dominant zonal soil order of the arid lands are aridisols. This soil order is dominated by two suborders, Argids and Orthids. Argids are characterized by the development of argillic or natric horizons while Orthids lack an argillic or natric horizon.

Each of these suborders can be further divided into a number of great groups. The principal great groups of Argids are Durargids, Haplargids, Natrargids, Nadurargids, and Paleargids (Guthrie 1982). Durargids possess a duripan that underlies an argillic horizon and lack natric horizons. The top of the duripan is within a metre of the ground surface. Haplargids lack a natric horizon and do not have a duripan or petrocalcic horizon within a metre of the surface. They have weakly developed argillic horizons with less than 35% clay accumulation. Natrargids are characterized by the occurrence of a natric horizon but lack a duripan or petrocalcic horizon within a metre of the ground surface. Nadurargids are Argids that possess a natric horizon above a duripan, and the surface of the duripan is within a metre of the ground surface. Paleargids are Argids that develop on old land surfaces. They are characterized by either a petrocalcic horizon below an argillic horizon or an argillic horizon with greater than 35% clay and an abrupt upper boundary (Soil Survey Staff 1975).

Orthids are aridic soils with one or more pedogenic horizons. However, they lack a natric or argillic horizon. They commonly have an accumulic horizon of soluble salts and calcium carbonate. Some orthids possess salic, calcic, gypsic, petrocalcic, petrogypsic, cambic, and duripan horizons (Soil Survey Staff 1975). Orthids are divided into six great groups: Calciorthids, Camborthids, Durorthids, Gypsiorthids, Paleorthids, and Salorthids. Calciorthids contain abundant amounts of lime derived from either the parent material or added from aeolian dust. Camborthids are characterized by the development of a cambic horizon that results in a brownish to reddish brown soil of uniform texture. Durorthids possess a duripan within a metre of the surface and are commonly calcareous throughout. Gypsiorthids contain a gypsic or petrogypsic horizon with an upper boundary within a metre of the soil surface. Paleorthids contain a petrocalcic horizon within a metre of the ground surface. They commonly display evidence of calcium carbonate engulfing a pre-existing argillic horizon. Salorthids are very salty soils commonly associated with the accumulation of salts from capilliary rise of salty waters. They are characterized by a salic horizon (Soil Survey Staff 1975).

In one of the most comprehensive studies of arid lands soils in the United States, the Desert Project (Gile et al. 1981) recognized that large parts of The Desert Project study sites in New Mexico are also occupied by Mollisols, Alfisols, and some Vertisols. Two suborders of Mollisols are widely recognized in arid regions: Ustolls, and Xerolls. Ustolls occur in regions with semi-arid and subhumid climates. They are free draining and show no evidence of saturation during the year. They possess ustic or aridic moisture regimes. The presence of lime or gypsum indicates that they are dry for extended periods of time through the year. Xerolls occur under environments of contrasing seasonal precipitation. Mean annual soil temperature is less than 22°C and seasonal soil temperatures differ by at least 5°C at a depth of 50 cm. Internal drainage is sufficient to preclude the development of mottles and Fe-Mn nodules larger than 2 mm in diameter. Xerolls have a xeric or aridic moisture regime. Vertisols are predominantly Usterts which are characterized by the presence of cracks that remain open for 90 consecutive days or more, but are closed for at least 60 consecutive days when soil temperature is above 8°C at 50 cm., mean annual soil temperature is greater than 22°C, and displays a seasonal difference of less than 5°C. Alfisols are dominated by Ustalfs and Xeralfs. Ustalfs possess a ustic moisture regime and thermal regimes that are at least thermic, but more commonly mesic, or isomesic. Calcic horizons commonly occur below the argillic horizon. Xeralfs possess moisture regimes that are xeric with extended dry summers and moist winters. Soil temperatures are typically thermic (Dregne 1976).

Aridic Soils and Landscape Development

Research on aridic soils in the United States in recent years has focused heavily on the relationship between soil development and landscape evolution. These studies have been concerned with the age of landscapes, correlation of Quaternary deposits, and climate change. The use of aridic soils as relative age indicators is well illustrated in studies from the western and south-western United States (Christianson and Purcell 1985, Harden et al. 1985, Machette 1985, Ponti 1985, Dohrenwend et al. 1991).

Detailed studies of soil development as an indicator of environmental change have been undertaken by many workers in the south-western United States. Major differences in aeolian dust addition to soils between the Holocene and Pleistocene are reported by Machette (1985), McFadden et al. (1984, 1986, 1987), Wells et al. (1987), Chadwick and Davis (1990), Reheis (1987a, b, 1990), and Bull (1991). Studies of carbonate accumulation amounts and rates, and depth of infiltration have been undertaken by Machette (1985), McFadden and Tinsley (1985), Reheis (1987a, b), and Bull (1991). These studies all point to maximum carbonate accumulation during the Holocene and accompanying reduction in the depth of carbonate infiltration. The polygenetic nature of many aridic soils is also highlighted in studies by Reheis (1987a, b) Nettleton et al. (1989), and Bull (1991). Variability in rates of soil formation depending on geomorphic setting and climate is stressed by Reheis et al. (1989) and Bull (1991). Recent more detailed studies by some of these same workers from California and Nevada (McFadden et al. 1998, Reheis et al. 1995) have established that accelerated dust deposition at the Pleistocene-Holocene boundary resulted in a period of active, rapid soil formation. McFadden et al. (1998) demonstrate further the intimate relationship between aeolian dust deposition, carbonate accumulation, vesicular soil horizon development and desert pavement formation.

Soils in arid environments in the western United States have also been widely used as stratigraphic markers and as indicators of periods of stability within Quaternary depositional systems. Such studies are well exemplified by the work of Gustavson and Winkler (1988), Gustavson et al. (1991), and Holliday (1985a, b, c, 1988, 1989, 1990) from the southern High Plains of Texas and New Mexico. Similar studies in aeolian environments in the southern Colorado Plateau by Wells et al. (1990) and California/Nevada (Reheis et al. 1995, McFadden et al. 1998) further demonstrates the importance of soil formation in understanding landscape evolution in desert environments.

Patterned Ground

Patterned ground is a common and widespread feature of desert surfaces. It includes a variety of forms including gilgai, surface cracking, microtopography related to surface crust formation, and vegetation patterning. At a variety of scales, all of these forms are related to patterns of water movement on landscape surfaces of diverse topography.

Gilgai

Patterned ground associated with both stony deserts and playa surfaces is a widespread phenomenon in warm deserts (Cooke and Warren 1973, pp. 129–49, Mabbutt 1977, pp. 130–4). Perhaps the most common and widespread type of patterned ground is gilgai. Gilgai is an Australian aboriginal word meaning small water hole, and while originally applied to small depressions that held water, it is now used to refer to a wide variety of soil patterned ground phenomena. Although the distribution and diversity of gilgai forms is perhaps greatest on the Australian continent, gilgai has also been reported from the Middle East by Harris (1958, 1959) and White and Law (1969), from South Dakota by White and Bonestall (1960), and from Death Valley, California, by Denny (1965, 1967) and Hunt and Washburn (1960). In addition to gilgai being frequently associated with soils of high swelling potential such as Vertisols, it is also a common feature in arid and semi-arid environments where strong textural contrasts exist within soils and where the climate is characterized by pronounced seasonality of precipitation. Gilgai morphology occurs in soils with annual rainfalls ranging from less than 150 mm to more than 1500 mm.

Gilgai Types

In early work on gilgai (Hallsworth and Robertson 1951, Hallsworth et al. 1955, Verger 1964, Stace et al. 1968, pp. 417 and 420, Hubble et al. 1983), six principal types of morphology were recognized (Fig. 5.6). (a) Normal gilgai is the most common form and is characterized by the development of randomly oriented mounds and shelves. The magnitude of these features ranges from imperceptible to as much as 3 m vertically with a wavelength of 15 m. If mounds are subcircular, they are referred to as puffy gilgai. (b) Melon-hole gilgai consists of large mounds separated by shelves of complex morphology. The shelves are commonly depressions with one or more sinkholes at the bottom; they are typically 1–3 m wide and 15–20 cm deep. (c) Stony gilgai, which most closely resembles the patterned ground found at high latitudes and high altitudes, has stone-covered mounds which are wide and flat in form. (d) Lattice gilgai is complex in morphology. It includes mounds that are discontinuous and oriented parallel to the direction of the slope as well as semi-continuous mounds that form networks of diverse orientation. (e) Linear or wavy gilgai develops on hillslopes ranging in gradient from 15" to 3°. Mounds and shelves are continuous and are arranged at right angles to the contour. These forms are 5–10 cm in height, and in the dry season they are very puffy in appearance. (f) Tank gilgai is large-scale gilgai that is usually rectangular in shape. Vertical dimensions range from 60 cm to 1.5 m, while depressions are 10–20 m long and 15–20 m wide. Verger's (1964) work recognizes essentially the same six basic forms, but he distinguishes between positive and negative relief groups and simulataneously between random and oriented distributions (Fig. 5.7).

The most widely developed type of gilgai in the Australian desert is stony gilgai. Ollier (1966) and Mabbutt (1977, p. 131, 1979) recognized two principal types: circular and stepped (Fig. 5.6c i and ii). Circular gilgai is characterized by the development of a relatively fine grained inner depression surrounded by a slightly raised stony rim. The depressions are commonly about 3 m in diameter, while the outer rims have diameters of approximately 8 m. This type of stony gilgai commonly forms on surfaces of low slope. The soil beneath the stony mound is generally clay-rich and has a silt crust with embedded pebbles. In contrast, the soil beneath the depression is typically sandier in the upper 30–50 cm but at depth resembles the mound soil and has abundant coarse fragments. Circular gilgai may occur either in random patterns or in networks. Stepped gilgai – or, as Mabbutt (1977, 1979) calls them, lattice systems – occur on steeper slopes with gradients of 0.5–6°. These forms are essentially distorted gilgai that become aligned across the slope. They are characterized by the development of stony risers upslope and downslope of practically stonefree treads. As with the circular gilgai forms, the risers are underlain by fine grained soils, while sandier soils underlie the treads. Treads commonly display sink holes at the base of the upslope riser (Ollier 1966, Cooke and Warren 1973, Mabbutt 1979).

Mabbutt (1979) recognized a third type of stony gilgai in the desert pavement-covered areas of the Northern Territory. He referred to this type as sorted stone polygons (Fig. 5.6c iii). These polygons are between 40 and 80 cm in diameter and are outlined by a rim of silcrete boulders sitting on a pavement of smaller pebbles. The silcrete boulders are absent from the interior depression, though the smaller pebbles are still present. Topsoil thicknesses are generally greater in the interior of the polygon (Mabbutt 1977).

Although gilgai in the arid areas of Australia is primarily associated with Desert Loams (Stace et al. 1968), it is also associated with a variety of other soils with strong textural contrasts in seasonally wet–dry environments. In particular, gilgai is strongly developed on grey cracking clays in west-central Queensland and in north-central and south-central New South Wales (Stace et al. 1968).

Detailed analyses of gilgai morphology have been undertaken by Paton (1974) and Knight (1980).

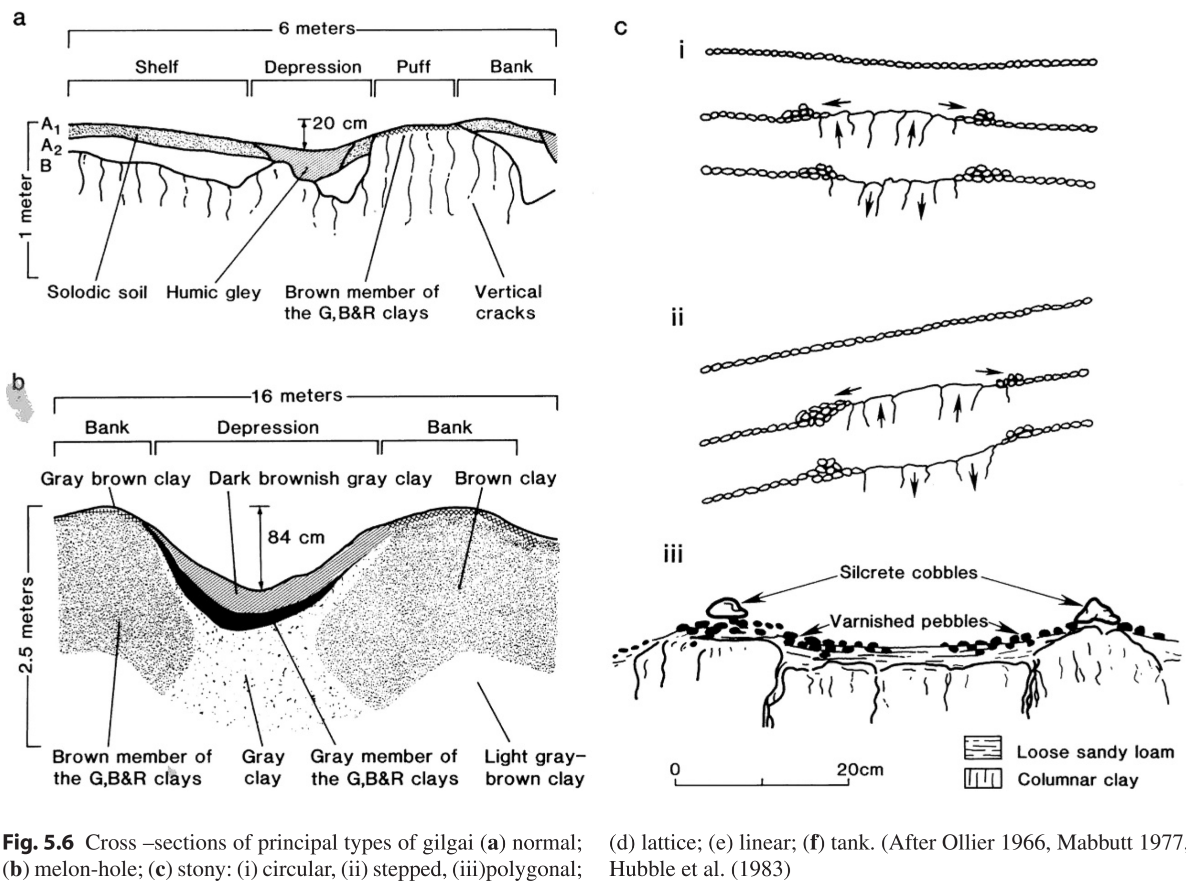

Fig. 5.6 Cross –sections of principal types of gilgai (**a**) normal; (**b**) melon-hole; (**c**) stony: (i) circular, (ii) stepped, (iii)polygonal; (d) lattice; (e) linear; (**f**) tank. (After Ollier 1966, Mabbutt 1977, Hubble et al. (1983)

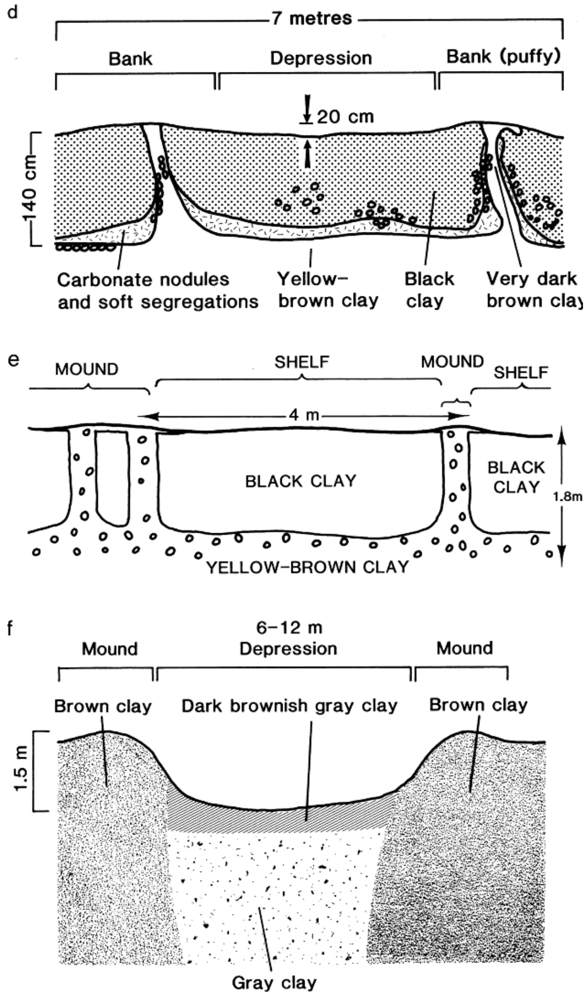

Fig. 5.6 (continued)

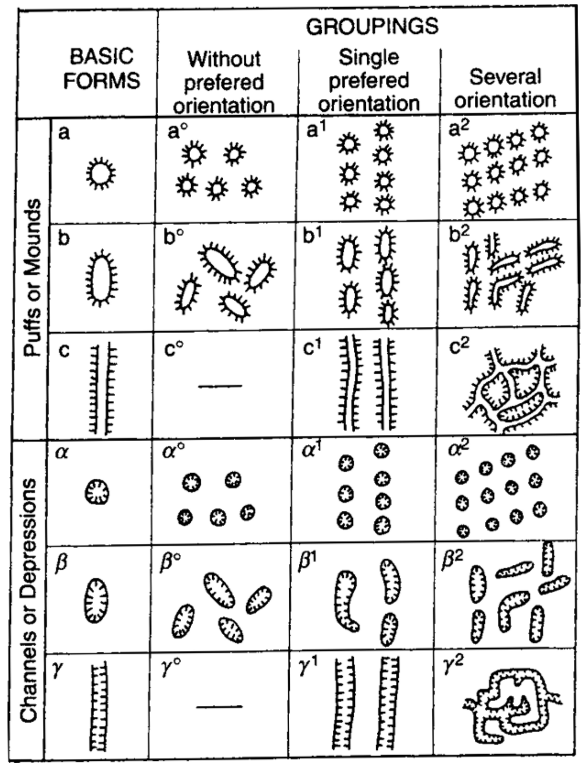

Fig. 5.7 Classification of gilgai morphology (After Verger 1964)

Knight analysed gilgai in south-eastern Australia using structural geological techniques. He recognized five recurring patterns based on the spatial arrangement of mounds, depressions, and shelves.

Gilgai Formation

A large literature exists on the origin of gilgai. Knight (1980) identified four types of gilgai-forming mechanisms: (a) heave between cracks, (b) heave over cracks, (c) contraction over cracks, and (d) heave due to loading (Fig. 5.8). Within the first type of mechanism he distinguished three subtypes (Fig. 5.8a). The first subtype involves soil compression, extrusion, and associated plastic flow, with resulting upward movement of the mound. This mechanism was originally proposed by Leeper et al. (1936). The second subtype is perhaps one of the most widely accepted mechanisms of gilgai development and involves compression and block fracture (Hallsworth et al. 1955, Verger 1964). This mechanism envisages the falling or washing of surface materials into cracks with subsequent exertion of force on subsurface clays resulting in vertical upthrusting. The third subtype calls for compression and oblique slip. This mechanism was originally proposed by White and Bonestall (1960). However, these three subtypes are all regarded by Knight (1980) as being mechanically unsound.

Two subtypes of the heave-over-cracks mechanism were recognized by Knight (1980) (Fig. 5.8b). The first he classified as being due to cumulative internal vertical movements due to small oblique slips. The second subtype is that of vertical block movement due to heave and was proposed by Howard (1939). It is also the mechanism proposed by Ollier (1966) for stony gilgai.

The contraction-over-cracks mechanism is a hypothetical one proposed by McGarity (1953) (Fig. 5.8c). It envisages the formation of depressions in the vicinity of adjacent cracks and mounds between cracks. The depressions are the result of downward movement of soil due to drying.

The heave-due-to-loading mechanism can be divided into two subtypes (Fig. 5.8d). The first involves the upward flow of soil due to density contrasts between layers (Beckman 1966, Paton 1974), whereas the second involves the flow of fluidized soil up a crack through a solid layer (Hallsworth and Beckman 1969).

Maxwell (1994) argues that most of the pre-existing models of gilgai formation are inconsistent with observed patterns of moisture distribution on undulating surfaces and that they are also inconsistent with known patterns of stress within overconsolidated expansive clays. He suggests that gilgai forms are directly related to horizonal stresses occuring naturally within these clay-rich materials. Ground surface deformation occurs when relatively small horizontal stresses modify normal vertical stresses and differential vertical rebound occurs.

Surface Cracking

The second major type of patterned ground in warm deserts is related to surface cracking. This cracking is interpreted to be largely a desiccation feature caused by drying of the surface crust (Cooke and Warren 1973, pp. 139–40). Considerable literature exists on the nature and origin of cracks in sediments. However, there has been relatively little discussion of the origin of small-scale surface cracks in deserts. An exception is the work by Tucker (1978) in northern Iraq on the origin of patterned ground associated with gypsum crusts. Tucker examined non-orthogonal crack types (Lachenbruch 1962) which range in diameter from 50 cm to 2 m, with the cracks themselves being no wider than 5 cm. The cracks are commonly infilled with gypsum and stand as ridges above the general level of the surface crust. Polygons commonly are upturned at their edges. Tucker suggested that the development of cracks and resulting polygons is the result of changes in volume of the gypsum crust. These volume changes he attributed to diurnal and seasonal changes in temperature.

Another prominent type of patterned ground occurring in deserts is associated with playa lake surfaces and is characterized by large scale dessication (Fig. 5.9). Such fissures have received considerable attention in the literature. In the United States extensive work on the nature and origin of these features has been undertaken in California by Neal (1965, 1968a, b, 1969), Neal and Motts (1967) and by Neal et al. (1968). These features are primarily the result of desiccation and subsequent cracking of crusts rich in swelling clays.

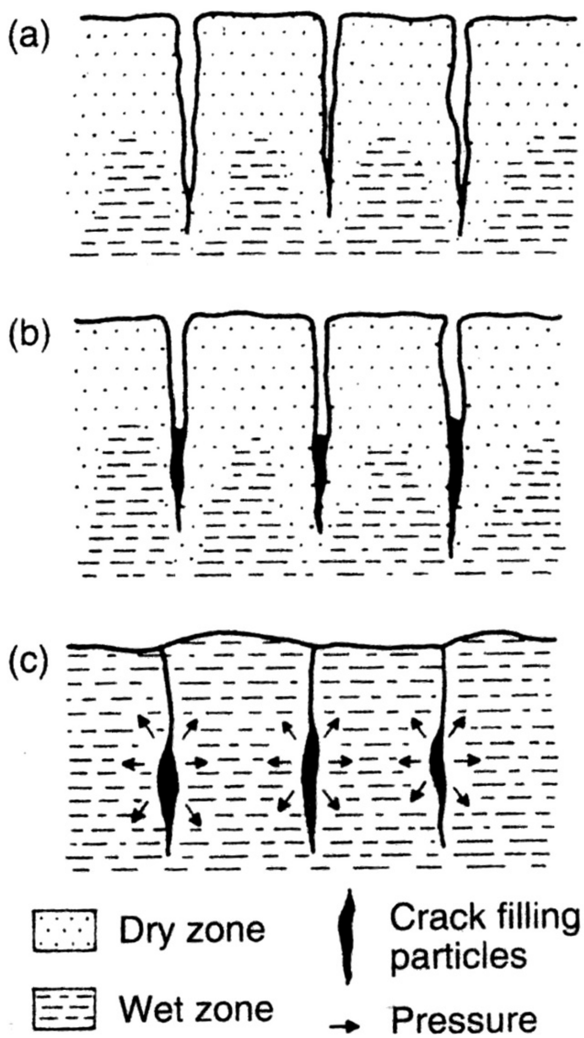

Fig. 5.8 Principal mechanisms of gilgai formation, (**a**) Heave between cracks: (i) by compression, extrusion and plastic flow, (ii) by compression and block fracture, (iii) by compression and oblique slip; (**b**) heave over cracks: (i) by cumulative internal oblique slip, (ii) by vertical block movement; (**c**) contraction over cracks due to downward movement associated with desiccation; (**d**) heave due to loading: (i) by upward flow due to density differences between layers, (ii) by upward flow of fluid soil through cracks in solid layer. (After Knight 1980)

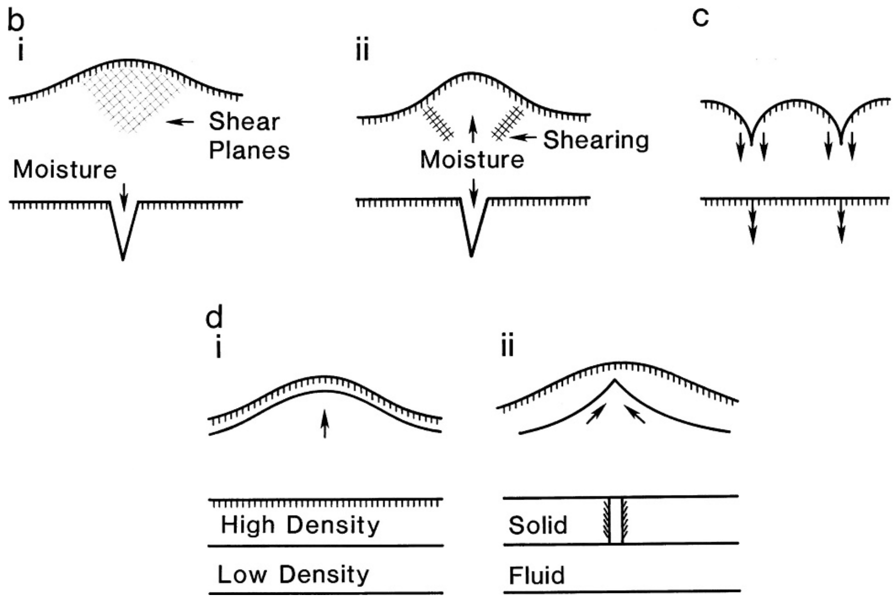

Fig. 5.8 (continued)

Fig. 5.9 Salt polygons, Death Valley Salt Pan, California

A great variety of patterned ground forms has been described from playa lakes in the western United States by Hunt and Washburn (1960, 1966). These forms are associated with different types of salt crusts. Smooth silt-rich surfaces with desiccation cracks and solution pits occur in the chloride zone beyond the limits of playa flooding. In the areas of gypsum crust, nets and polygons develop. In the carbonate zone near the edge of the salinas, sorted and non-sorted nets develop. The sorted nets form as a result of the washing of debris into the cracks. Sorted polygons develop where rock salt occurs beneath the surface. Coarse debris accumulates in depressions within silt overlying the salts. In addition, Hunt and Washburn (1966) have described a variety of patterned ground features developed on alluvial fan surfaces, including sorted steps, sorted polygons, and stone stripes.

The various patterned ground phenomena described from playa lake basins are all essentially the result of the action of salt crystallization. In addition to the desiccation of salt-rich sediment crusts on and adjacent to playas, the expansion and contraction of salts as a result of heating and cooling as well as wetting and drying play an important role in the development of patterned ground features.

Patterned ground phenomena have been described from the Great Kavi of Iran. Krinsley (1968, 1970) reported a variety of morphologies (Fig. 5.10) including salt polygons, thrust polygons, desiccation cracks, and mud pinnacles. He attributed these features largely to the influx and episodic evaporation and subsequent desiccation of salt-rich ground waters. Patterned ground has also been described from the bed of Lake Eyre in central Australia, where its origin has been ascribed to the growth of salt crystals in unconsolidated muds and accompanying ground heave along desiccation cracks (Summerfield 1991, p. 147).

Microtopographic Patterned Ground

Microtopographic patterned ground forms are also common in desert environments. These forms are often

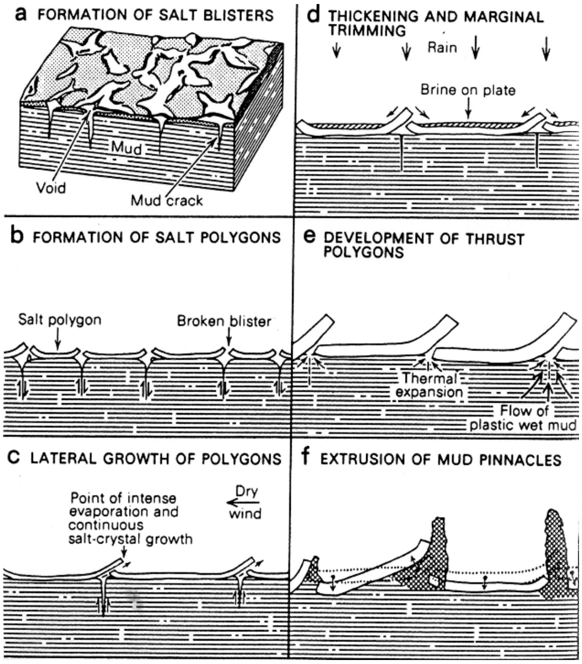

Fig. 5.10 Development stages of polygons on salt crusts. (After Krinsley 1970)

not apparent to the casual observer but are highlighted by patterns of vegetation growth, such as vegetation arcs and water lanes recognized by MacFadyen (1950) and Hemming (1965) in Somalia. Water lanes have also been described from the Somaliland Plateau by Boaler and Hodge (1962). Similar patterns have been observed in Western Australia by Mabbutt (1963). He attributed the sorting of sediment to the combined action of wind and sheet flow. Large arcuate ripples have been described in Utah by Ives (1946). Similarly, microforms are reported from playa surfaces and other gypsum-rich environments (Gutiérrez 2006) and include hemispheric domes or tumuli which commonly display central collapse depressions. These forms are attributed to numerous processes including *in situ* dissolution and precipitation of surface gypsum which fills in existing pores of the deposit. Volume increase and associated crystallization pressures then lead to lateral gypsum expansion and dome building (Artieda 1993 in Gutiérrez 2006). More recently, Calaforra (1996 in Gutiérrez 2006) suggested several alternative hypotheses including compressive tectonic forces, and volumetric increases associated with the transformation of anhydrite to gypsum. Saline ramps composed of salt crystals and shaped as half-moons have been reported by Millington et al. (1995) from playa surfaces in Tunisia. These forms are attributed to salt crystal redistribution by wind.

Desert Pavement

Morphology

Desert pavement is a stony surface generally composed of a layer of angular or subrounded gravels one or two stones thick sitting on a mantle of finer stone-free material (Mabbutt 1965, 1977, p. 119, Cooke and Warren 1973, p. 120) (Figs. 5.11 and 5.12). Desert pavement occurs widely in the warm deserts of the world (Cooke and Warren 1973, p. 121, Mabbutt 1977, p. 119). This surficial phenomenon is known by a

Fig. 5.11 Desert pavement, Cima volcanic field, California. Pavement underlain by a thick layer of aeolian-derived fine sediment (photo courtesy of L.D. McFadden)

Fig. 5.12 Desert pavement, Afton Canyon, California. Pavement underlain by a thick layer of aeolian-derived fine sediment

variety of names depending on the particular character of the pavement and the region of the world in which it occurs. Where the pavement is dominated by rock outcrops and boulders are relatively few, the landscape is referred to as hamada, an Arabic word meaning 'unfruitful'. Boulder hamada consists of extensive surfaces of large angular rock fragments and occurs extensively in deserts such as the Libyan Sahara (Mabbutt 1977, p. 121). Pavement dominated by smaller size gravel is referred to as reg in the old world and derives its name from the Arabic word meaning 'becoming smaller' (Dan et al. 1982, Amit and Gerson 1986). In the central Sahara, reg surfaces are referred to as serir (Mabbutt 1977). In Australia stone pavements are called gibber.

Genesis

Deflation

The development of desert pavement is generally attributed to one of five stone-concentrating processes. Perhaps the most commonly invoked process is that of deflation. The concentration of the coarse debris is believed to be the result of the removal of fine-grained material from the desert surface by wind, leaving the coarser debris behind as a lag deposit (e.g. Cooke 1970, Dan et al. 1981, 1982). Many workers, however, have questioned the ability of wind to transport fine-grained materials from desert surfaces. These materials are often incorporated into crusts which effectively preclude subsequent transportation (Cooke 1970). Several workers have also shown that as a gravel surface develops, the areas between the coarse fragments in fact become sheltered from the wind and fine-grained materials in these locations become progressively less likely to be transported (Pandastico and Ashaye 1956, Symmons and Hemming 1968). Finally, the fact that many desert pavements are underlain by a relatively thick layer of fine debris suggests the relative inefficiency of aeolian transport of fines from the desert surface. Some desert pavements, however, may in fact be the result of deflation, such as those observed in the Peruvian desert by Crolier et al. (1979). These workers noted that the cobbles forming the pavement are not supported by silts and clays but rather rest on sands and gravels.

Wash

A second mechanism for gravel concentration that has been widely proposed is the winnowing of fines by surface wash (Lowdermilk and Sundling 1950, Sharon 1962, Denny 1965, Cooke 1970, Dan et al. 1981, 1982 Parsons et al. 1992). Several of these workers have shown quantitatively that erosion of disturbed pavements yields considerable amounts of fine wash debris (Sharon 1962, Cooke 1970). McHargue (1981), working in the vicinity of the Aguila Mountains in Arizona, demonstrated that surface wash is a necessary component of pavement formation. He argued that incipient pavements form after the erosion of 1–3 cm of fine sediment and that stable pavements form after the removal of 3–15 cm of sediment. McHargue envisaged long-term stability of the pavement being related to subsequent addition of finegrained aeolian materials to the landscape. Similarly, Parsons et al. (1992) and Williams and Zimbelman (1994) emphasized the role of wash processes in the formation of desert pavements in southern Arizona, and Wainwright et al. (1999) demonstrated the crucial role played by wash in the reestablishment of pavement after its disturbance. However, the widespread occurrence of vesicular horizons beneath pavements has been argued to be incompatable with extensive wash and associated erosion (Williams and Zimbelman 1994)

Upward Migration of Stones

The third and perhaps most widely accepted explanation for the origin of desert pavements is the progressive upward migration of coarse particles through the underlying finer material. The mechanism most commonly invoked for this upward movement is alternate wetting and drying and associated swelling and shrinking of the fine-grained subpavement materials (Springer 1958, Jessop 1960, Cooke 1970, Dan et al. 1982). As the fine debris swells, coarse material is forced upward. When shrinkage occurs the coarse material fails to return to its former position which is occupied by fines. Repetition of this sequence of events causes the coarse material slowly to make its way to the ground surface. This process is clearly most pronounced in subpavement soils which are characterized by strong textural contrasts between the

A and B horizons. The presence of gypsic horizons beneath pavements may also impede the vertical migration of cobbles from the subsurface. Cooke and Warren (1973, p. 128) have suggested that freeze–thaw processes may also account for the formation of stone pavements in high altitude deserts such as those in South America and central Asia. The appeal of this mechanism is the fact that it offers an explanation for the fine-grained material found beneath the pavement. One of the limitations of this explanation is the fact that seldom are stones observed to be 'in transit' from below the fine-grained mantle to the ground surface (Mabbutt 1979). Further, adequate wetting depths for coarse fragment migration in deserts are difficult to invoke.

Cumulic Pedogenesis

Mabbutt (1977, 1979) proposed that perhaps some of the desert pavements in central Australia are the result of upward sorting above a fine textured mantle that is largely aeolian in origin. He suggested that aeolian dust is trapped by the rough surface of the pavement and that there is a consequent upward displacement of the pavement as dust accumulation proceeds. Sorting, if any, is limited to the uppermost part of the fine-grained aeolian mantle. More recently, a similar mechanism of pavement formation has been suggested by McFadden et al. (1987) Wells et al. (1995) and Haff and Werner (1996), who argued that the formation of desert pavement in the Cima volcanic field, California, is the result of the *in situ* mechanical weathering and accompanying downslope transport of basalt at the ground surface and the subsequent raising of the pavement as aeolian silts and clays accumulate below the gravels. In essence, the pavement is born at the earth's surface and, once formed, changes little through time. Soil development proceeds in the aeolian mantle beneath the pavement. McFadden et al. (1998) and Anderson et al. (2002) discuss the intimate relationship between vesicular horizon formation and pavements. They propose a three-stage model of pavement development involving (1) formation of a laminar crust adhered to the bottom of clasts. Crust formation is probably related to forces of adhesion and surface tension as water and sediment combine. (2) vesicular horizon development due to aeolian sediment and carbonate influx and associated clast lifting (3) formation of well developed soil structure permiting further development and thickening of the Av horizon.

Soil properties beneath desert pavements vary considerably depending on the precise nature of the pavement with respect to such characteristics as pavement coverage and clast arrangement. These characteristics have profound influences on soil properties and processes including soil texture, depth of leaching of fines and soluble salts (Wood et al. 2005). The effectiveness of cumulic pedogenesis appears to be also strongly influenced by clast size. Valentine and Harrington (2006) have demonstrated that if clasts are too small they do not form the interlocking pavement essential for silt capture and vesicular horizon formation. Support for this cumulic model of pavement formation has been provided by Wells et al. (1991) through cosmogenic dating of gravel clasts. These workers found that, using ^{3}He and ^{21}Ne, lava flow samples and clast samples have ages that are statistically inseparable from each other.

Subsurface Weathering

Mabbutt (1977) suggested a fifth process to account for the formation of desert pavement, namely differential weathering. He argued that moisture conditions are more favourable for rock weathering in the subsoil environment than at the desiccated surface, leading to the more rapid breakdown of coarse debris. The result of this enhanced weathering at depth is a layer of relatively fine materials with a lack of coarse debris. Mabbutt reasoned that this process may be particularly important in the generation of pavements in granitic terranes and that salt weathering may also be effective in the breakdown of coarse debris below the surface.

Desert pavements clearly result from a variety of processes. Some processes act independently, while others act in combination. The processes responsible for the formation of pavements vary from location to location depending on the climate, geomorphic setting, the nature of the clastic materials available, and the nature of the local soils (Bull 1991, p. 66).

Gravel Source

Central to an understanding of the origin of desert pavements is the source of the gravels. Two principal sources are generally recognized. Some pavements are composed of gravels that are fluvial in origin (Denny 1965, Cooke 1970, Mabbutt 1977, p. 122, Dan et al. 1982, Bull 1991, p. 65). These gravels are commonly rounded and occur in landscape settings in close proximity to stream channels. Some rounding of gravels can be accomplished by surface weathering processes, and so grain shape alone is not necessarily indicative of a fluvial origin. The second source is mechanical weathering of the local bedrock. This source has frequently been associated with the origin of hamadas (Cooke 1970, Mabbutt 1977, p. 121) but is now widely ascribed to finer grained pavement surfaces as well (Wells et al. 1985, McFadden et al. 1987).

Various processes have been postulated for the disintegration of bedrock and production of pavement gravels. Cooke (1970) suggested that pavement gravels in California and Chile might result from the combined influences of diurnal changes in insolation, the growth of frost needles and salt crystals, and a variety of chemical weathering processes. Peel (1974) proposed that gravels on pavements in the central Sahara are the result of insolation weathering, a view reiterated by Dan et al. (1982) for pavements in southern Israel and the Sinai. However, Dan et al. stress that cracking associated with salt crystal growth is a more important mechanism. In their detailed study of pavements in the Cima volcanic field, McFadden et al. (1987) suggested that much of the disintegration of the basalt flows is the result of mechanical disintegration processes. They proposed that salt- and clay-rich aeolian material deposited in fractures in the lava-flow surfaces experiences volumetric increase due to crystal growth and wetting and drying. These processes result in the vertical and lateral displacement of basalt clasts from the parent flows as well as subsequent breakdown of dislodged fragments. McFadden et al. suggested that some of the sand and silt in the fine-grained mantle may be derived from the mechanical breakdown of surface clasts. Some chemical alteration of the clasts has also occurred but appears to be limited to alteration of volcanic glass and iron oxides as well as olivine in the groundmass. Some authigenic clay formation is also suggested. However, much of the clay alteration seems to be related to chemical weathering of aeolian fines. A detailed discussion of weathering processes in arid environments is provided in Chapter 3.

Conclusions

Desert soils display a distinctive assemblage of physical chemical and biological characteristics. They are typically thin, gravelly, salt-dominated, and organic-poor. Their surfaces are characterized by the development of crusts and gravel lag deposits. The processes responsible for the development of these distinctive soils also result in the development of an assemblage of distinctive landform features.

The addition of considerable amounts of airborne clay to landscape surfaces in desert environments results in the development of soils with considerable swell–shrink capacity. Repeated swelling and shrinking of soils leads to the development of a variety of patterned ground forms.

Aeolian processes not only contribute significantly to soil formation and patterned ground development but they represent an important component of desert pavement formation. Silts and clays added to the desert landscape act to buoy the coarse gravel lag on the landscape surface. In addition, these fine particles as well as airborne salts serve as important agents of weathering, producing coarse debris that dominates some desert landscapes.

From the preceding discussion it is clear that a complete understanding of the geomorphic processes operating in desert environments cannot be obtained without a careful assessment of the nature of desert pedogenic processes.

References

Akpokodje, E.G. 1984. The influence of rock weathering on the genesis of gypsum and carbonates in some Australian arid zone soils. *Australian Journal of Soil Research* **22**, 243–51.

Alexander, E.B. and W.D. Nettleton 1977. Post-Mazama Natrargids in Dixie Valley, Nevada. *Soil Science Society of America Journal* **41**, 1210–12.

Amit, R. and R. Gerson 1986. The evolution of Holocene reg (gravelly) soils in deserts: an example from the Dead Sea region. *Catena* **13**, 59–79.

Amit, R. and J.B.J. Harrison 1995. Biogenic calcite horizon development under extremely arid conditions, Nizzana Sand Dunes, Israel. *Advances in Geoecology* **28**, 65–88.

Anderson, K., S.Wells and R. Graham 2002. Pedogenesis of vesicular horizons, Cima Volcanic Field, Mojave Desert, California. *Soil Science Society of America* **66**, 878–87.

Bachman, G.O. and M.N. Machette 1977. Calcic soils and calcretes in the southwestern United States. *U.S. Geological Survey Open File Report* 77–796.

Beckman, G.G. 1966. Gilgai phenomena as geologic structures: a theoretical discussion. *Proceedings of the Australian Conference of Soil Science* **3**, 1–2.

Benhur, M., I Shainberg, D Baker and R. Keren. 1985. Effect ofsoil texture and $CaCO_3$ content on water infiltration in crusted soil as related to water salinity. *Irrigation Science* **6**, 281–94.

Bettenay, E. 1962. The salt lake systems and their associated aeolian features in the semi-arid region of Western Australia. *Journal of Soil Science* **13**, 10–17.

Bettenay, E., A.V. Blackmore and F.J. Hingston 1964. Aspects of the hydrological cycle and related salinity in the Belka Valley, Western Australia. *Australian Journal of Soil Research* **2**, 187–210.

Birkeland, P.W. 1990 Soil-geomorphic research – a selected overview. *Geomorphology* **3**, 207–24.

Blackburn, W.H. 1975. Factors influencing infiltration infiltration rates and sediment production of semi-arid rangelands in Nevada. *Water Resources Research* **11**, 929–37**.**

Blank, R.R., J.A.Young and T.Lugaski 1996. Pedogenesis on talus slopes, the buckskin Range, Nevada. *Geoderma* **71**, 121–42.

Boaler, S.B. and C.A.H. Hodge 1962. Vegetation stripes in Somaliland. *Journal of Ecology* **50**, 465–76.

Boettinger, J.L. and R.J. Southard 1991. Silica and carbonate sources for aridisols on a granitic pediment, western Mojave Desert. *Soil Science Society of America Journal* **55**, 1057–67.

Boyadgiev, T.G. and W.H. Verheye 1996. Contribution to a utilitarian classification of gypsiferous soil. *Geoderma* **74**, 321–38.

Brasher, B.R., G. Borst and W.D. Nettleton 1976. Weak duripans in weathered rock in a Mediterranean climate. *Annual Meeting, American Society of Agronomy, Abstracts* 158.

Brewer, R., E. Bettenay and H.M. Churchward 1972. Some aspects of the origin and development of the red and brown hardpan soils of Bulloo Downs, W.A. *CSIRO (Australia) Division of Soils Technical Paper* 1B.

Bull, W.B. 1991. *Geomorphic responses to climate change*. New York: Oxford University Press.

Buol, S.W., F.D. Hole and R.J. McCracken 1997. *Soil genesis and classification, 4th edition.* Ames IA: Iowa State University Press.

Chadwick, O.A. and J.O. Davis 1990. Soil forming intervals caused by eolian sediment pulses in the Lahonton Basin, northwestern Nevada. *Geology* **18**, 243–6.

Chartres, C.J. 1982. The pedogenesis of desert loams in the Barrier Range, Western New South Wales, I. Soil parent materials. *Australian Journal of Soil Research* **20**, 269–81.

Chartres, C.J. 1983. The pedogenesis of desert loam soils in the Barrier Range, Western New South Wales. II. Weathering and soil formation. *Australian Journal of Soil Research* **21**, 1–13.

Chartres, C.J. 1985. A preliminary investigation of hardpan horizons in north-west New South Wales. *Australian Journal of Soil Research* **23**, 325–37.

Christianson, G.E. and C. Purcell 1985. Correlation and age of Quaternary alluvial fan sequences, Basin and Range province, southwestern United States. In *Soils and Quaternary geology of the southwstern United States*, D.L. Weide (ed.), 115–22. Geological Society of America Special Paper 203.

Conacher, A.J. 1975. Throughflow as a mechanism responsible for excessive soil salinization in non-irrigated, previously arable lands in the Western Australia wheat belt: a field study. *Catena* **2**, 31–67.

Cooke, R.U. 1970. Stone pavements in deserts. *Annals of the Association of American Geographers* **60**, 560–77.

Cooke, R.U. and A. Warren 1973. *Geomorphology in deserts*. Berkeley: University of California Press.

Crolier, M.J., G.E. Ericksen, J.F. McCauley and P.C. Morris 1979. The desert landforms of Peru; a preliminary photographic atlas. *U.S. Geological Survey Interagency Report: Astrogeology* 57.

Dan, J. 1983. Soil chronosequences in Israel. *Catena* **10**, 287–399.

Dan, J., R. Gerson, H. Kogumdjisky and D.H. Yaalon 1981. *Arid Soils of Israel*. Agricultural Research Organization, Special Publication 190.

Dan, J., D.H. Yaalon, R. Moshe and S. Nissim 1982. Evolution of reg soils in southern Israel and Sinai. *Geoderma* **28**, 173–202.

Denny, C.S. 1965. Alluvial fans in the Death Valley region, California and Nevada. *U.S. Geological Survey, Professional Paper* 466.

Denny, C.S. 1967. Fans and pediments. *American Journal of Science* **265**, 81–105.

Dimmock, G.M., E. Bettenay and M.J. Mulcahy 1974. Salt content of laterite profiles in the Darling Range, Western Australia. *Australian Journal of Soil Research* **12**, 63–9.

Dohrenwend, J.C., W.B. Bull, L.D. McFadden, G.I. Smith 1991. Quaternary geology of the Basin and Range Province in California. In *Quaternary non-glacial geology: conterminous United States*, R.B. Morrison (ed.), 321–52. Geology of North America, Vol. K-2. Boulder, CO: Geological Society of America.

Dregne, H.E. 1976. *Soils of arid regions*. Amsterdam: Elsevier.

Dunkerley, D.L. and K.J. Brown 1997. Desert Soils. In *Arid Zone Geomorphology*, D.S.G. Thomas (ed.), 55–68. Chichester: John Wiley and Sons.

Eswaran, H. and G. Zi-Tong 1991. Properties, genesis, classification and distribution of soils with gypsum. In *Occurrence, characteristics, and genesis of carbonate, gypsum and silica accumulations in soils*, W.D. Nettleton (ed.), 89–119. Soil Science Society of America Special Publication 26.

Evenari, J., D.H. Yaalon and Y. Gutterman 1974. Note on soils with vesicular structures in deserts. *Zeitschrift für Geomorphologie* **18**, 162–72.

Flach, K.W., W.D. Nettleton, L.H. Gile and J.G. Cady 1969. Pedocementation: induration by silica, carbonates, and sesquioxides in the Quaternary. *Soil Science* **107**, 442–53.

Flach, K.W., W.D. Nettleton and R.E. Nelson 1973. The micromorphology of silica cemented soil horizons in western North America. In *Soil microscopy*, G.K. Rutherford (ed.) 714–29. Kingston: Queens University.

Fletcher, J.E. and W.P Martin 1948. some effects of algae and moulds in the raincrusts of desert soils. *Ecology* **29**, 95–100.

Frenkel, H., J.O. Goertzen and J.D. Rhoads 1978. Effects of clay type and content, exchangeable sodium percentage and electrolyte concentration on clay suspension and hydraulic conductivity. *Journal of the Soil Science Society of America* **42**, 32–9.

Gerson, R. and R. Amit 1987. Rates and modes of dust accretion and deposition in an arid region: the Negev, Israel. In *Desert sediments: ancient and modern*, L. Fostick and I. Reid (eds), 157–69. Geological Society Special Publication 35.

Gerson, R., R. Amit and S. Grossman 1985. *Dust availability in desert terrains, a study in the deserts of Israel, and the Sinai*. Jerusalem: Institute of Earth Sciences, the Hebrew University of Jerusalem.

Gifford, G.F. 1972. Infiltration rate and sediment production on a plowed big sagebrush site. *Journal of Range Management* **25**, 53–5.

Gile, L.H. 1967. Soils of an ancient basin floor near Las Cruces, New Mexico. *Soil Science* **103**, 265–76.

Gile, L.H. 1970. Soil of the Rio Grande Valley border in southern New Mexico. *Soil Science Society of America Journal* **34**, 465–72.

Gile, L.H. 1975. Holocene soils and soil geomorphic relations in an arid region of southern New Mexico. *Quaternary Research* **5**, 321–60.

Gile, L.H. and R.B. Grossman 1968. Morphology of the argillic horizon in desert soils of southern New Mexico. *Soil Science* **106**, 6–15.

Gile, L.H., J.W. Hawley and R.B. Grossman 1981. Soils and geomorphology in the Basin and Range area of Southern New Mexico: guidebook to the Desert Project, *New Mexico Bureau of Mines and Mineral Resources Memoir* **39**.

Gunn, R.H. 1967. A soil catena on denuded laterite profiles in Queensland. *Australian Journal of Soil Research* **5**, 117–32.

Gunn, R.H. and D.P. Richardson 1979. The nature and possible origins of soluble salts in deeply weathered landscapes of eastern Australia. *Australian Journal of Soil Research* **17**, 197–215.

Gustavson, T.C., R.W. Baumgardner Jr, S.C. Caran, V.T. Holliday 1991. Quaternary geology of the Southern Great Plains and an adjacent segment of the Rolling Plains. In *Quaternary non-glacial geology: conterminous United States*, R.B. Morrison (ed.), 477–501. Geology of North America, Vol. K-2. Boulder, CO: Geological Society of America.

Gustavson, T.C. and D.A. Winkler 1988. Depositional facies of the Miocene–Pliocene Ogallola Formation, northwestern Texas and eastern New Mexico. *Geology* **16**, 203–6.

Guthrie, P.L. 1982. Distribution of great groups of aridisols in the United States. *Catena Supplement* **1**, 29–36.

Gutiérrez M. 2006. *Climatic geomorphology*, Amsterdam: Elsevier.

Haff, P.K. and B.T. Werner 1996. Dynamical processes on desert pavements and the healing of surficial disturbances. *Quaternary Research* **45**, 38–46.

Hallsworth, E.G. and G.G. Beckman 1969. Gilgai in the quaternary. *Soil Science* **10**, 409–20.

Hallsworth, E.G. and G.K. Robertson 1951. The nature of gilgai and melonhole soils. *Australian Journal of Science* **13**, 181.

Hallsworth, E.G., G.K. Robertson and F.R. Gibbons 1955. Studies in pedogenesis in New South Wales. VII. The gilgai soils. *Journal of Soil Science* **6**, 1–31.

Harden, D.R., N.E. Biggar and M.L. Gillam 1985. Quaternary deposits and soils in and around Spanish Valley, Utah. In *Soils and Quaternary geology of the southwestern United States*, D.L. Weide (ed.), 43–64. Geological Society of America Special Paper 203.

Harden, J.W., E.M. Taylor, L.D. McFadden and M.C. Reheis 1991. Calcic, gypsic, and siliceous soil chronosequences in arid and semiarid environments. In *Occurrence, characteristics, and genesis of carbonate, gypsum and silica accumulations in soils*, W.D. Nettleton (ed.), 1–16. Soil Science Society of America. Special Publication 26.

Harris, S.A. 1958. Gilgaied and land structured soils of central Iraq. *Journal of Soil Science* **9**, 169–85.

Harris, S.A. 1959. The classification of gilgaied soils: some evidence from northern Iraq. *Journal of Soil Science* **10**, 27–33.

Hemming, C.F. 1965. Vegetation arcs in Somaliland. *Journal of Ecology* **53**, 57–67.

Holliday, V.T. 1985a. Holocene soil geomorphological relations in a semi-arid environment: the Southern High Plains of Texas. In *Soils and Quaternary landscape evolution*, J. Boardman (ed.), 325–57. New York: Wiley.

Holliday, V.T. 1985b. Morphology of late Holocene soils at the Lubbock Lake archeological site, Texas. *Soil Science Society of America Journal* **49**, 938–46.

Holliday, V.T. 1985c. Early and middle Holocene soils at the Lubbock Lake archeological site, Texas. *Catena* **12**, 61–78.

Holliday, V.T. 1988. Genesis of late Holocene soils at the Lubbock Lake archaeological site, Texas. *Annals of the Association of American Geographers* **78**, 594–610.

Holliday, V.T. 1989. The Blackwater Draw Formation (Quaternary). A 1.4 + m.y. record of eolian sedimentation and soil formation on the Southern High Plains. *Bulletin of the Geological Society of America* **101**, 1598–607.

Holliday, V.T. 1990. Soils and landscape evolution of eolian plains: the Southern High Plains of Texas and New Mexico. *Geomorphology* **3**, 489–515.

Howard, A. 1939. Crab-hole gilgai and self-mulching soils of the Murrumbidgee Irrigation Area. *Pedology* **8**, 14–8.

Hubble, G.D. and R.F. Isbell 1958. The occurrence of strongly acid clays beneath alkaline soils in Queensland. *Australian Journal of Science* **20**, 186–7.

Hubble, G.D., R.F. Isbell and K.H. Northcote 1983. Features of Australian soils. In *Soils: an Australian viewpoint*, 17–47. Melbourne: CSIRO/London: Academic Press.

Hunt, C.B. and A.L. Washburn 1960. Salt features that simulate ground patterns formed in cold climates. *U.S. Geological Survey Professional Paper* 400-B.

Hunt, C.B. and A.L. Washburn 1966. Patterned ground. *U.S. Geological Survey Professional Paper* 494-B, 104–33.

Hutton, J.T. and J.C. Dixon 1981. The chemistry and mineralogy of some South Australian calcretes and associated soft carbonates and their dolmitization. *Journal of the Geological Society of Australia* **28**, 71–9.

Hutton, J.T. and T.I. Leslie 1958. Accession of nonnitrogenous ions dissolved in rainwater to soils in Victoria. *Australian Journal of Agricultural Research* **14**, 319–29.

Isbell, R.F. 1962. Soils and vegetation of the brigalow lands, eastern Australia. *CSIRO (Australia) Division of Soils, Soils and Landuse Series* 43.

Isbell, R.F., R. Reeve and J.T. Hutton 1983. Salt and sodicity. In *Soils: an Australian viewpoint*, 107–17, Melbourne: CSIRO/London: Academic Press.

Ives, R.L. 1946. Desert ripples. *American Journal of Science* **244**, 492–501.

Jessop, R.W. 1960. The lateritic soils of the southeastern portion of the Australian Arid Zone. *Journal of Soil Science* **11**, 106–13.

Knight, M.J. 1980. Structural analysis and mechanical origins of gilgai at Boorook, Victoria, Australia. *Geoderma* **23**, 245–83.

Krinsley, D.B. 1968. Geomorphology of three kavirs in northern Iran. *Air Force Cambridge Research Laboratories Environmental Research Papers* 283.

Krinsley, D.B. 1970. A geomorphological and paleoclimatological study of the playas of Iran. *U.S. Geological Survey. Final Scientific Report Contract PROCP-70-800.*

Lachenbruch, A.H. 1962. Mechanics of thermal contraction cracks and ice-wedge polygons in permafrost. *Geological Society of America Special Paper* 70.

Leeper, G.W., A. Nicholls and S.M. Wadham 1936. Soil and pasture studies in the Mount Gellibrand area, Western District of Victoria. *Proceedings of the Royal Society of Victoria* **49**, 77–138.

Litchfield, W.H. and J.A. Mabbutt 1962. Hardpan soils of semiarid Western Australia. *Journal of Soil Science* **13**, 148–60.

Lowdermilk, W.C. and H.L. Sundling 1950. Erosion pavement, its formation and significance. *Transactions of the American Geophysical Union* **31**, 96–100.

Mabbutt, J.A. 1963. Wanderrie banks: micro-relief patterns in semiarid Western Australia. *Bulletin of the Geological Society of America* **74**, 529–40.

Mabbutt, J.A. 1965. Stone distribution in a stony tableland soil. *Australian Journal of Soil Research* **3**, 131–42.

Mabbutt, J.A. 1977. *Desert landforms*, Cambridge, Mass. MIT Press.

Mabbutt, J.A. 1979. Pavements and patterned ground in the Australian stony deserts. *Stuttgarter Geographische Studen* **93**, 107–23.

Machette, M.N. 1985. Calcic soils of the Southwestern United States. In *Soils and Quaternary geology of the southwestern United States*, D.L. Wide (ed.) 1–21. Geological Society of America Special Paper 203.

Maxwell, B. 1994. Influence of horizontal stresses on gilgai landforms. *Journal of Geotechnical Engineering* **120**, 1437–44.

McFadden, L.D. 1988. Climatic influences on rates and processes of soil development in Quaternary deposits of southern California. In *Paleosols and weathering through geologic time: principles and applications*, J. Reinhardt and W.R. Sigleo (eds), 153–77, Geological Society of America Special Paper 206.

McFadden, L.D. and J.C. Tinsley 1985. Rate and depth of pedogenic-carbonate accumulation in soils: formulation and testing of a comparintment model. In *Soils and Quaternary geology of the southwestern United States*, D.L. Weide (ed.), 23–41. Geological Society of America Special Paper 203.

McFadden, L.D., S.G. Wells, J.C. Dohrenwend and B.D. Turrin 1984. Cumulic soils formed in eolian parent materials on flows of the late Cenozoic Cima volcanic field, California. In *Surficial geology of the eastern Mojave Desert, California*, J.C. Dohrenwend (ed.), 134–49. Field Trip Guide, 97th Annual Meeting of the Geological Society of America.

McFadden, L.D., S.G. Wells and J.C. Dohrenwend 1986. Influences of Quaternary climate changes on processes of soil development on desert loess deposits of the Cima volcanic field; California. *Catena* **13**, 361–89.

McFadden, L.D., S.G. Wells and M.J. Jercinovich 1987. Influences of eolian and pedogenic processes on the origin and evolution of desert pavements. *Geology* **15**, 504–8.

McFadden, L.D., R.G. Amundson and O.A. Chadwick 1991. Numerical modeling, chemical, and isotopic studies of carbonate accumulation in soils of arid regions. In *Occurrence, characteristics, and genesis of carbonate, gypsum and silica accumulation in soils*, W.D. Nettleton (ed.), 17–35. Soil Science Society of America. Special Publication 26.

McFadden, L.D., E.V. McDonald, S.G. Wells, K.Anderson, J. Quade and S.L. Forman 1998. The vesicular layer and carbonate collars of desert soils and pavements: formation, age and relation to climate change. *Geomorphology* **24**, 101–45.

MacFadyen, W.A. 1950. Vegetation patterns in the semidesert plains of British Somalia. *Geographical Journal* **116**, 199–211.

McGarity, J.W. 1953. Melon hole formation in the Richmond River District of New South Wales. *Proceedings of the Australian Conference on Soil Science* **2**, 1–7.

McHargue, L.E. 1981. Late Quaternary deposition and pedogenesis on the Aquila Mountains piedmont, southeastern Arizona. M.S. thesis, University of Arizona, Tucson.

McIntyre, D.S. 1958a. Permeability measurements of soil crusts formed by raindrop impact. *Soil Science* **85**, 185–89.

McIntyre, D.S. 1958b. Soil splash and the formation of surface crusts by raindrop impact. *Soil Science* **85**, 261–66.

Millington, A.C., N.A. Drake, K. White and R.G. Bryant. 1995. Salt ramps: wind-induced depositional features on tunisian playas. *Earth Surface Processes and Landforms* **20**, 103–13.

Miller, D.E. 1971. Formation of vesicular structure in soil. *Soil Science Society of America Journal* **35**, 635–7.

Milnes, A.R. and J.T. Hutton 1983. Calcretes in Australia. In *Soils: an Australian viewpoint*, 119–62, Melbourne: CSIRO/London: Academic Press.

Milnes, A.R., M.J. Wright and M. Thiry 1991. Silica accumulations in saprolites and soils in South Australia. In *Occurrence, characteristics, and genesis of carbonate, gypsum and silica accumulations in soils*, W.D. Nettleton (ed.) 121–49. Soil Science Society of America Special Publication 26.

Monger, H.C., L.A. Daugherty and L.H. Gile 1991. A microscopic examination of pedogenic calcite in an aridisol of southern New Mexico. In *Occurrence, characteristics, and genesis of carbonate, gypsum and silica accumulations in soils*, W.D. Nettleton (ed.), 37–60. Soil Science Society of America. Special Publication 26.

Neal, J.T. 1965. Giant desiccation polygons of Great Basin playas. *Air Force Cambridge Research Laboratories Environmental Research Papers* 123, 1–30.

Neal, J.T. 1968a. Playa surface changes at Harper Lake California: 1962–1967. *Air Force Cambridge Research Laboratories Environmental Research Papers* 283.

Neal, J.T. (ed.) 1968b. Playa surface morphology: miscellaneous investigations. *Air Force Cambridge Research Laboratories Environmental Research Papers* 283.

Neal, J.T. 1969. Playa variations. In *Arid lands in perspective*, W.G. McGimies and B.J. Goldman (eds), 13–44. Tucson: University of Arizona Press.

Neal, J.T. and W.S. Motts 1967. Recent geomorphic changes in playas of the western United States. *Journal of Geology* **75**, 511–25.

Neal, J.T., A.M. Langer and P.F. Kerr 1968. Giant desiccation polygons of Great Basin playas. *Bulletin of the Geological Society of America* **79**, 69–90.

Nelson, R.E., L.C. Klameth and W.D. Nettleton 1978. Determining soil gypsum content and expressing properties of gypsiferous soils. *Soil Science Society of America Journal* **42**, 659–61.

Nettleton, W.D. and F.F. Peterson 1983. Aridisols. In *Pedogenesis and soil taxonomy II. The soil orders. Developments in Soil Science IIB*, L.P. Wilding, N.E. Smeck and G.F. Hall (eds), 165–215. Amsterdam: Elsevier.

Nettleton, W.D., J.G. Witty, R.E. Nelson and J.W. Hanley 1975. Genesis of argillic horizons in soils of desert areas of the southwestern. *U.S. Soil Science Society of America Journal* **39**, 919–26.

Nettleton, W.D., R.E. Nelson, B.R. Brasher and P.S. Deer 1982. Gypsiferous soils in the western United States. In *Acid sulfate weathering*, J.A. Kittrick, D.S. Fanning and L.R. Hossner (eds), 147–68. Soil Science Society of America. Special Publication 10.

Nettleton, W.D., E.E. Gamble, B.L. Allen, G. Borst 1989. Relief soils of subtropical regions of the United States. *Catena Supplement* **16**, 59–93.

Nettleton, W.D., B.R. Brasher and S.L. Baird 1991. Carbonate clay characterization by statistical methods. In *Occurrence, characteristics, and genesis of carbonate, gypsum and silica accumulations in soils* W.D. Nettleton (ed.), 75–88. Soil Science Society of America. Special Publication 25.

Northcote, K.H. and J.K.M. Skene 1972. Australian soils with saline and sodic properties. *CSIRO (Australia) Soil Publication* 27.

Ollier, C.D. 1966. Desert gilgai. *Nature* **212**, 581–3.

Painuli, D.K. and I.P. Abrol 1986. Effect of exchangeable sodium percent on surface scaling. *Agricultural Water Management* **11**, 247–256.

Pandastico, E.B. and T.I. Ashaye 1956. Demonstration of the effect of stone layers on soil transport and accretion. *In Experimental pedology*, E.G. Hallsworth and D.V. Crawford (eds), 384–90. London: Butterworth.

Parsons, A.J., A.D. Abrahams and J.R. Simanton 1992. Microtopography and soil-surface materials on semi-arid piedmont hillslopes, southern Arizona. *Journal of Arid Environments* **22**, 107–15.

Paton, T.A. 1974. Origin and terminology for gilgai in Australia. *Geoderma* **11**, 221–42.

Peck, A.J. 1978. Salinization of non-irrigated soils and associated streams: a review. *Australian Journal of Soil Research* **16**, 157–68.

Peck, A.J. and D.H. Hurle 1973. Chloride balance of some farmed and forested catchments in southwestern Australia. *Water Resource Research* **9**, 648–57.

Peel, R.F. 1974. Insolation weathering: some measurements of diurnal temperature change in exposed rocks in the Tibest region, central Sahara. *Zeitschrift für Geomorphologie Supplement Band* **21**, 19–28.

Peterson, F.F. 1980. Holocene desert soil formation under sodium salt influence in a playa-margin environment. *Quaternary Research* **13**, 172–86.

Ponti, D.J. 1985. The Quaternary alluvial sequence of the Antelope Valley, California. In *Soils and Quaternary geology of the southwestern United States*, D.L. Weide (ed.), 79–96. Geological Society of America Special Paper 203.

Prescott, J.A. 1931. The soils of Australia in relation to vegetation and climate. *CSIRO (Australia) Bulletin* **52**.

Rabenhorst, M.C., L.T. West and L.P. Wilding 1991. Genesis of calcic and petrocalcic horizons in soils over carbonate rocks. In *Occurrence, characteristics, and genesis of carbonate, gypsum and silica accumulations in soils*, W.D. Nettleton (ed.), 61–74, Soil Science Society of America. Special Publication 26.

Reheis, M.C. 1987a. Gypsic soils on the Kane alluvial fans, Big Horn Country, Wyoming. *U.S. Geological Survey Bulletin* **159-C**.

Reheis, M.C. 1987b. Climate implications of alternating clay and carbonate formation in semi arid soils of south-central Montana. *Quaternary Research* **29**, 270–82.

Reheis, M.C. 1990. Influence of climate and eolian dust on major-element chemistry and clay mineralogy of soils in the northern Bighorn Basin, U.S.A. *Catena* **17**, 219–48.

Reheis, M.C., J.W. Harden, L.D. McFadden and R.R. Shroba 1989. Development rates of late Quaternary soils: Silver Lake Playa, California. *Soil Science Society of America Journal* **53**, 1127–40.

Reheis, M.C., J.C. Goodmacher, J.W. Harden, L.D. McFadden, T.K. Rockwell, R.R. Shroba, J.M. Sowers and E.M. Taylor 1995. Quaternary soils and dust deposition in southern Nevada and California. *Geological Society of America Bulletin* **107**, 1003–22.

Schaetzl, R. and S. Anderson. 2005. *Soils: Genesis and Geomorphology*. Cambridge University Press: Cambridge.

Sharon, D. 1962. On the nature of hamadas in Israel. *Zeitschrift für Geomorphologie* **6**, 129–47.

Shlemon, R.J. 1978. Quaternary soil-geomorphic relationships, southeastern Mojave Desert, California and Arizona. In *Quaternary soils*, W.C. Mahany (ed.), 187–207, Norwich: Geo Books.

Smith, B.R. and S.W. Buol 1968. Genesis and relative weathering intensity studies in three semiarid soils. *Soil Science Society of America Journal* **32**, 261–5.

Soil Survey Staff 1975. Soil taxonomy: a basic system of soil classification for making and interpreting soil survey. *U.S. Department of Agriculture Agriculture Handbook* 436.

Soil Survey Staff 1994. *Keys to Soil Taxonomy 6th edition*. Pocahontas Press: Blacksburg VA.

Southard, R. 2000. Aridisols. In *Handbook of Soil Science*, M.E. Sumner (ed.), E-321- E 338, CRC Press: Boca Raton.

Springer, M.E. 1958. Desert pavement and vesicular layer in some soils of the Lahontan Basin, Nevada. *Soil Science Society of America Journal* **22**, 63–6.

Stace, H.C.T., G.D. Hubble, R. Brewer, K.H. Northcote 1968. *A handbook of Australian soils*, Glenside, South Australia: Rellim Technical Publications.

Stephens, C.G. 1962. *Manual of Australian soils*. 3rd edn. Melbourne: CSIRO.

Sullivan, L.A. and A.J. Koppi 1991. Morphology and genesis of silt and clay coatings in the vesicular layer of a desert loam soil. *Australian Journal of Soil Research* **29**, 579–86.

Summerfield, M.A. 1991. *Global geomorphology*, New York: Wiley.

Symmons, P.M. and C.F. Hemming 1968. A note on wind-stable stone-mantles in the southern Sahara. *Geographical Journal* **134**, 60–4.

Tucker, M.E. 1978. Gypsum crusts (gypcrete) and patterned ground from northern Iraq. *Zeitschrift für Geomorphologie* **22**, 89–100.

Valentine, G.A. and Harrington, C.D. 2006. Clast size controls and longevity of pleistocene desert pavements at Lathrop Wells and Red Cone volcanoes, southern Nevada. *Geology* **34**, 533–36.

Verger, F. 1964. Mottereaux et gilgais. *Annales de Géographie* **73**, 413–30.

Wainwright, J., A.J. Parsons and A.D. Abrahams 1999. Field and computer simulation experiments on the formation of desert pavement. *Earth Surface Processes and Landforms* **24**, 1025–1037.

Weide, D.L. (ed.) 1985. Soils and Quaternary geology of the southwestern United States. *Geological Society of America Special Paper* 203.

Wallace, A., E.M. Romney and V.O. Hale 1973. Sodium relations in desert plants: 1. Cation contents of some desert plant species from the Mojave and Great Basin deserts. *Soil Science* **115**, 284–87.

Wells, S.G., J.C. Dohrenwend, L.D. McFadden, B.D. Turrin 1985. Late Cenozoic landscape evolution on lava flow surfaces of the Cima volcanic field, Mojave Desert, California *Bulletin of the Geological Society of America* **96**, 1518–29.

Wells, S.G., L.D. McFadden and J.C. Dohrenwend 1987. Influence of late Quaternary climatic changes on geomorphic and pedogenic processes on a desert piedmont eastern Mojave Desert, California. *Quaternary Research* **27**, 130–46.

Wells, S.G., L.D. McFadden and J.D. Schultz 1990. Eolian landscape evolution and soil formation in the Chaco dune field, southern Colorado Plateau, New Mexico. *Geomorphology* **3**, 517–46.

Wells, S.G., L.D. McFadden and C.T. Olinger 1991. Use of cosmogenic 3Ne and 21Ne to understand desert pavement formation. *Geological Society of America Abstracts with Programs* **23** (5), 206.

Wells, S.G., L.D. McFadden, J. Poths and C.T. Olinger 1995. Cosmogenic ^{3}He surface exposure dating of stone pavements: implications for landscape evolution in deserts. *Geology* **23**, 613–16.

Wetselaar, R. and J.T. Hutton 1963. The ionic composition of rainwater of Katherine, N.T. and its part in the cycling of plant nutrients. *Australian Journal of Agricultural Research* **14**, 319–29.

White, E.M. and R.G. Bonestall 1960. Some gilgaied soils of South Dakota. *Soil Science Society of America Journal* **24**, 305–9.

White, L.P. and R. Law 1969. Channeling of alluvial depression soils in Iraq and Sudan. *Journal of Soil Science* **20**, 84–90.

Williams, S.H. and J.R. Zimbelman 1994. Desert pavement evolution: an example of the role of sheetflood. *Journal of Geology* **102**, 243–48.

Wood, Y.A., R.C. Graham and S.G. Wells 2005. Surface control of desert pavement pedologic process and landscape function, Cima Volcanic field, Mojave Desert, California. *Catena* **59** 205–30.

Wright, M.J. 1983. Red-brown hardpans and associated soils in Australia. *Transactions of the Royal Society of South Australia* **107**, 252–4.

Yaalon, D.H. 1964. Airborne salt as an active agent in pedogenic processes. *Transactions of the 8th Congress of the International Soil Science Society* **5**, 994–1000.

Yaalon, D.H. and E. Ganor 1973. The influence of dust on soils during the Quaternary. *Soil Science* **116**, 146–55.