

Chapter 8

Rock Slopes

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Introduction

The very slow chemical and physical weathering rates in desert areas coupled with a relatively high efficiency of wash processes, due to the general sparseness of vegetation, results in more widespread occurrence of slopes with little or no regolith than in areas with humid climates. This chapter outlines the processes and landforms occurring on desert slopes that are either massive bedrock or are scarps and cuerdas in layered rocks dominated by outcropping resistant rock layers.

Slopes in Massive Rocks

Hillslopes formed on bodies of massive rock – that is, with few joints and high intact strength – are found in most of the world's deserts. These deserts vary in age, past climatic regimes, structure, and geological history. It seems inherently improbable, therefore, that all hillslopes will have one set of controls on their form, or even that the same controls have prevailed throughout the evolution of the hillslope. In this section the following influences on hillslope form will be considered: rock mass strength and slope forms; structural influences; sheeting; bornhardts; karst in siliceous rocks; etch processes and inherited forms.

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Rock Mass Strength and Slope Forms

There has been a general recognition by geomorphologists that rock resistance to processes of erosion has played a part in controlling the form of hillslopes, particularly in arid regions, but attempts to study such relationships in detail were dependent upon the development of rock mechanics as a discipline. The work of tunnelling and mining engineers has been important in the establishment of methods of investigation (see, for example, Terzaghi 1962, Brown 1981, Hoek and Bray 1981, Brady and Brown 1985, Bieniawski 1989). The application of rock mechanics approaches to the understanding of hillslopes has been particularly fruitful in arid and semi-arid areas where excellent exposure of rock outcrops is available (Selby 1980, 1982a,b,c,d, Moon and Selby 1983, Moon 1984a, 1986, Allison and Goudie 1990a,b).

The underpinning of the geomorphological studies listed above was the formulation of a semiquantitative method of evaluating the strength of rock masses (Selby 1980). The method involves a five-level ordinal rating of seven characteristics of exposed rock related to mass strength. These are intact rock strength (typically measured by Schmidt hammer), degree of weathering, spacing of joints, orientation of joints relative to the slope surface, joint width, joint continuity, and degree of groundwater outflow. The individual ratings of these characteristics are incorporated into a weighted sum characterizing rock mass strength. Details of this method have been referred to in accessible sources, such as Selby (1982d) and Gardiner and Dackombe (1983). The original method is fundamentally unchanged but Moon (1984b) has suggested some refinements to measurements and

Abrahams and Parsons (1987) have improved the definition of the envelope defining strength equilibrium slopes.

The contention that strength equilibrium slopes are widespread is confirmed by independent work and has a number of implications. (a) Rock slopes in strength equilibrium retreat to angles which preserve that equilibrium: if the process of retreat brings to the slope surface rocks of lower mass strength the slope angle will become lower; if higher strength rock masses are exposed, slope angles will steepen; only if rock mass strength is constant will slopes retreat parallel to themselves. (b) Only if controls other than those of mass strength supervene will the above generalization be invalid. (c) Rock slopes evolve to equilibrium angles relatively quickly. If this were not so, slope angles too high or too low for equilibrium would be more common than they are. (d) Gentle rock slopes remain essentially uneroded until either weakened by weathering or steepened by undercutting to the strength equilibrium gradient. Application of the technique is useful for distinguishing the effects of controls other than rock mass strength on slope inclination.

The assessment of strength equilibrium for hillslopes forming escarpments and inselbergs in South Africa, the Namib Desert, and Australia and on a great variety of lithologies has demonstrated that there is no simplistic model for slope evolution of the kind implied in such terms as parallel retreat or downwearing. If such ideas have any validity, it can only be in application to a particular variation in mass strength into a rock mass as the slopes on it evolve.

In all of the early work on the form of rock slopes the units for assessment were chosen as having a uniform slope and angle or an obvious uniformity of rock properties. Such a selection process has certain advantages, but it limits the application of statistical assessment of parameters. Allison and Goudie (1990a) have introduced the use of a 'kennedy parameter' for measuring slope shape and a fixed distance of 5 m over which slope angle changes are recorded. This method permits assessment of slope curvature. The kennedy parameter has a value of 0.5 for straight slopes, increasingly concave slopes have values tending towards 0.0 and increasingly convex slopes towards 1.0. Symmetrical concavo-convex or convexo-concave slopes also have a value of 0.5. Profile records must therefore include profile plots if errors are to be avoided. The same

workers also advocate the use of the sine of the slope angle if frequency distributions and statistical analyses are to be applied to slope profile data.

Structural Influences

The rock mass strength rating system is insensitive of the condition in which critical joints or weak bedding planes dip steeply out of a slope – that is, where the stability of the overall slope is controlled by deep-seated structural influences rather than processes and resistance operating at the scale of individual joint blocks. If joints are infilled with weak materials, such as clays, the critical angle for stability could be as low as 10° (see Selby 1982d, pp. 72 and 158). A careful assessment of the shear strength along critical joints is then required. Basic methods are described by Brown (1981) and their application to geomorphic studies by Selby (1987).

Moon (1984a) has studied steep slopes in the Cape Mountains and recognized both unbuttressed and buttressed slopes with units which are as steep as 90°. Buttressed slopes are supported by rock units lower on the hillslope whenever the dominant joints fail to outcrop. This happens most obviously where bedding joints plunge at the same angle as the hillslope angle. Buttressed slopes have slope angles less than their apparent strength based upon the rock mass strength parameters due to lack of exposed dominant jointing. Such slopes demonstrate the importance of a lack of weathering along the joints and the lack of dilation which would otherwise allow sliding to take place. The effective friction angle along critical joints is assumed to exceed 55° in some cases, as there are few signs of mass failure of scarp faces. The elimination of buttressing by weathering, buckling, or other processes is an essential prerequisite to the development of strength equilibrium.

Rock slopes out of equilibrium with the rock mass strength rating can also result from rapid undercutting (by stream erosion of mass wasting lower on the slope), form dominance by non-mass-wasting processes (such as solution of limestones), rocks with a regolith cover, exhumation (for example, structural planes on resistant caprocks exposed by erosion of overlying weak rocks, former Richter slopes denuded of their debris cover, and some bornhardts), and slopes dominated by sheeting fracturing.

Sheeting

The formation of planar or gently curved joints conformable with the face of a cliff or valley floor is called sheeting. To merit this name there usually is evidence that the joints form more than one layer of separating rock. Such evidence is obtained from exposures which reveal parallel shells of rock separating from the parent rock mass, which is massive. Sheeting joints have been described from several rock types which can form massive bodies; granite and hard, dense sandstones (Bradley 1963) are the most common. In deserts domed inselbergs, called bornhardts, and high cliffs are the most common features which develop sheet structures (Figs. 8.1 and 8.2).

At least four major hypotheses have been proposed to account for sheeting: (a) sheeting results from stress release; (b) sheet structures are developed in granitic rocks by the formation of stretching planes during intrusion into the crust; (c) sheeting is the result of faulting with the development of secondary shears; and (d) lateral compression within the crust creates dome-like forms with concentric jointing being developed during the compression. The expansion of rock during weathering produces thin slabs of rock which, if confined laterally, may arch and create small-scale features which are similar to the larger-scale forms created by crustal stresses.

The unloading or stress release hypothesis was expressed in its most persuasive form by Gilbert (1904). He, like many geologists, was impressed by the



Fig. 8.1 A sheet of arkosic sandstone has separated from the dome of Ayers Rock, central Australia. The bedding of the rock lies in a vertical plane and has no influence on the sheeting. Note the tafoni on the dome face, which have been interpreted as having an origin within a weathering mantle (photo M.J. Selby)



Fig. 8.2 Sheeting on the surface of a granite dome, showing various degrees of separation. Gross Spitzkoppe, Namibia (photo M.J. Selby)

evidence of rock bursts in mines and deep tunnels, of the springing up of rock slabs after retreat of glaciers and by the common occurrence of sheeting on the walls of deglaciated valleys and in quarries. Work in quarries (Dale 1923, Jahns 1943) showed clearly that the thickness of sheets increases with depth into the rock mass. Sheet thicknesses range from less than one metre to more than ten metres and transect rock structures, or even dykes in rock bodies, which have few or no other joint sets. Sheet structures terminate laterally where they meet other joints or weak rock units.

Stresses in the upper crust are usually described as being derived from four major sources: gravitational, tectonic, residual, and thermal (see Selby et al. 1988 for a review). Gravitational stresses at a point within a rock body are induced by the weight of the column of rock above that point. Rocks under load tend to expand transversely to the direction of the applied load, with the resulting transverse stress having a magnitude which is approximately one-third that of the vertical stress. Even if the overburden is removed the tendency for transverse expansion to occur is locked into the rock mass as a residual stress (Haxby and Turcotte 1976). Tectonic stresses result from convergence, and their presence is indicated by thrust faulting and conjugate joint sets. Stress fields in areas of convergence may yield horizontal stresses which exceed the local vertical stresses (McGarr and Gay, 1978). Thermal stresses result from the prevention of expansion or contraction of a solid during heating or cooling. The magnitude of the stresses in a confined rock body which cools through 100°C may exceed the tensile strength of the rock (Voight and St Pierre 1974). The stresses from

gravitational, tectonic, and thermal effects may all be locked in the rock body as residual stresses if they cannot be relieved by expansion of the rock, internal deformation, or the development of joints.

In weak rocks and soils, continuous joint systems do not develop freely because the high void space and the presence of many microcracks and plastic clays within their structure prevent fracture propagation. In dense, strong rocks, by contrast, joints can propagate with few impediments (for a discussion of fracture mechanics and crack propagation see Einstein and Dershowitz 1990).

The tendency of rock bodies to expand as a result of gravitational loading and thermal stress from cooling has been analysed in some detail by finite-element analyses. Such analyses have now been applied to a number of geomorphic phenomena at scales ranging from small cliffs and individual slopes to mountain peaks and mountain massifs (see, for example, Yu and Coates 1970, Sturgul et al. 1976, Lee 1978, Augustinus and Selby 1990). It is evident from such analyses that all of the four forms of stress described above can create a tendency for rock masses to expand laterally and for stresses to be concentrated at particular sites along cliff faces, usually at the bases of cliffs (Yu and Coates 1970). Furthermore, the magnitude of the stresses commonly exceeds the tensile strength of the rock. As fractures are propagated in directions which are normal to the direction of the principal stress (Einstein and Dershowitz 1990), it is evident that stress relief can be a major cause of sheeting which can operate when the confining pressure of surrounding rock masses is removed by erosion.

Yatsu (1988) has offered a dissenting view of the role of residual or 'locked-in' stresses on development of sheeting and rock bursts, maintaining that steady-state creep can erase such stresses over the long timescales required for exposure of formerly deeply buried rock. Yatsu emphasized the role of neotectonic stress fields and gravitational stress related to present-day topography.

The evidence for steady-state creep is likely to be found in thin sections taken from rock at shallow depth. Such sections may be expected to show alignment of minerals such as micas and development of shears in crystals and silica overgrowths in orientations which are unrelated to original rock microstructure and are aligned downslope. No such evidence has been reported.

Neotectonic stress fields would be expected to create preferred alignments of sheeting joints, but sheeting occurs parallel to valley walls and dome surfaces with many orientations in a small area. Furthermore, many areas of sheeting occur in cratons far from areas of Cenozoic tectonism.

Gravitational stress is caused by existing overburdens and is relevant to flanks of steep-sided peaks and ridges and to walls of deep valleys. It is not relevant to areas of low relative relief where many examples of stress relief are found. It should be noted also that, although a few studies of stresses in individual boulders have been carried out, residual stresses have been measured in joint-bounded columns of basaltic lava (Bock 1979). Much further work is needed, but Bock's work has the implication that residual stresses, of thermal or gravitational origin, may be locked into small rock units.

The alternative mechanisms listed in the second paragraph of this section are not necessarily excluded from being contributing factors to sheeting but are, in essence, special cases which can apply to only a few cases in specific environments.

Mass Wasting Influences

Many steep bedrock slopes in desert environments exhibit chutes – linear depressions oriented downslope. Although most of these are developed along zones of structural weakness, their deepening and downslope integration may be enhanced by the motion of rockfalls and avalanches through the chutes. Such features are most elegantly developed as 'spur and gully' terrain in arctic and alpine environments where competition from runoff processes is negligible (Matthes 1938, Blackwelder 1942, Rapp 1960a,b, Luckman 1977, 1978, Akerman 1984, Rudberg 1986) and on the sides of steep scarps on Mars (Sharp and Malin 1975, Blasius et al. 1977, Lucchitta 1978). In arctic and alpine terrain dry rockfalls, debris avalanches, and snow avalanches appear to be capable of rock erosion (Matthes 1938, Blackwelder 1942, Rapp 1960a,b, Peev 1966, Gardiner 1970, 1983, Hewett 1972, Luckman 1977, 1978, Corner 1980, O'Loughlin and Pearce 1982, Ackroyd 1987). Lucchitta (1978) and Patton (1981) cited examples from desert areas, but competition from fluvial erosion makes the evidence for rockfall and avalanche erosion equivocal.

Howard (1990) has modelled gully erosion by avalanching using avalanche rheology and an approach similar to that used from modelling of snow avalanches (Perla et al. 1980, Dent and Lang 1980, 1983, Martinelli et al. 1980) and some debris avalanches (Cannon and Savage 1988, McEwen and Malin 1990). These models generally assume that the avalanche moves as a unit with the net downslope surface shear τ being expressed by a relationship

$$\tau = \rho_b g h (\cos \theta \tan \Phi - \sin \theta) - \rho_b C_f V^n \quad (8.1)$$

where h is the avalanche thickness (often assumed to be constant during the avalanche motion), g is the gravitational acceleration, θ is the slope angle, Φ is a friction angle, V is the avalanche velocity, ρ_b is the flow bulk density, n is an exponent, and C_f is a friction coefficient. A theoretical basis for C_f is not firmly established and may represent air drag, internal frictional dissipation, and ‘ploughing’ of surface material (Perla et al. 1980). Some models utilize a ‘turbulent’ friction with $n = 2$, and others assume ‘laminar’ friction with $n = 1$ (Buser and Frutiger 1980, Dent and Lang 1980, 1983, Lang and Dent 1982, Perla et al. 1980, Martinelli et al. 1980, McClung and Schaerer 1983, Schiewiller and Hunter 1983, McEwen and Malin 1990, Cannon and Savage 1988). Rather than finding analytical solutions for travel distance, in these models avalanche motion is generally routed downslope over the existing terrain, with deposition where V drops to zero. These models also often account for lack of flow contact in overhangs and momentum loss at sudden decreases of slope angle. Howard (1990) showed that chutes can be created either through the action of debris motion triggering failure of partially weathered bedrock or regolith under conditions where they would otherwise be stable or through direct scour of the substrate.

Bornhardts

Bornhardts are steep-sided, domical hills with substantial surfaces of exposed rock (Willis 1936; Fig. 8.3). In detail they vary considerably in form from being nearly perfect hemispheres, through cylinders with domed tops, to elongated ovoids. These different forms have been given a variety of local and general descriptive names (see Twidale 1981, 1982a,b).



Fig. 8.3 Gross Spitzkoppe, Namibia, is a granite bornhardt which rises 600 m above the surrounding desert plain. Sheeting is occurring on the main face and at depth in the central section which is in deep shadow. On the *right*, sheets are being subdivided by cross-joints (photo M.J. Selby)

The name bornhardt honours the German geologist Wilhelm Bornhardt who gave evocative accounts of the landscapes of East Africa where granite domes stand above extensive plains (Bornhardt 1900, Willis 1936). The association of granite with isolated domical hills standing above plains is clearly implicit in the original designation, but hills with similar forms occur on other rock types of which Ayres Rock and the Olgas of central Australia are among the best known. Ayres Rock is formed of a coarse-grained arkose – that is, a sandstone rich in feldspars. The Olgas, on the other hand, consist of a massive conglomerate, the matrix of which is as resistant as the boulder- and gravel-sized clasts (Twidale 1978) (Figs. 8.3 and 8.4).



Fig. 8.4 Domes of the Olgas, central Australia, showing the effect of major joints in subdividing the massif into elongated domes (photo M.J. Selby)

The outstanding feature of bornhardts is their massive form with few major or continuous joints passing through the body of rock, but the margins of bornhardts can often be seen to be determined by bounding joints. Large rock masses may be partitioned by widely spaced joints which then separate the mass into a series of domes. Such partitioning can be seen in the Pondok Mountains of the Namib Desert (Selby 1982a) and in the Olgas (Fig. 8.4). Sheeting, with separation of curved plates of rock conformable with the dome surface, is a feature of many bornhardts. It is easily recognized that such jointing will permit survival of the dome form even after successive sheets are broken by cross joints, and weathering has caused disintegration of the resulting slabs. Sheet jointing, however, is not always concentric with the dome surface, and where it has a curvature with a shorter radius or longer radius than that of the dome, the result will be a steepening or flattening of the ultimate dome form.

Bornhardts have been reported from virtually all climatic zones, but they are especially abundant in the humid, subhumid, and arid landscapes of the Gondwana cratons where large exposed intrusions of granitic rocks are common. In areas of younger granitic and gneissic intrusions, granitic domes form the cores of mountain massifs and may be revealed by valley erosion along their flanks, as in the Yosemite Valley, California, or as growing intrusions in areas of active uplift, as in Papua New Guinea (Ollier and Pain 1981). In other continental settings diapiric intrusions form domed hills, as at Mount Duval, Australia (Korsch 1982). Such features, and the recognition that granite domes are created as part of the intrusion process and revealed as cover beds are removed, as in the inland Namib escarpment (Selby 1977), indicate clearly that some dome forms are entirely structural in origin and have forms maintained by sheeting processes (Fig. 8.5).

There is no evidence, or process known, for a structural origin for domes of sedimentary rock; for them, and possibly for some domes of igneous rock, other mechanisms have to be considered. It is widely recognized that rounded core stones develop from cuboid joint blocks within weathering profiles. On much larger scales, it is often assumed that similar rounding of corners will form domical rock masses within weathering profiles. Such processes may be aided, or even made possible, by stress release joints opening along the edges of large blocks and thus promoting the formation of rounded forms.



Fig. 8.5 Domes of granite being stripped of their cover rocks which are unweathered schists. This exposure is one of the best pieces of evidence for the creation of domes by intrusive emplacement and the inheritance of that form in the landscape as cover rocks are removed by erosion (photo M.J. Selby)

Construction of railways, roads, and mine tunnels (Boyé and Fritsch 1973) has revealed that many domes occur as compartments of solid relatively unjointed rock within a deeply weathered regolith. The widely accepted hypothesis that many bornhardts are resistant bodies of rock which have survived attack by subsurface weathering and then been revealed by erosional stripping of the regolith (Falconer 1911) is clearly demonstrable where domes stand above erosion surfaces below which deep regoliths survive. For example, in central Australia the exposure of Ayres Rock and the Olgas can be put at the end of the Mesozoic, yet these inselbergs stand above plains that have silcretes of early to middle Tertiary age at the surface (Twidale 1978).

Much of the controversy about bornhardts has been concerned with why bodies of rock should be relatively free of major joints while the rock around them is more closely jointed and therefore subject to penetration by water and chemical alteration. The focus of the debate has been on the lack of joints which has commonly been ascribed to the massive bodies being in a state of compression due to horizontal stresses in the crust (e.g. Twidale 1981, 1982a,b). The arguments, however, will be convincing only when stress levels in the rock are measured. Standard methods exist (Brady and Brown 1985) and have been used by mining engineers for some time. The difficulty is that the material around the bornhardt usually has either been removed by erosion or been deeply weathered, so comparable data cannot be obtained from the dome and its surroundings. In the Namib Desert, however, emerging domes and their surrounds are available for study.

It is evident from the above discussion that domed hills are not unique to any climatic environment. They are found in many of the world's deserts, but primarily because these lie on the surfaces of cratons which have been deeply weathered and stripped. The distinct feature of domes in deserts is their excellent exposure, with limited weathering of their surfaces by modern processes because the domes shed to their margins such rain as falls on them. Even the tafoni, pits, and other superficial features are usually attributed to weathering within the regolith. The process of stripping, whether it be by downwearing or backwearing of the surrounding material is irrelevant to understanding the origins of the domes. Similarly, the mineralogy and petrology of the rock is relevant only in so far as any bornhardt that survives has had to resist the processes acting on it.

Karst in Siliceous Rocks

Karst forms, possibly inherited from past, more humid environments, may be important parts of some present desert environments. Such an area is the Bungle Bungle of the south-eastern Kimberley region of north-western Australia (Young 1986, 1987b, 1988) and similar areas have been reported from the Sahara (Mainguet 1972). In both of these areas the rocks are quartzose sandstones and quartzites.

The siliceous rocks of the south-eastern Kimberley region are nearly horizontally bedded and of Devonian age. Their karstic features include complex fields of towers and sharp ridges, cliff-lined gullies, steep escarpments with cave and tube systems and conical hills standing above flat-floored valleys which debouch on to pediments which may have near right-angle junctions with the scarps. The local relief between the pediment surface and the ridge crests is seldom greater than 300 m (Fig. 8.6). The present climate is semi-arid with a dry season lasting from April to November and a wetter season occurring between December and March. The best estimate for the annual rainfall is 600 mm. On pediment surfaces, to the south, are widely spaced sand dunes now fixed by stable vegetation but indicating a formerly dry climate; the last arid phase ended 10,000–15,000 years BP (Wyrwoll 1979).

Studies by Young (1988) have indicated the processes which have contributed to, or perhaps controlled, the development of the present landscape. The primary evidence comes from scanning electron



Fig. 8.6 Conical hills and sinuous ridges of the Bungle Bungle, northern Australia, rising above a pediment (photo M.J. Selby)

microscope analyses of the rocks. The sandstones have quartz grains cemented by quartz overgrowths which did not eliminate all primary porosity. As a result water could circulate through the rock and etch the grains and dissolve much of the overgrowth silica. The sandstones have, in some formations, been left as granular interlocking grains with few cementing bonds.

The rock bodies now have relatively high compressive strength (35–55 MPa), as applied loads are transmitted by point-to-point contacts between grains, but very low tensile or shear strengths because of weak cementation. The sandstones are now friable and are readily broken down into small blocks or single grains by sediment-laden water. On cliff faces, bedding forms stand out clearly where small-scale fretting and granular exfoliation has undercut more coherent rock units (Fig. 8.7). Stream channels have obviously followed major joints and, in many canyons, streams have undercut valley walls. Below cliffs and escarpments the boundary between the cliff and the pediment is extremely sharp and has angles as high as 90°. Sheet flows crossing the pediments can readily transport the dominantly sand-size grains carried by wet-season flows.

Stream flow and sheet wash are clearly active and the presence of streams in the month of May suggests substantial storage of water in the rocks and the possibility of sapping as a significant factor in slope development. Case hardening is common on most of the slopes to depths of 1–10 mm, but whether the deposition of silica and clays to form the casing is still active or is a relic from former dry periods is unclear. There are a few sites in the Bungle Bungle, and in Hidden Valley near Kununurra to the north (see



Fig. 8.7 Kings Canyon, central Australia, with bedding forms of the sandstone exposed (photo M.J. Selby)

Young 1987b), where thin slab exfoliation has occurred in the relatively recent past and left bare unfretted faces which do not show evidence of case hardening; this may indicate either inactivity of case hardening or just very slow hardening.

The evidence for solution and weakening of the rocks is very clear, but the period over which this has occurred is much more difficult to establish. Caves and dolines in sandstone have usually been attributed to removal of silica in acidic solutions under tropical humid conditions (Pouyllau and Seurin 1985) where there is an abundance of organic molecules. Palaeoclimatic data from Australia indicate that much of the continent had a rainforest cover under which deep weathering profiles and lateritized surfaces developed until at least the middle Miocene when grasslands became common (Kemp 1981). However, laboratory studies of the solubility of silica show that this is greatest when pH is high and that it is enhanced by high chloride concentrations (Young 1987a). Potassium chloride crystals have been found in close association with etched quartz grains in the Kimberley sandstones. Whether much, or all, of the solution took place in humid or in arid conditions is not clear, but it seems probable that it may have occurred in both of these extreme tropical environments.

Some of the landforms described in this section have very similar forms to those found closer to the heart of the Australian arid zone near Alice Springs, where annual rainfall is in the range of 200–300 mm (Slatyer 1962). The horizontally bedded sandstones of Kings Canyon area (Fig. 8.7), for example, exemplify a joint-controlled incised landscape in which the canyon walls have a generally convex form and stand above flat-floored valleys and pediments.

Local controls by hard bands in the rock create a waterfall and plunge-pool long profile to channel beds. Local rockfalls are of greater significance here than they are in the Bungle Bungle. Whether or not solution of the silicate rocks has played any part in the evolution of these landforms is unknown.

The case for calling landforms developed under dominant solution processes ‘karstic’ has been made by Jennings (1983). They are relatively common in the tropical zone of the Gondwana continents and are of significance in several desert areas. The extent to which they are wholly or only partly inherited from humid environments is unclear.

Etch Processes and Inherited Forms

Discussions of the development of hillslopes in terms of rock resistance to modern processes and of the effect of modern processes are inevitably limited in their relevance to slopes which may have a history of 100 m.y. or more. The idea of very long periods for landscape development in the core zones of the continents is not new and has a history which can be traced back at least to Suess (German edition, 1885–1909; English edition, 1904–1924) who introduced the word ‘shield’ for describing the exposed nuclear core of each major land mass. The addition of the word ‘craton’, by Stille (1924) to distinguish the long-stabilized foundations of Precambrian age of every continent, added to the recognition of great geological age for extensive parts of all continents. Only the platforms of relatively undeformed sedimentary rocks deposited predominantly during marine transgressions, the marginal mobile belts, and the hotspot-generated zones of volcanism and rifting disturb the pattern of continental stability. The relevance of these comments to desert hillslopes is that many of the world’s most extensive deserts are on cratons and may therefore have had upland masses with a long evolutionary history for their hillslopes.

Remote sensing has given a clear indication of both the extent of bedrock exposures and of features veneered by thin cover beds on desert cratonic surfaces (for example, Brown et al. 1989; Burke and Wells 1989). Broad swells and undulations of cratonic surfaces, and continental rifting with elevation of rift margins may be attributable to mantle convection. It

has been suggested by Fairbridge (1988) that tectono-eustasy related to seafloor spreading has given rise to a thalossostatic condition linked to a biostatic regime (Erhart 1956) marked by a worldwide moist and warm climate with low relief, continuous vegetation cover, abundant non-seasonal rainfall, and strong biochemical weathering. This thalossostatic condition alternated with a disturbed or rhexistatic state in which high-standing continental surfaces (epeirostatic conditions) with low base levels, rejuvenation of fluvial systems, land erosion, seasonal rainfalls, zonally contrasting climates, formation of deserts, and monsoonal weather patterns were major features. Cratonic regions, according to this concept, tend to be characterized throughout geological history by alternation of chemical leaching, weathering, and duricrust formation with periods of erosion, desiccation, and offshore deposition.

Whether or not this grand but simple scheme is valid, it does add a possible explanation for the clear evidence of periodic stripping of weathering mantles from cratonic surfaces with the consequence that many hillforms, and especially those of inselbergs and related forms, have features that have been attributed to weathering beneath deep regoliths, followed by erosional stripping of the mantles and exposure of bedrock.

The idea of nearly world-wide cratonic weathering mantles was most vigorously expressed by Budel (1957) and has subsequently been vigorously espoused as the basis of many geomorphological explanations by Ollier (1988), Thomas (1989a,b), and Twidale (1990). The essence of the theory is that deep weathering mantles are developed through the action of meteoric waters penetrating along joints and fissures and progressively forming a mantle of saprolite which may extend to depths of tens or even hundreds of metres. The processes of alteration are fundamentally chemical, with the production of a residual soil that is leached of soluble material and depleted of the colloids removed by drainage waters. That the saprolite is formed without loss of volume is indicated by the survival in it of relict joints, veins of quartz and other structural forms derived from the original bedrock. The loss of mobile elements leaves a low-density residuum.

Weathering mantles commonly develop two major zones: an upper unsaturated zone which has a red coloration and is oxidized, and a lower saturated zone

which has white to pale green coloration and a reducing environment. The boundary between the two zones is the water table which fluctuates in level seasonally or as a result of storms and droughts. At depth, a zone is reached where joints are closed, water cannot penetrate, and therefore weathering cannot operate; this is the basal weathering front. All that is required for deep weathering is deep penetration of ground water in the liquid state (i.e. not frozen) and time. There is no necessary relationship between climate at the ground surface and chemical weathering at depth. At depths ranging from about one to ten metres, depending upon climatic zones and penetration of fresh water, soil temperatures are constant at the mean annual surface temperature. At greater depths the geothermal heat flow controls soil temperatures. The availability of groundwater depends mostly on fresh inputs from percolation but is commonly available as 'old' water in arid and semi-arid zones.

It has been pointed out by Habermehl (1980) and Ollier (1988) that the Great Artesian Basin of Australia has groundwater extending to depths of 3,000 m, so weathering by hydrolysis can also extend to this depth. Stable isotope studies show that the water is of meteoric origin (not connate) and flow rates indicate that it takes two million years for the groundwater to flow from recharge areas to discharge zones in artesian springs. The present land surface is a desert, but weathering of the basal rock is occurring under the control of geothermal heat and Pliocene, or early Pleistocene, meteoric water. The Great Artesian Basin is, no doubt, an extreme example but it illustrates the general principle.

Deep weathering profiles have often been associated with current humid tropical climates, perhaps because the surfaces of cratons which are now within the tropics have many land surfaces formed upon deeply weathered mantles. There is increasing evidence that some weathering profiles are of great age; in Australia, for example, Mesozoic and early Cenozoic ages are recognized (Idnurm and Senior 1978). More widely it has been recognized that relics of deep weathering profiles occur in areas which are far from the tropics. Hack (1980) noted saprolite in the Piedmont and Blue Ridge areas of the Appalachians with an average depth of 18 m and a maximum depth which may exceed 90 m; Hall (1986) identified preglacial saprolites in Buchan, Scotland, up to 50 m deep, and in the Gaick area of the Grampian Mountains up to 17 m (Hall 1988); Bouchard (1985) has found saprolites up

to 15 m thick at protected sites covered by Quaternary ice sheets in Canada.

The relevance of all of these observations for an understanding of hillslope development in the modern desert zones is that many features now visible may have developed at the basal weathering front and later been exposed at the ground surface as regolith was stripped in periods of rhexistasy.

Etch Forms of Deserts

The recognition that boulders, and other minor and major forms, have developed within weathering mantles and at the weathering front has a long history which goes back to the beginning of modern geology (Twidale 1990). These forms when exposed by mantle stripping are collectively known as etch forms (Wayland 1933). Weathering fronts are particularly sharp in granitic and other crystalline rocks of low permeability. The front is more diffuse, and essentially a zone, in the weaker sedimentary rocks. Crystalline rocks commonly form the bedrock of cratons, and the various associations of boulders recognized as boulder-strewn platforms (Oberlander 1972), tors or koppies (Fig. 8.8) are readily identified as remnants of corestones because corestones are common in exposures through weathering mantles. Perhaps more significant was the proposal by Falconer (1911) that the inselberg landscapes of northern Nigeria were shaped not by epigene processes but by chemical action of waters acting at the weathering front. Furthermore, he recognized that the

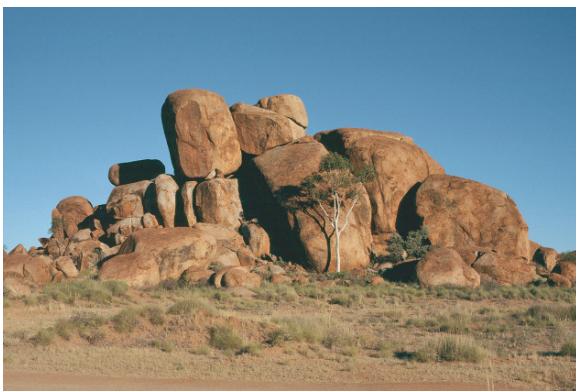


Fig. 8.8 Devils Marbles, central Australia, which have been interpreted as residual etchforms developed within a weathering mantle which has now been stripped (photo M.J. Selby)

variations in depth of penetration of the front were controlled by the spacing of joints in the bedrock. The result of mantle stripping was therefore an irregular land surface with inselbergs standing above plains with residual mantles of varying depth. Such ideas were elaborated into a system of geomorphological evolution of the cratonic surfaces by Budel (1957, 1982).

The evidence in favour of the concept of etchforms developing on cratonic surfaces is now very strong. It indicates the development of currently exposed land surfaces over periods of 100 m.y. or more. This timespan far exceeds the period of existence of the world's major deserts which are mid to late Tertiary in origin (see Selby 1985 for a review). The obvious conclusion is that many of the major hill mass and plain landforms had an origin beneath deep weathering mantles of ancient origin. Subsequent mantle stripping has exposed them to subaerial processes of weathering and erosion. In the time since exposure, denudation of bare rock has been sufficient for many hillslopes to achieve slope angles which are in equilibrium with the mass strength of their rocks, but on hillslopes formed of massive rocks with little pre-existing jointing the possible controls on form include survival of etch forms; survival of the original form through stress release joints developing nearly parallel to the original rock surface; the development of cross fractures which permit an incomplete adjustment to the slope angle controlled by rock mass strength; survival of structural influences and controls; dominance by solution processes; control by talus covers and formation of Richter-denudation slope units; undercutting of upper slope units by sapping, stream action, or recession of an underlying weak rock unit.

Slopes in Layered Rocks

Many desert areas are underlain by generally flat-lying sedimentary rocks of varied composition, sometimes intermingled with tabular intrusive or extrusive volcanic rocks. Examples include the Colorado Plateau in the south-western United States, North Africa, the Arabian peninsula, and portions of the other major deserts. The erosion of such sedimentary sequences creates a landscape of scarps or cuestas capped by the more resistant rock units. By contrast, a few desert areas, such as the Zagros Mountains region of Iran, have

complex patterns of cuestas, hogbacks, strike valleys, etc. developed in strongly folded or faulted layered sedimentary rocks. The discussion here will focus on the simple cuesta landforms of flat or inclined beds, although the general principles are applicable in areas of more complicated structure.

The classification of rock slope types introduced by Selby (1980, 1987) and Moon (1984b) can also be applied to cuesta landforms. Figure 8.9 shows a classification of slope elements on an eroded anticline in layered rocks. Strength equilibrium slopes (shown by =) occur primarily on cliff faces eroded by small rockfalls, but where such cliffs are actively eroded by stream undercutting (+U+) they may be steeper than strength equilibrium. Scarp faces dominantly eroded by landslides may be less steep than strength equilibrium and those dominantly wasted by rockfalls involving the entire cliff face may be steeper than strength

equilibrium. Structurally controlled slopes (-S-) occur where erosion of weaker layers exposes the top surface of resistant rocks. Normal slopes in weak rock (typically badlands with shallow regolith) are shown by -E-. Rampart slopes in weak rocks below cliffs of resistant rock will generally be talus (-D-) or Richter slopes (not shown).

In a general sense cuestas are an automatic adjustment of the landscape to permit rocks of varied erosional resistance to be eroded at roughly equivalent rates (Hack 1966). When a resistant rock layer is first exposed the erosion rate diminishes on the exposed top of the resistant layer as the overlying weaker rocks are removed, often creating lithologically controlled near-planar upland surfaces called stripped plains. As the resistant layer becomes more elevated relative to surrounding areas of weaker rocks, the caprock is eventually breached, exposing underlying weaker rock units,

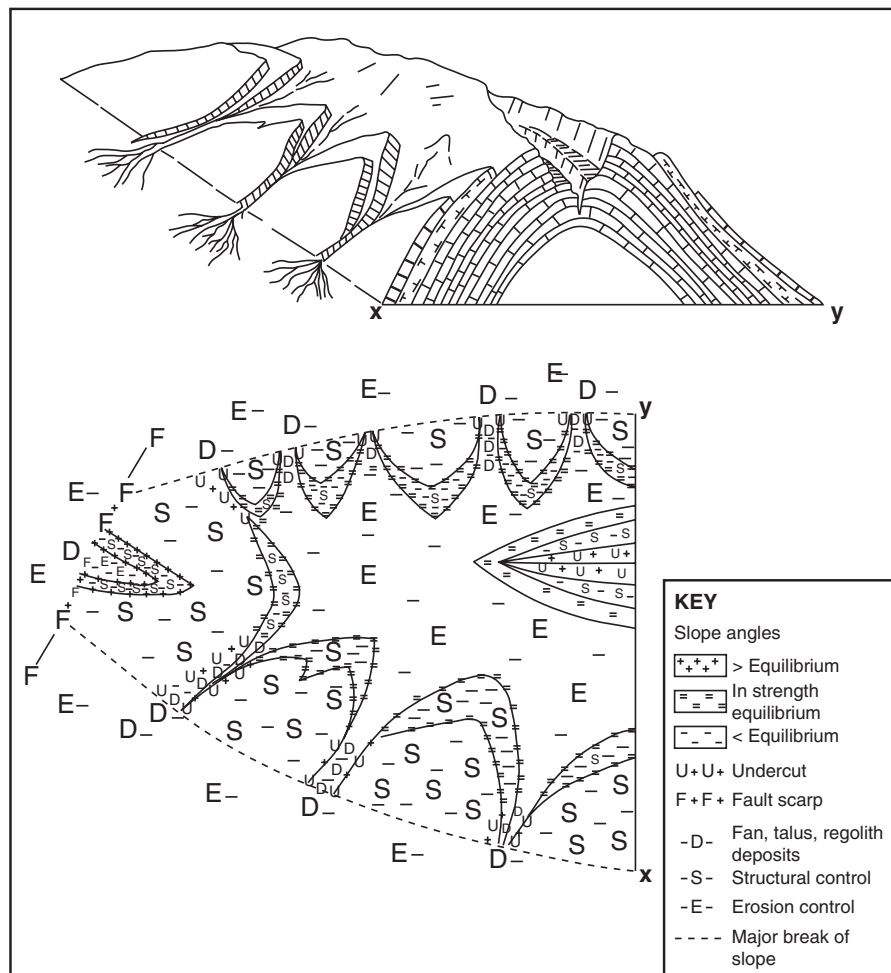


Fig. 8.9 Identification of landform elements on cuestas in folded rocks (from Selby 1987, Fig. 15.13)

whose rapid erosion creates a steep scarp and accelerated caprock erosion by virtue of the steep gradients relative to the superjacent stripped plain. This discussion of such scarps is organized into two general headings: evolution of scarps in profile and evolution of scarp planforms. Examples are taken primarily from the Colorado Plateau, south-western United States. Additional illustrations of the features described here and a road log of sapping and geologic features are given in Howard and Kochel (1988).

Evolution of Scarps in Profile

Resistant sandstones, limestones, and volcanic flows and sills in desert areas are generally exposed as bare rock slopes except where mantled with aeolian sands or alluvium. However, areas of very low relief, such as stripped plains, may be mantled with sandy, cobbly, poorly horizonated soils and scrubby vegetation. Two morphological end members characterize the resistant rock exposures, low to high relief slopes in massive rock (in particular, the rounded sandstone exposures termed 'slickrock') and cliff or scarp slopes developed where the caprocks are being undermined. Emphasis in this section is placed on scarp-front processes and morphology. An intermediate landform type, termed

segmented cliffs by Oberlander (1977), will also be discussed.

The more readily weathered and eroded rock units, generally shales or poorly cemented sandstones or alluvium, are commonly eroded into badlands where they are thick (Chapter 10). But in areas with thick interbedded caprock-forming units the easily eroded strata are usually exposed primarily on the subcaprock slopes or ramparts. Terms used here to describe the characteristic parts of an escarpment are shown in Fig. 8.10. Several alternative terms have been used to denote the rampart, including subtalus slope (Koons 1955), debris-covered slope (Cooke and Warren 1973), footslope (Ahnert 1960, Oberlander 1977), lower slope (Schmidt 1987), and substrate ramp (Oberlander 1989).

The exposure of weaker strata (generally shales or poorly cemented and/or highly fractured sandstones) beneath massive sandstones causes undermining of the sandstone, leading to cliff development and rapid scarp backwasting. A far greater volume of rock is initially broken up by scarp retreat than by erosion on backslopes when considering average rates over large areas. Because of the rapid retreat of scarp slopes, the cliffs generally eat back into pre-existing backslopes. Figure 8.11 shows an example where updip exposure of shale near the stream level (right side of picture) has caused development of cliffs and their backwasting into sandstone slickrock slopes; such undermining was

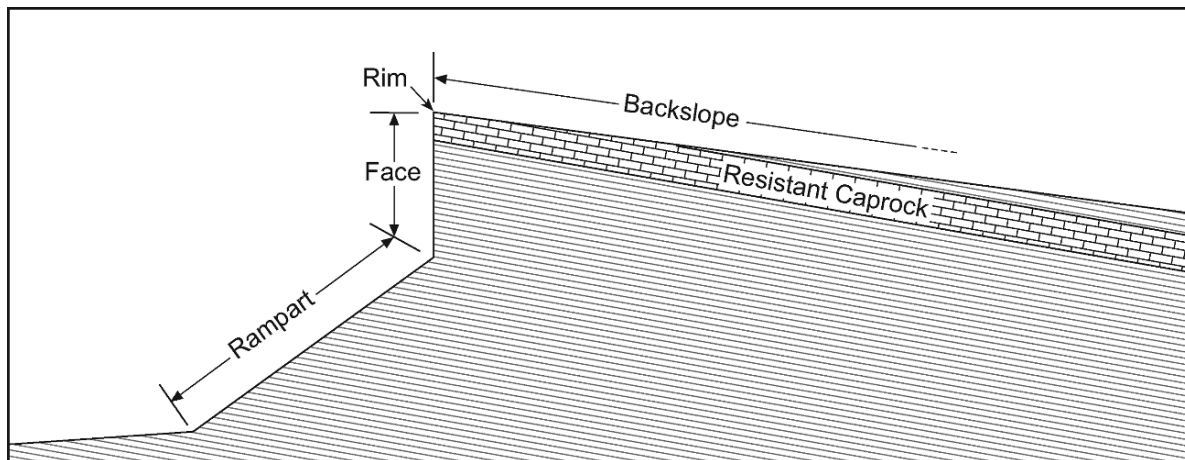


Fig. 8.10 Escarpment components. The face is the vertical to near-vertical cliff developed at the top of the escarpment mostly in the caprock but occasionally extending into the non-resistant layer beneath. The rampart is a slope of lesser inclination

extending from the bottom of the face to the base of the escarpment which is commonly partly mantled with debris from backwasting of the face (from Howard 1970, Fig. 55)



Fig. 8.11 Landforms of De Chelly Sandstone north of Kayenta, Arizona along US Highway 163. Slickrock slopes to left give way to vertical scarps at right where undermining is active due

to exposure of shaly Organ Rock Tongue at the base of the scarps (photo A. Howard)

discussed by Ahnert (1960) and Oberlander (1977). The relief developed on scarp backslopes depends upon the erodibility of the caprock unit compared with overlying units, the thickness of the caprock, the types of weathering and erosional processes acting on the exposed rock, and rates of basal level lowering.

Processes of Scarp Erosion

Wasting of caprocks occurs primarily by rockfall, block-by-block undermining, and slumping. Some sandstone caprocks (e.g. the Morrison Formation of the American South-west) are undermined block-by-block by weathering and erosion of the underlying shale without rapid fall of the undermined blocks (Fig. 8.12). The blocks may be repeatedly lowered

with little downslope sliding or rolling, but typically the blocks slide and occasionally roll a short distance upon being undermined. Block-by-block undermining requires a relatively thin caprock, well-developed jointing, and shale that weathers easily by addition of water (e.g. the smectitic Morrison Shale). Slumping is prevalent on relatively few escarpments, where it may dominate as the mechanism of scarp retreat (Fig. 8.13). The Toreva block slumps are a classic example (Reiche 1937), and other examples have been discussed by Strahler (1940) and Watson and Wright (1963). Conditions leading to slumping failure have not been firmly established, but a low shear strength of the unweathered subcaprock unit is probably the major factor. Low shear strength can result from low bulk strength or a high degree of fracturing and/or abundant bedding plane partings. Other factors may be deep



Fig. 8.12 Block-by-block undermining of channel sandstones in shales of Morrison Formation, near Hanksville, Utah (a and b) (photo A. Howard)



Fig. 8.13 Slump failure of Morrison–Summerville escarpment near Hanksville, Utah (photo A. Howard)

weathering of the subcaprock unit by groundwater flow and high pore water pressures. Landslide and rockfall processes were summarized in Anderson and Richards (1987) and Brunsten and Prior (1984).

Rockfall is the most common form of scarp retreat (Fig. 8.14), involving events ranging from calving of individual blocks to the failure and fall of a wide segment of the face, resulting in a rock avalanche on the scarp rampart. Debris produced by rockfalls with high potential energy may result in powdering of a large percentage of the original rock (Schumm and Chorley 1966), but on most scarps the coarse debris produced by the rockfalls must be weathered and eroded before further scarp retreat can occur (Fig. 8.15). Weathering processes acting on the debris are similar to those occurring on slickrock slopes, including splitting or shattering, granular disintegration, and solution of cement (or the rock *en masse* in the case of limestones). The necessity for weathering of scarp-front debris before further erosion of the subcaprock unit leads to a natural episodic nature of rockfalls and scarp morphology, as outlined by Koons (1955), Schipull (1980), and Schmidt (1987) (Fig. 8.14). Where caprocks are eroded primarily by large rockfalls continued erosion of the subcaprock unit at the margins or base of the rockfall eventually raise the debris blanket into relief, sometimes forming subsidiary small escarpments where the debris blanket is subject to further mass wasting. Thus old rockfalls stand well above surrounding slopes of both exposed subcaprock unit and younger rockfalls. When these old rockfalls are contiguous with the scarp face, they prevent the development of high relief at the cliff face, and thereby inhibit further rockfalls until the talus



Fig. 8.14 Portion of escarpment of Emery Sandstone over Mancos Shale on North Caineville Mesa near Hanksville, Utah, along Utah Highway 24. Note extensive deposits of rockfall debris that are dissected by continuing erosion of shale on rampart. Cliff in alcove extends far into the underlying Mancos Shale. Seepage lines are present at the base of the Emery Sandstone in the alcove. High debris blankets the *right* and *left* of the alcove that extend upward to the base of the sandstone are strongly dissected at their base. Presence of the debris blanket inhibits further scarp backwasting until the debris is weathered and eroded. These high debris blankets are probably a Bull Lake equivalent (early Wisconsinan?) pluvial deposit. The picture is taken from the top of a sandstone debris-capped Bull Lake pediment that formerly merged with the dissected talus deposits. The Bull Lake pediment formerly extended behind the photographer to merge with Fremont River gravels (now terraces about 60 m above present river level)

is weathered and eroded. In some cases the talus at the foot of the rampart becomes isolated from the scarp face as erosion of the subcaprock unit continues, forming a talus flatiron (Koons 1955, Schipull 1980, Gerson 1982, Schmidt 1987). Talus flatirons thus formed have been termed non-cyclic flatirons as contrasted with similar features resulting from climatic fluctuations (Schmidt 1989a) (Chapter 22). Schmidt (1987, 1989a) suggested that flatirons are best developed in scarps with heterogeneous subcaprock strata, including beds of variable resistance and slope inclination. However, flatirons are also well developed on the scarps of the Colorado Plateau composed of massive sandstones over homogeneous marine shales. Gerson and Grossman (1987) noted that flatirons are absent on desert scarps lacking a strong caprock over a weaker subcaprock unit.

Most prominent scarps on the Colorado Plateau are formed of massive sandstone underlain by shale or other easily weathered rock, so that backwasting is caused by a combination of weathering of exposed



Fig. 8.15 Recent rockfall in Navajo Sandstone in the Inscription House area of the Navajo Indian reservation showing abundant rockfall debris (photo A. Howard). Cliff above rockfall is approximately 30 m tall

caprocks (e.g. off-loading fracturing, freeze–thaw, and groundwater sapping) and loss of bulk strength of the underlying layer accompanied by erosion of the scarp rampart. However, some of the incompetent layers producing scarps are strong in bulk but are eroded primarily because of denser fracturing relative to more massive (but not necessarily stronger) overlying sandstones (Oberlander 1977, Nicholas and Dixon 1986). Creep of subcaprock shales has been implicated in breakup of caprocks (block gliding) in humid environments (e.g. Zaruba and Mencl 1982) but has not been noted on desert scarps. It is possible that slight creep or off-loading expansion in shales may locally be a factor in development of off-loading fractures in overlying sandstone caprocks. Gravity-induced creep of evaporite beds has been suggested as the mechanism responsible for creating the miniature horst and graben structure of the Needles District in Canyonlands National Park, Utah (McGill and Stromquist 1975).

Although rockfall and slumping are the major transport processes in scarp backwasting, weathering and erosion by groundwater commonly can be as, or more, important in weakening the scarp face than is undermining by erosion of the rampart. The role of groundwater sapping is discussed separately below.

The amount of caprock talus exposed on scarp ramparts depends in part on the planform curvature of scarps, being greater in re-entrants where caprock debris converges on the lower rampart and lesser in front of headlands or projections (or around small buttes)

where debris is spread radially. The amount of debris is also controlled by spatial variations in rates of scarp retreat (generally higher at the head of re-entrants) and by the volume of caprock eroded per unit amount of backwasting, which is higher in re-entrants and lower at headlands.

Scarp ramparts are eroded by a variety of processes, including normal slope and rill erosion where the subcaprock unit is exposed and weathering and erosional processes act on caprock detritus. Weathering processes on talus include frost and/or hydration splitting or shattering, spalling due to salt crystallization (primarily on unexposed surfaces), granular disintegration, and solution. The relative mix of these weathering processes depends upon the size and composition of the talus. Calcareous-cemented sandstones, common on the Colorado Plateau, are primarily shattered in large blocks, but succumb to granular disintegration in small blocks, yielding easily eroded sand-sized detritus. On the other hand, siliceous-cemented sandstones do not weather by granular disintegration, and large blocks of caprock commonly remain behind as the scarp retreats (Fig. 8.12b).

Where caprock debris is copious or very resistant, the talus material may be reworked several times before its final removal due to continuing erosion of the subcaprock unit, forming elevated blankets eroded at its margins as subsidiary scarps (e.g. the flatirons discussed above), development of individual rocks on subcaprock unit pedestals (damoisselles), or less dramatic undermining and rolling of individual boulders. Slope-wash and gullying are important in removing sand- to gravel-sized weathering products, and some talus blankets are extensively modified by, or even emplaced by, wet debris flows (Gerson and Grossman 1987).

Absolute and Relative Rates of Scarp Erosion

A variety of methods has been used to date regional rates of scarp retreat in desert areas, including archaeological dating (Sancho et al. 1988), beheading of consequent valleys (Schmidt 1980, 1989b), stratigraphic relationships with dated volcanic or sedimentary deposits (Lucchitta 1975), and age of faulting initiating scarp retreat (Yair and Gerson 1974). Inferred retreat rates vary over two decades from about 0.1 to 10 m

1000 y^{-1} (Schmidt 1988, 1989b, Oberlander 1989). The primary long-term controls over erosion rates are rate of base level lowering and rock dip, as discussed below. However, short-term erosion rates are strongly influenced by climate variations and climate-related changes in local base level.

The relative rates of erosion on different parts of cuestas can be illustrated by considering, as a first approximation, that the form elements maintain a constant gradient and a constant position relative to the stratigraphic layers through time. These assumptions require a constancy of both stream erosion and slope processes through time, which probably approximates the long-term average behaviour of scarp erosion but not the short-term changes due to climatic fluctuations. The assumption of a constant position of slope elements relative to the stratigraphy (Fig. 8.16b) is clearly a closer approximation to scarp evolution than the assumption that slope elements retain a constant position through time (i.e. a constant rate of vertical erosion on all elements of the scarp (Fig. 8.16a)).

In horizontal stratified rock these assumptions predict a rate of vertical erosion proportional to the slope tangent, whereas the horizontal rate of erosion (lateral backwasting) is identical on all slope elements (Fig. 8.16b). This implies an infinite rate of downwasting for a vertical cliff, which is an artifact of considering cliff retreat as continuous erosion rather than as discrete events such as rockfalls. Therefore, as mentioned above, on a typical escarpment the downwasting of the

slickrock slope on top of the caprock is very slow compared with both cliff retreat and vertical erosion below the rim.

The relative rates of erosion on various slope elements are also affected by the structural dip. If all form elements erode at an equal rate parallel to the structure (that is, in a downdip direction) with constant gradient, then the instantaneous rate of vertical downwasting V_t is given by the structural dip d , the slope angle s , and the rate of downdip backwasting of the escarpment D_t . Where the slope is inclined with the dip (Fig. 8.17a),

$$V_t = D_t(\sin d - \cos d \tan s), \text{ for } 90^\circ > d > s > 0^\circ \quad (8.2)$$

and (Fig. 8.17b),

$$V_t = D_t(\tan s \cos d - \sin d), \text{ for } 90^\circ > s > d > 0^\circ \quad (8.3)$$

Where the slope opposes the structural dip (Fig. 8.17a):

$$V_t = D_t(\sin d + \cos d \tan s), \text{ for } 90^\circ > s > 0^\circ. \quad (8.4)$$

During continued downcutting by streams draining the escarpment, the relief should adjust until downdip exposure of new caprock and updip removal by backwasting are roughly balanced. Therefore the rate of vertical reduction of the rim should be independent of the dip (maintaining a constant relief through time), whereas horizontal retreat of the escarpment would be inversely proportional to the tangent of the dip, and

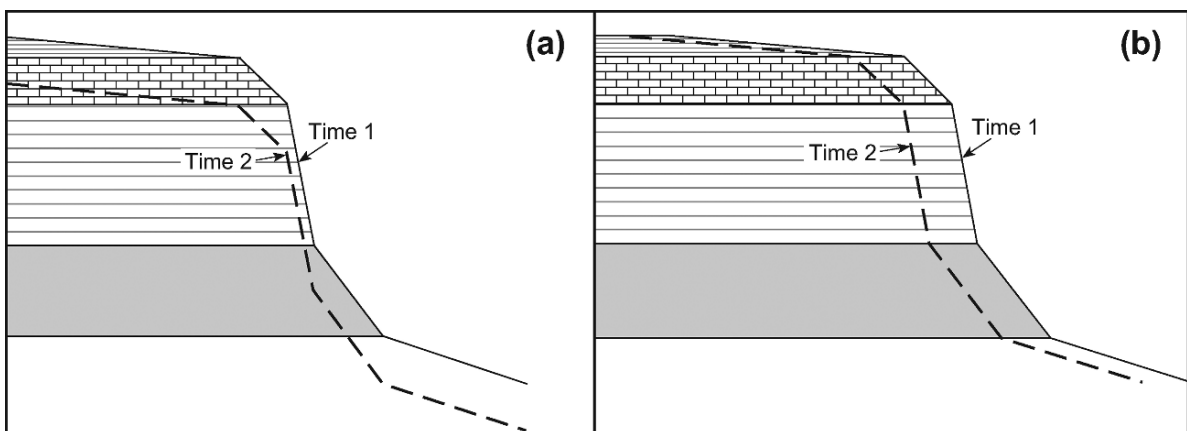


Fig. 8.16 Erosion of a compound hillslope under the assumptions of (a), uniform rate of downwasting on all slope elements, and (b), maintenance of correlations between slope elements and stratigraphic units. Both assume that gradients of the respec-

tive elements remain constant through time. Under the first assumption (a) the volume of material removed increases with decrease of gradient, whereas the volume decreases in (b) (from Howard 1970, Fig. 56)

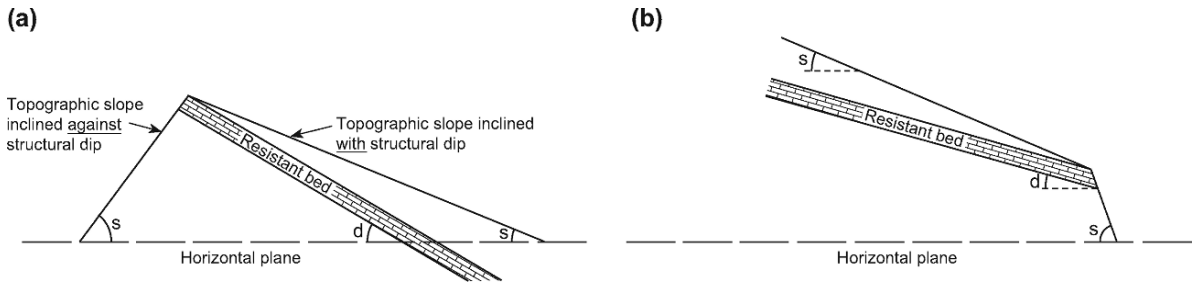


Fig. 8.17 Definition of angles and slope elements on escarpments in dipping strata (from Howard 1970, Fig. 58)

the volume of caprock eroded per unit time would be inversely proportional to the sine of the dip. The very rapid rate of horizontal retreat predicted for low dips does not occur because the escarpment becomes segmented by erosion along drainage lines into isolated mesas and buttes whose local relief, distance from the main escarpment, and rate of backwasting increase through time. Nevertheless, these considerations imply that, in general, a greater volume of rock must be eroded per unit time from gently dipping scarps than from steeper ones. The gradients and total relief on a given scarp should increase where the structural dip decreases to maintain relatively constant rates of vertical reduction of the rim. Figure 8.18 compares relief of escarpments on two sandstones in the Henry Mountains area, Utah, as a function of the reciprocal of the sine

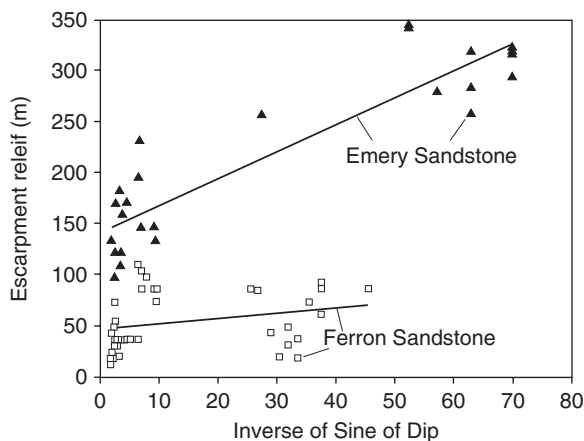


Fig. 8.18 Relief of escarpments in dipping strata as a function of the angle of dip for two sandstones sandwiched between over- and underlying shales. Relief is measured from crest of escarpment to base of scarp rampart (see Fig. 8.10). Each data point is a measurement from a separate location along the western edge of the Henry Mountains, Utah. A least-squares regression line has been fitted to observations from each sandstone scarp (based on Howard 1970, Fig. 59)

of the dip, showing that there is a relationship of the type predicted for the Emery Sandstone. The relationship is poor for the low escarpment of Ferron Sandstone. Areal variations in fluvial incision at the scarp base since the Bull Lake glacial maximum (varying from about 60 m along Fremont River to a few metres in remote locations) may be responsible for the large scatter for this low scarp.

The overall height and steepness of a given scarp should be controlled by one of three factors: (a) development of sufficient relief to trigger caprock mass wasting (which will be related to caprock resistance); (b) the rate of weathering and erosion of caprock debris on scarp ramparts; and (c) the rate of erosion of the subcaprock unit.

Caprock resistance has been suggested as a controlling factor by Schumm and Chorley (1966), Nicholas and Dixon (1986) (who related scarp backwasting rates to degree of fracturing of the lower caprock unit), and Schmidt (1989b) (who found a strong relationship between regional backwasting rates and the product of caprock thickness and a measure of caprock resistance). To the degree that scarps adjust over long time intervals to balance rate of caprock removal with long-term rates of base level lowering (i.e. V_t in Equations 8.2, 8.3, 8.4 is areally uniform), relationships between backwasting *rate* and caprock resistance of the type found by Nicholas and Dixon (1986) and Schmidt (1989b) cannot be universally valid (although *local* variations in back-wasting rates on a given scarp may be related to caprock resistance as found by Nicholas and Dixon (1986) due to exposure of zones of caprock of differing resistance to erosion). Rather, the back-wasting rate D_t will adjust to be a function only of rock dip and regional erosion rate V_t . As suggested above, overall scarp height and steepness adjust to equalize erosion rates, and these should be related to caprock resistance and thickness (the scarps

in the thin Ferron Sandstone are lower than those of the massive, thick Emery Sandstone (Fig. 8.18)). It seems reasonable that a threshold scarp steepness and/or height would be required to trigger the more energetic types of mass wasting processes, such as rock avalanche and landsliding. Block-by-block undermining small rockfalls and caprock weakening by groundwater sapping are less clearly related to overall scarp relief.

Koons (1955) emphasized that accumulated caprock talus protects the caprock from further large rockfalls until it is removed. Koons (1955) and Howard (1970) suggested that the length of the rampart self-adjusts over the long run to provide a surface area sufficient to weather talus at the rate that it is supplied. Schumm and Chorley (1966) introduced the talus weathering ratio, which they defined as the ratio of the rate of talus production from the cliff to the rate of talus destruction on the rampart. They noted that a ratio greater than unity leads to a moribund scarp choked in its own detritus, similar to the models of Lehmann (1933) and others (summary in Scheidegger 1991, pp. 130–4). Schumm and Chorley (1966) suggested that ratios less than unity characterize certain nearly debris-free scarps encountered on the Colorado Plateau. However, over the long run, continuing scarp retreat implies the talus weathering ratio equals unity, since only as much talus can be weathered as is produced (Gerson and Grossman 1987, Howard and Kochel 1988). This concept can be illustrated by letting P equal the volumetric rate of supply of caprock talus per unit time per unit width of scarp, K equal the potential volumetric erosion rate (either weathering- or erosion-limited) of talus per unit area of scarp rampart, and L the required rampart length. Then for balance of addition and removal

$$L = P/K \quad (8.5)$$

However, as discussed below, this ratio may vary considerably as a result of climatic fluctuations as well as local short-term imbalances of P and K , as discussed by Koons (1955). Since talus on ramparts is composed of caprock debris, scarp relief controlled by talus weathering will be indirectly controlled by caprock resistance.

The rate of erosion of the subcaprock bedrock (commonly shales) has been cited as the controlling factor for scarp erosion rates by Gerson and Grossman (1987) and Schipull (1980). Strictly speaking, if the rate of

subcaprock bedrock erosion were the dominant factor in scarp retreat, scarps should not be higher or different in form than other slopes in the subcaprock unit where no caprock is present. None the less, it is true that base level control is transmitted to the scarp via the channels and slopes developed in the subcaprock unit.

It is likely that these three factors vary in relative importance in controlling scarp form from scarp to scarp, from place to place on the same scarp, and through time as climate and/or base level control varies. Scarps that backwaste largely by rockfall are most likely to have planforms controlled over the long run by requirements for weathering of caprock debris, leading to the observation that 50–80% of the scarp front is covered by talus at equilibrium (Gerson and Grossman 1987). As will be discussed further below, many scarps on the Colorado Plateau are quite stable under the present climatic regime, so that the ramparts are nearly bare of talus, and tall cliffs have developed in both the caprock and in the subcaprock shales. Thus both base level control and talus weathering at present have little influence on scarp form, and the backwasting that does occur is largely due to caprock weathering and small rock-falls. On the other hand, for escarpments characterized by block-by-block undermining (Fig. 8.12) caprock resistance is less important than erosion of the subcaprock shales. In fact, where the caprocks consist of hard-to-weather silica-cemented sandstones, caprock boulders may be gradually let down by undermining and rolling while the escarpment continues to retreat, leaving piles of large caprock fragments (Fig. 8.12b).

Schmidt (1989b) noted that very resistant caprocks often include less resistant beds in the subcaprock strata that, in the absence of the more resistant overlying bed, would independently form scarps. In sections of scarp that are linear or indented, debris shed from the overlying caprock largely prevents development of subsidiary lower scarps, but in front of headlands, where rapid retreat of the main caprock and radial dispersal of its detritus leave the subcaprock units largely free of talus, the lower resistant units commonly form low scarps.

Backslopes on Caprock

In the classic case of a scarp composed of a thin resistant layer sandwiched between thick, easily eroded

strata (Fig. 8.10) the caprock is exposed as a low-relief stripped plain on the scarp backslope, and caprock erosion occurs primarily at the scarp face. However, when the caprock is thick, erosional sculpting of the backslope may be as, or more, important than the lateral attack at scarp faces. Landform development on thick, homogeneous rocks is discussed in the Slopes in Massive Rocks section, but processes and landforms on slickrock slopes are outlined here because of their importance in development of segmented scarps.

The striking and unusual slickrock slopes occur on desert sandstone exposures as low, generally rolling relief on bare rock slopes (Figs. 8.11 and 8.19). Hill forms are generally convex to convex-concave and rather irregular due to the prevalence of small-scale structural and lithologic controls exerted by the exposed rock upon weathering and erosional processes. Slickrock slopes occur most commonly on the backslopes of cuestas, but high relief forms occur in thick, massive sandstones such as the Navajo Sandstone. Where strong structural control by jointing or faulting occurs, the fractures tend to be eroded into furrows or valleys, and the sandstone landscape takes on a reticulated or maze-like appearance as at Arches National Park, Utah. Doelling (1985) noted that sandy colluvium collecting along depressions developed on joints accelerates weathering of the sandstone by providing a moist environment; thus the influence of fractures on the topography is enhanced by a positive feedback on weathering rates. Small-scale horst-and-graben development associated with extension caused by flow of underlying evaporites has created the



Fig. 8.19 Polygonal superficial fracturing of Navajo Sandstone along US 160 west of Kayenta, Arizona superimposed upon weathered joints. Individual polygons range from 1 to several meters in size (photo A. Howard)

‘needles’ section of Canyonlands National Park, Utah (McGill and Stromquist 1975).

Slickrock slopes are weathering limited (Carson and Kirkby 1972, pp. 104–6) in that transport processes are potentially more rapid than weathering processes. That is, loose debris is removed from the slopes as fast as it is produced by weathering so that little or no loose residuum covers the bedrock. On slickrock slopes the bedding is emphasized by the grain-by-grain loosening or disintegration of thin surface crusts or whole layers of the sandstone exposed on these weathering-limited slopes, particularly on exposures of the massively cross-bedded Navajo Sandstone (Fig. 8.20). Coarser sand layers with fewer grain-to-grain contacts weather and loosen most readily, aiding differential surface expression of minor lithologic variations (Hamilton 1984). Despite these microscale lithologic controls, the slickrock slopes generally show only minor form control by bedding and the fairly planar to rounded slopes cut across bedding planes (Figs. 8.11 and 8.19).

The major reason for development of smooth, generally convex slopes is the development of ex-foliation or sheeting fractures in massive, poorly jointed (referring here to pre-existing regional or systematic jointing) sandstones such as the Navajo Sandstone (Bradley 1963). The fracturing may be due to stress relief (Bradley 1963), expansive stresses due to weathering of near-surface beds (Hamilton 1984), and possibly freeze–thaw. These mechanisms produce lenticular sheets of thickness ranging from a few centimetres to a metre or so, with more widely spaced



Fig. 8.20 Expression of bedding in slickrock slopes in Navajo Sandstone near Utah Highway 12 west of Boulder, Utah (photo A. Howard)

fractures below these, fading out within 10–20 m from the surface. Thus the exfoliation and sheeting joints form ‘a crude, somewhat subdued replica of the surface form’ (Bradley 1963, p. 521). Although grain-by-grain removal of sand grains loosened by solution of the calcite cement or peeling of thin (<1 cm) weathered rinds seems to be the dominant erosional process, deeper weathering is indicated locally by the presence of shallow jointing perpendicular to the sandstone surface. In massive sandstones these cracks create a network pattern with a scale of 1–5 m that clearly wrap around existing topography, creating an ‘elephant hide’ pattern (Fig. 8.19). Where strong layering is exposed in cross-section, the cracks follow bedding planes and also create fractures cutting across the bedding, forming a ‘checkerboard’ or ‘waffle’ pattern as at Checkerboard Mesa, Zion National Park, Utah. The effective depth of these fractures is probably about 1/5–1/2 their lateral spacing. Hamilton (1984, pp. 32–4) suggested the fractures result from cyclic near-surface volume changes resulting from thermal cycling, wetting and drying, or freeze-thaw.

In addition to exfoliation, any weathering process that acts through some depth from the surface will tend to erode away projecting masses due to the greater surface area relative to volume and lead to a ‘grading’ of surface slopes, with a characteristic scale of action of the same order of magnitude as the depth of weathering (presumably a few centimetres to a few metres). Such processes may include solution of cement, weathering of feldspars and clays, disruption of the rock along microfractures and between grains due to differential volume changes produced by temperature changes, freeze-thaw, or shrink–swell of clays.

Segmented Scarps

Many areas of moderate relief on sandstones on the Colorado Plateau exhibit a complex topography embodying both elements of slickrock morphology and of scarps. Such landscapes developed in the Slick Rock member of the Entrada Sandstone at Arches National Park, Utah, were the object of a comprehensive study by Oberlander (1977). In this area slickrock slopes are interrupted by nearly vertical cliffs which Oberlander termed slab walls due to their erosion by failure along sheeting (off-loading) fractures parallel to the scarp face. The slab walls terminate at their base at inden-

tations developed in thin weak zones (partings) whose weathering and erosion cause the slab wall backwasting (Fig. 8.21). Partings that readily weather (effective partings) are either closely spaced bedding planes with highly fractured sandstone sandwiched in between, or are one or more thin (2–5 cm) layers of fissile ferri-ferrous shale. The partings commonly are of limited horizontal extent, so that slab walls die out laterally (Fig. 8.21). Some slopes may have more than one slab wall where partings occur at two or more levels.

Oberlander presented convincing evidence that slope erosion occurs both by erosion of slickrock slopes and slab-wall backwasting. This, coupled with intersection of new partings and lateral dying out of other partings during slope retreat, leads to progressive changes in slope profile form (Fig. 8.22). In Oberlander’s model the gradient of slickrock slopes below effective partings largely depends upon the relative rates of scarp backwasting by parting erosion and the rate of weathering and erosion on the slickrock slopes, with gentler slickrock slopes associated with rapid parting erosion. Sometimes backwasting at a parting may cease, either due to playing out of the parting or local conditions less conducive to parting erosion. In such cases continued erosion of the slickrock slope below the parting leads to development of a near-vertical slope below the parting; such slopes were called secondary walls by Oberlander. Such inactive slab walls also commonly develop alveolar



Fig. 8.21 Segmented scarp in Entrada Sandstone at Arches National Park, Utah, showing slickrock slopes, slab walls, and effective partings at base of slab walls. Note that the prominent slab wall in the middle of the scarp at left dies out in the right side of the photo due to pinching out of the effective parting (photo A. Howard)

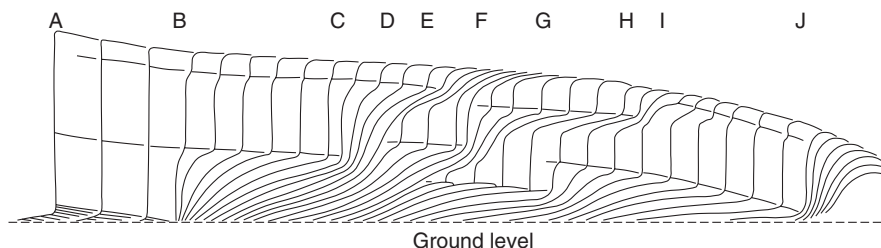


Fig. 8.22 Development of scarp forms in massive sandstone through time and space. For simplicity a constant ground level (*dashed line*) is assumed during scarp retreat, along with equal thicknesses of removal from major slab walls in each unit of time. At A a through-going cliff is present due to sapping above a thin-bedded substrate. At B the thin-bedded substrate passes below ground level, scarp retreat slows, and effective intraformational partings assume control of scarp form. Effective partings

close at C, E, F, G, H, and J, leading to local slab wall stagnation and rounding into slickrock. Partings that open at D, E, F, G, and I, initiate growth of new slab walls. Note that the effect of former partings in rocks that have been removed continues to be expressed in the form of slickrock ramps and concave slope breaks. Lowering of ground level during backwearing would cause cliff extension upward from major contact (caption and illustration from Oberlander 1977, Fig. 8)

weathering, discussed further below. An important conclusion of Oberlander's study is that thin partings in otherwise massive bedrock cause a complicated slope form (in particular the slab wall), so that slope breaks are not necessarily an indication of lithologic contrast above and below the break, but may imply only a thin discontinuity.

Similar slope forms occur in other sandstone units on the Colorado Plateau, especially the Navajo Sandstone and the Cedar Mesa Sandstone at Natural Bridges National Monument, Utah. In these formations the slab wall is commonly strongly overhung into a thin two-dimensional arch or alcove presumably backwasted along sheeting fractures. One puzzling aspect of these prominent indentations is a general paucity of mass wasting debris on the lower floor. Schumm and Chorley (1966) cited the ready breakup of the wasted debris as an explanation, but slab wall failures from relatively short cliffs yield abundant debris (Fig. 8.15), and the alcoves are a relatively protected environment. Another possible explanation is present-day inactivity of parting erosion and resulting slab failure due to aridity. The effects of climatic change on scarp morphology are discussed further below.

Hoodoos and Demoiselles

A few scarps are sculpted into highly intricate forms, such as occur at Bryce Canyon, Utah, and smaller forests of boulders on shale pedestals called hoodoos (Fig. 8.23). These forms occur where the caprock unit is discontinuous but massive and the underlying shales

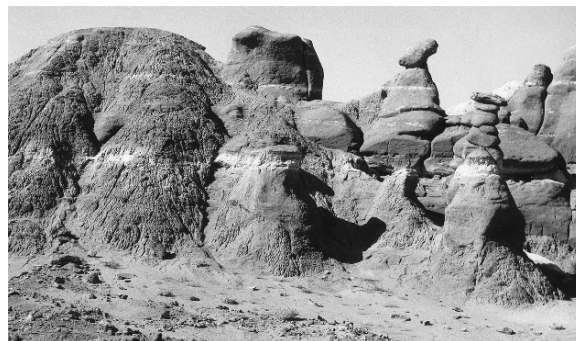


Fig. 8.23 Hoodoos in Entrada Sandstone near Hanksville, Utah. Hoodoos are capped by resistant concretions, whereas nearby poorly cemented sandy shales exhibit regolith-mantled badland slopes. Hoodoos are often supported by thin pedestals of shale (photo A. Howard)

are readily eroded where exposed to rain but easily protected from weathering by slight overhangs. Many scarps in the south-western United States are composed of cliffs that extend well below the caprock unit into the underlying shales (Fig. 8.24). These cliffs are remarkably stable, having persisted and grown vertically (downward) throughout the Holocene (and late Pleistocene?), showing the efficacy of overhangs as small as a few tens of centimetres in restricting surface weathering of shales in a desert environment. Typically the caprock units are concretions or discontinuous interbeds of cemented sandstones or limestones. They are embedded in shales or shaly, poorly cemented sandstones that typically erode into badlands where relief is sufficient and caprocks are absent. Exposure of the concretion or resistant interbed greatly reduces

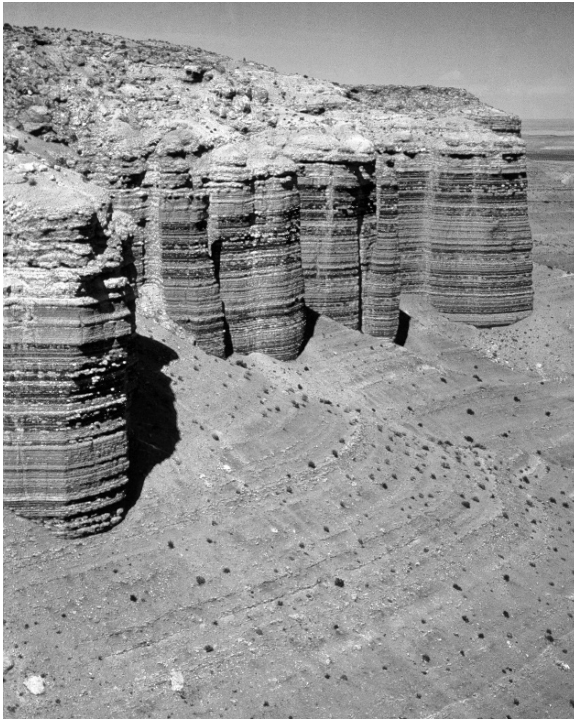


Fig. 8.24 Typical view of Morrison–Summerville escarpment at Sandstone Point off Utah Highway 95 near Hanksville, Utah, showing the cliff face extending into the weakly resistant Summerville Formation, the badland rampart in the same formation abutting the escarpment face, and the absence of debris on the rampart. Note that the rampart is shallowly underlain by bedrock and is not a colluvial apron (photo A. Howard)

erosion rates while the surrounding shale badlands continue to erode, producing a dichotomy in the shales between the vertical slopes protected by the caprock and the subjacent badlands. Development of stucco-like coatings of clay and lime by vertical drainage on the shale cliffs may be a contributing factor in protecting the shale from erosion (Lindquist 1979). Rates of vertical erosion in the shale badlands may be 40 times the rate of lateral retreat of the vertical shale cliffs (Lindquist 1979).

Role of Sapping Processes in Scarp Erosion and Morphology

Various geomorphologists have suggested that rock weathering and erosion at zones of groundwater discharge have contributed to the backwasting of scarps and valleys in sandstones exposed on the

Colorado Plateau (Gregory 1917, Bryan 1928, Ahnert 1960, Campbell 1973, Laity and Malin 1985). General discussions of groundwater sapping and its landforms have been provided in Higgins (1984), Howard et al. (1988), Baker et al. (1990), and Higgins and Osterkamp (1990). Laity and Malin (1985, p. 203) defined sapping as ‘the process leading to the undermining and collapse of valley head and side walls by weakening or removal of basal support as a result of enhanced weathering and erosion by concentrated fluid flow at a site of seepage’. Higgins (1984) distinguished between ‘spring sapping’ caused by concentrated water discharge and ‘seepage erosion’ resulting from diffuse discharge at lithologic contacts or other lithologic boundaries. This discussion addresses both the role of seepage groundwater in scarp erosion and the development of deeply incised valleys in sandstone by spring sapping.

Such definitions of sapping and seepage erosion are complicated by marginal and transitional situations. Scarp erosion processes that are clearly not sapping erosion include plunge-pool undermining and rock weathering by moisture delivered to the scarp face by precipitation, condensation, or absorption of water vapour. However, other circumstances are not as clear-cut. For example, water penetrating into tensional and exfoliation joints close to cliff faces may cause rockfalls as a result of freezing or water pressure but would probably not be classified as sapping by most geomorphologists. Similarly, corrasional erosion of shale beneath sandstone by water penetrating along wide fractures (termed subterranean wash by Ahnert 1960) is similar to piping, but probably should not be included as a process of groundwater sapping. On the other hand, rockfall caused by weathering of shales beneath a sandstone scarp in which water is delivered by flow along joints within the sandstone is more likely to be considered to be sapping, even in the absence of obvious water discharge along the scarp face. Weathering processes resulting from intergranular flow within sandstone would generally be considered sapping.

Groundwater flow plays an uncertain role in weathering of the shales and weakly cemented layers whose erosion causes scarp retreat in overlying sandstones. Oberlander (1977, 1989) mentioned spring sapping of shale partings as a possible process of scarp retreat in segmented scarps. Schumm and Chorley (1966), while providing experiments and observations on surface

weathering of caprock units, essentially avoided the issue of processes of caprock undermining. Koons (1955) was similarly vague. Ahnert (1960) clearly felt that sapping processes are of general importance in scarp retreat in sandstone-shale scarps of the Southwest USA, but he provided little evidence. Laity and Malin (1985) suggested that disruption of surface exposures by salt crystal growth where seepage emerges and sloughing of thin sheets of the bedrock are the major processes of sapping erosion in massive sandstones, and that sapping is usually concentrated in thin zones above less permeable boundaries within or below the sandstone. This backwasting and undermining of the overlying sandstone then occasions development of slab failure and, locally, alcove development associated with development of exfoliation jointing as outlined by Bradley (1963). Laity and Malin primarily discussed spring sapping processes occurring at canyon headwalls, and it is uncertain the degree to which they felt sapping or seepage erosion occurs more generally on sandstone scarps. Observations reported below suggest an important role of shallow groundwater circulation in scarp retreat in sandstone–shale sequences of the Southwest USA under present climates.

Many scarps are indented by V-shaped re-entrants or canyons excavated by erosion along streams passing over the escarpment. Such erosion is generally considered to result from corrasion of the bed (e.g. Howard and Kerby 1983) or from plunge-pool action. However, examination of canyon heads in sandstone suggests that sapping processes may play a role in channel erosion, at least for washes with drainage areas less than several square kilometres. Washes passing over scarps in sandstone commonly occupy only a fraction of the total scarp width. In addition, the scarps are commonly overhung when developed in massive sandstone and plunge pools are rare and small below the waterfalls. Even steep streams on thin sandstone beds sandwiched between shale layers exhibit overhangs considerably wider than the stream bed and show little development of plunge pools. Much of the erosion of the sandstone beds in such cases may occur due to sapping, with a very localized water source from the overlying stream. Additionally, small washes developed on slickrock slopes above scarps are commonly interrupted by solution pits and the washes exhibit flutes and furrows that suggest that solutional removal of calcite cement is more important than mechanical corrasion in bed erosion.

Weathering and erosion of sandstone by the effects of crystal growth occur at a variety of scales in sheltered locations. Steep slopes and scarps in sandstone are frequently interrupted by rounded depressions, often overlapping, which intersect sharply with the general slope (Fig. 8.25). Such alveolar weathering, or tafoni, occurs not only in sandstone, but also in granites, tuff, and other massive rocks (Mustoe 1982, 1983). Both accelerated erosion in the hollows and case hardening of the exposed portions of the slope (Conca and Rossman 1982) may contribute to development of tafoni. Salt accumulations are often quite apparent in the cavernous hollows, and the backwasting results in spalling of sheets of weathered rock up to a few centimetres in thickness. Mustoe (1983) noted high soluble cation contents in the spall detritus in tafoni and the presence of the mineral gypsum. Laity (1983) and Laity and Malin (1985) found calcite deposition on spalling walls. The mechanisms by which such mineral deposition may contribute to the spalling include pressures exerted by crystal growth, thermal expansion and contraction of the crystal filled rock, and expansion and contraction due to hydration of deposited minerals (Cooke and Smalley 1968). Freeze–thaw disruption on the moist seepage faces may also contribute, and spalling may be aided by the weight of accumulated winter ice (Laity and Malin 1985).

A surface protected from surface runoff is a necessary condition for tafoni and alcove development. On steep sandstone scarps surface runoff commonly flows as sheets down the scarp, held by surface tension on slightly overhung slopes. Such runoff paths are commonly accentuated by rock varnish. Where

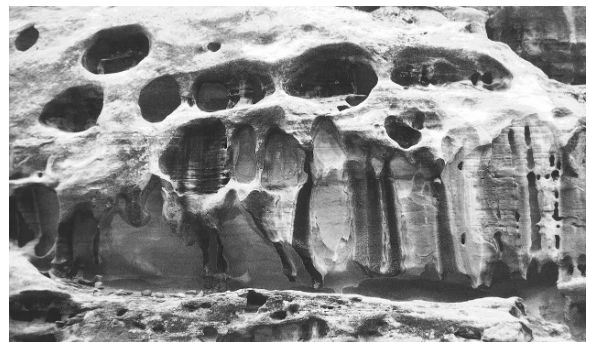


Fig. 8.25 Alveolar weathering, or tafoni, developed in Navajo Sandstone at Capital Reef National Monument, along Utah Highway 24. Note uneroded ribs where surface wash occurs. Image height about 10 m (photo A. Howard)

such runoff paths cross zones of tafoni development, backwasting is inhibited, and the tafoni are separated by columns that often resemble flowstone columns in caves (Fig. 8.25). Surface runoff might inhibit salt fretting simply by solution and removal of salts brought to the surface by evaporating groundwater or more actively by case hardening of the exposed surface by deposition of clays or calcite (Conca and Rossman 1982).

Two intergrading types of sapping landforms develop on massive sandstones. The more exotic form is the development of tafoni on steep scarps and on large talus blocks. Such tafoni may literally riddle certain steep slopes (Fig. 8.25), with the tafoni concentrated along certain beds that are either more susceptible to the salt fretting or receive greater ground water discharge. Talus blocks generally develop tafoni on their lower, overhung portions. The concentration of tafoni development at the base of such blocks may be due both to the protection from surface wash as well as upward wicking of salts from underlying soils or shales, a process that contributes to weathering of the bases of tombstones (Mustoe 1983, Hamilton 1984). Oberlander (1977) pointed out that tafoni develop most strongly on scarps initially steepened by basal undermining but presently no longer backwasting because alluviation or aeolian deposition covers and protects the basal backwasting face. Thus generalized tafoni development on a scarp indicates relative inactivity of backwasting by surface attack or basal undercutting.

On the other hand, large alcoves are common in massive sandstones and are often actively retreating as a result of sapping erosion (Fig. 8.26). However, direct sapping usually is localized to zones less than 2 m thick along permeability discontinuities where the discharge of groundwater is concentrated (Fig. 8.27), although at major valley heads the seepage zone may be 20–25 m thick. These sapping zones generally backwaste by processes similar to those of tafoni, and locally tafoni are superimposed upon the sapping face. In addition to salt fretting, backwasting by groundwater discharge can also occur by cement dissolution and by weathering of shale beneath or interbedded in the sandstone. The retreat of the active zone of sapping undermines the sandstone above, with the result that occasional rockfalls occur (Fig. 8.15). In massive sandstone the undermining occasions the development of exfoliation sheeting fractures, resulting in large arches or alcoves, with the deepest parts of the alcoves presumably corresponding to the most rapid sapping at-



Fig. 8.26 Theatre-headed valley in Navajo Sandstone near Utah Highway 95 along North Wash (southern Henry Mountains area). Note cottonwood trees and *dark* figure in wash at bottom centre of photo for scale. The active seep is the *dark* band at the base of the alcove. Note the evidence of offloading fracturing. The stream passing over the *top* of the alcove is clearly inadequate in size to have created the alcove as a plunge pool. Water draining down the face of the alcove from the lip of the falls has created the dark streaks. However, delivery of water along the alcove walls occurs only locally, and is insufficient to account for the backwasting of the alcove (photo A. Howard)

tack (Fig. 8.27). In well-jointed sandstones, such as the Wingate Sandstone, arches and overhanging cliffs are less common, and the role of scarp retreat by sapping processes is not as obvious but may be just as important.



Fig. 8.27 Seepage face at the theater head alcove shown in Fig. 9.26. Zone of seepage is 1–2 m thick. Floor of alcove is highly weathered sandstone debris supporting a vegetative cover. The seepage face is encrusted with ferns, mosses, algae, and other phreatophytes. Salt efflorescences occur above moist zone. Note person for scale at arrow (photo A. Howard)

The major aquiclude for the Navajo Sandstone is the underlying Kayenta Formation, and the major seeps develop at this discontinuity (Figs. 8.26 and 8.27). However, thin shales and limestone interbeds (interdunal deposits) create minor aquicludes within the Navajo (Laity 1988, Kochel and Riley 1988), leading to frequent development of multiple levels of seeps and associated alcoves at valley headwalls.

A distinction may be made between wet sapping with a damp rock face and an effluent discharge and dry sapping, where the rock face is generally incrustated with mineral salts (Laity and Malin 1985). In general, tafoni are associated with dry sapping because of their localized development, whereas large alcoves are generally associated with a more regional groundwater flow and exhibit faces that are at least seasonally wet.

Neither active seepage nor deposition of mineral crusts on protected sandstone walls are necessarily correlated with rapid weathering and backcutting of the sandstone walls. For example, the Weeping Wall at Zion National Park, Utah, is an impressive seep emerging from the Navajo Sandstone, but the associated alcove and canyon are small. Many other examples of fairly high discharge rates but only minor, or non-existent, alcoves can be found throughout the Colorado Plateau. Too rapid a seepage may in fact discourage deposition of salts. Although rapid seepage can also cause backwasting by dissolution of calcite or gypsum cement, this would only occur if the groundwater were under-saturated. Similarly, many examples of thick mineral incrustations at seeps lacking evidence of backwasting can be found on sandstones throughout the Southwest USA. Several factors control whether minerals deposited by evaporating seepage are deposited intergranularly within the rock (encouraging exfoliation and granular disintegration) or at the rock surface (with little resulting sapping), including the type and concentration of salts, the average and variance of water discharge to the surface, the distribution of pore sizes and their interconnectivity, the presence and size of fractures, the temperature regime at the rock face, local humidity and winds, and the frequency of occurrence of wetting of the rock face by rain or submergence (if along a stream or river). The interaction of these factors is poorly understood. Discrepancies between size of alcove or sapping valley and the magnitude of the seep can also result from differences in the length of time that sapping has been active. Despite these cautions, it

is reasonable to expect within a specific physiographic, structural, and stratigraphic setting that the degree of alcove development and degree of headward erosion of sapping valleys would correlate with the discharge of the seeps involved (Laity and Malin 1985).

In summary, most evidence suggests that sapping processes are common in sandstones of the Colorado Plateau and the processes result in the widespread development of tafoni and alcoves. The major caution is the evidence cited by Oberlander suggesting that groundwater sapping is unimportant in erosion of effective partings at thin shale layers in the Entrada and other sandstones. One of several possibilities may account for this difference in interpretation. The effective shale partings may, in fact, be so readily weathered by atmospheric moisture due to high salt content or other factors that seepage is not required for such partings as it is for cavernous weathering and alcove development elsewhere. The Entrada Sandstone exhibits considerable variation in lithology and cementation both vertically and areally, so that in the Arches National Park the Slick Rock Member may be relatively impermeable. However, well-developed arches and tafoni with evidence of active salt fretting are found a few tens of kilometres south in this same member. Another possibility is that backwasting along effective partings is not very active under present climatic conditions, but was more active during the late Pleistocene. The evidence for relatively inactive scarp retreat (and, by extension, seepage erosion) under present climatic conditions is discussed below.

Climatic Change and Scarp Morphology

Most recent discussions of scarp morphology and associated weathering and erosional processes have implied that the morphological elements can be explained by presently active processes (e.g. Koons 1955, Schumm and Chorley 1966, Oberlander 1977). However, other authors have maintained that escarpments in the Colorado Plateau region are undergoing considerably less caprock erosion under the present climate than during the late Pleistocene (Reiche 1937, Ahnert 1960, Howard 1970).

Given our poor understanding of present scarp processes and little historical control, it is not surprising that different conclusions have been drawn about

present activity and the presence or the lack of a direct relationship between present processes and present scarp form. For example, Schumm and Chorley (1964, 1966) cited numerous examples of historical rockfalls as evidence for present activity of scarps. However, a simple analysis suggests just the opposite. Inferred long-term rates of scarp retreat on the Colorado Plateau average about 3 m per 1000 years (Schmidt 1989b). If rockfalls occur randomly through time and characteristically involve a block of caprock 1 m thick and 10 m in horizontal dimension, then one should expect about three rockfalls per year on 10 km of scarp front. If the characteristic rockfall is larger, say 3 m thick and 100 m long, then there should be a rockfall every ten years. Since there are tens to hundreds of thousands of kilometres of scarp front on the Colorado Plateau, occurrences of large rockfalls should be much more frequent than appears to be the case, possibly by as much as an order of magnitude.

The escarpment of Emery Sandstone near Caineville, Utah, is a case in point. Extensive old rockfall deposits on North and South Caineville Mesas, dissected to a depth of as much as 50 m at their lower and lateral margins, were interpreted by Howard (1970) to be Bull Lake equivalent in age (Illinoian?) because some of the debris blankets interfinger with pediment deposits of Bull Lake age, whereas the remainder extend no further downslope than the level of the former pediment surface (Fig. 8.28). The old debris blankets have been dissected by a similar amount and project concordantly laterally along the



Fig. 8.28 Pleistocene-age (Early Wisconsinan?) rockfall on escarpment of Emery Sandstone over Mancos Shale on South Caineville Mesa near Hanksville, Utah. Note heavy dissection of lower end of rockfall blanket. Note remnant of pediment of same age in front of escarpment on left side of photo (photo A. Howard)

escarpment. Much of the scarp rampart is now devoid of extensive debris, so that extensive badlands have developed on the underlying Mancos Shale. In places the cliff capped by the Emery Sandstone extends downward by as much as 70 m into the underlying shale (Fig. 8.14). Koons (1955) demonstrated that development of cliff faces in the non-resistant subcaprock unit is a normal occurrence in the cycle of rockfall, debris blanket erosion, and erosion of the subcaprock unit which triggers the next rockfall. However, for reasons outlined above, extensive cliff development in the subcaprock unit may also indicate stagnation of caprock mass wasting while erosion of the subcaprock unit continues on the scarp rampart.

Schmidt (1988, 1989a) and Gerson and Grossman (1987) noted similar remnant debris blankets related to climatic change and, in particular, talus flatirons and associated pediment flatirons (Fig. 8.29). Schmidt (1988, 1989a) termed climatically controlled flatiron development cyclical flatirons as contrasted with the non-cyclic flatirons discussed previously, and also cited the presence of associated dissected pediments (pediment flatirons) as a distinguishing characteristic of cyclic flatirons.

An outstanding example of a presently stagnant scarp morphology is the escarpment capped by the lower units of the Morrison Formation overlying the thin bedded sand and shale beds of the Summerville Formation near Hanksville, Utah. Much of the escarpment is capped by a thin gypsum layer which is stable enough to support vertical cliffs up to 30 m high in the Summerville Formation (Fig. 8.24). The gypsum cap has been eroded into small crenellations mimicked in the underlying Summerville, indicating that the caprock protects the vertical face from weathering and erosional processes. The rampart generally consists of debris-free badlands likewise carved in the Summerville Formation, which meet the vertical face at an abrupt angle. The rampart is bordered downslope by dissected pluvial (Bull Lake?) pediments paved with agate derived from the caprock, giving evidence that abundant debris was transported from the escarpment during that time. Where a caprock of gypsum is present, even a thick sequence of overlying sandstone sheds little debris on to the rampart. But where the gypsum unit is discontinuous or missing (Fig. 8.30), the overlying sandstone wastes by undermining and rockfall. However, during the Bull Lake pluvial climate the gypsum caprock evidently succumbed more

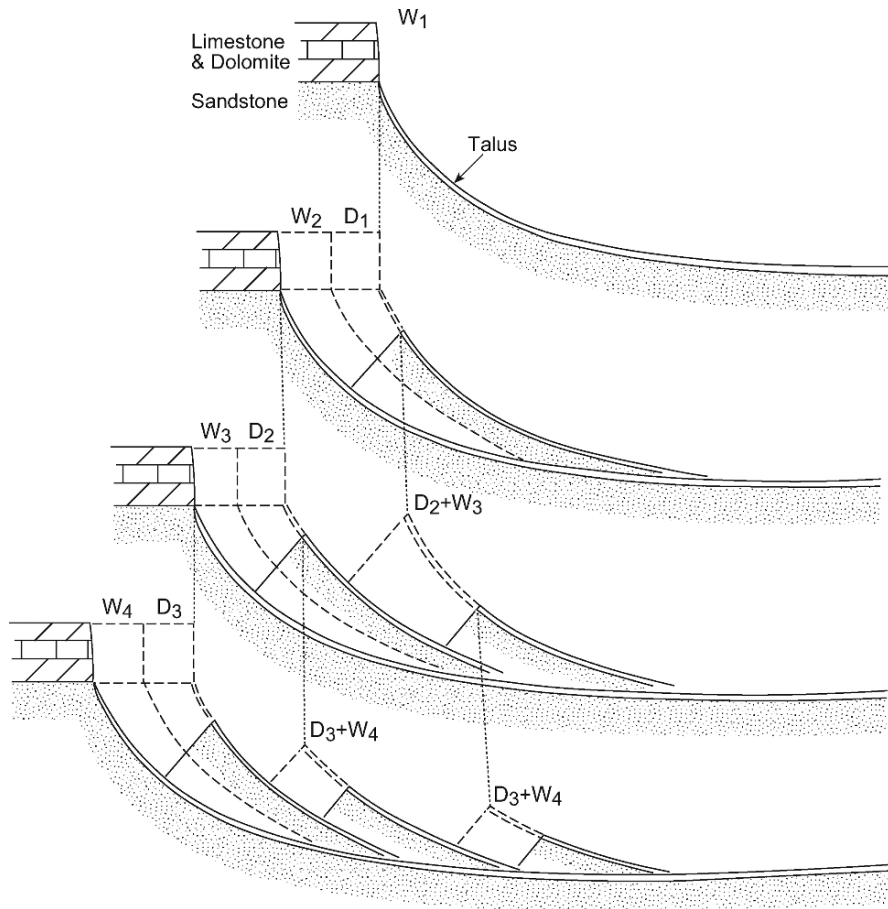


Fig. 8.29 Talus flatirons developed by alternating wet (W) and dry (D) climates. Talus aprons and associated pediments develop during enhanced weathering and transport of wet periods and are

dissected during dry epochs (from Gerson and Grossman, 1987, Fig. 17–8)



Fig. 8.30 View of northwest face of Goatwater Point, along Utah Highway 95 near Hanksville, Utah, showing rapid backwasting of the sandstones capping the Morrison–Summerville escarpment in the absence of the gypsum caprock. Vertical cliffs

in the Summerville Formation occur only where the gypsum unit is present, such as on the left edge and centre of the picture (photo A. Howard)

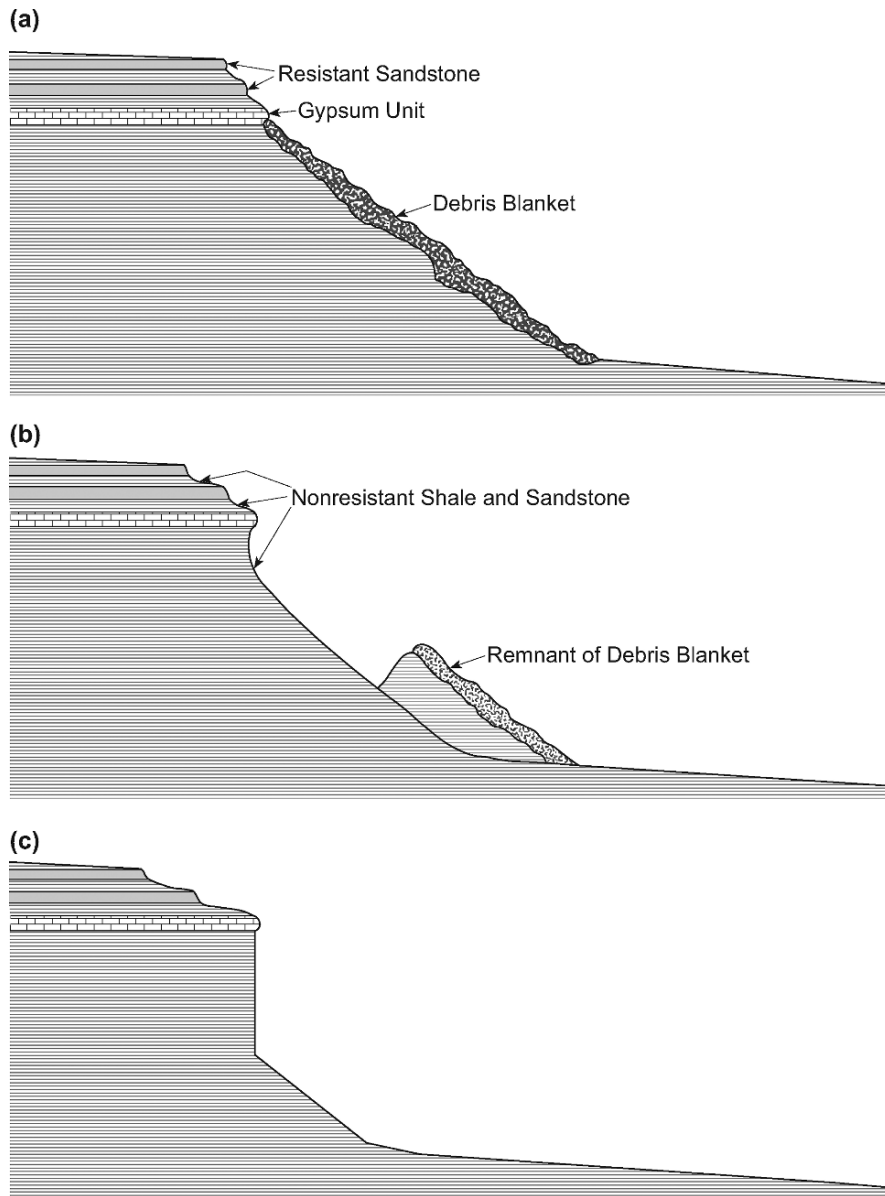


Fig. 8.31 Cross-sections through the Morrison–Summerville escarpment showing presumed Pleistocene conditions compared with the present-day profile. During the Pleistocene pluvials, backwasting of the caprock units by rockfall and block-by-block undermining was probably rapid, and the debris-covered rampart extended up to the caprock without tall vertical cliffs in the underlying Summerville Formation (cross-section **a**). Between rockfalls, weathering and erosion eventually removed the sandstone and gypsum debris, and a cliff would begin to form in the

Summerville Formation (**b**). Because of the greater supply of water and lower temperatures, vertical cliffs higher than those in profile (**b**) were probably unstable, for another rockfall would return the escarpment to the conditions at (**a**). However, under the drier and warmer Holocene conditions, the gypsum caprock has become very resistant, and backwasting of the escarpment has essentially halted, with the result that the escarpment rampart is rapidly eroding away, leaving behind high cliffs in the Summerville Formation (**c**) (from Howard 1970, Fig. 81)

readily to rockfall and occasional slumping (Fig. 8.31). The scarps illustrated in photos 3 and 22 of Schumm and Chorley (1966) are probably further examples of such presently inactive scarps.

In canyon areas, a cliff in a massive sandstone often is underlain by a nearly bare rock slope with a gradient of about 30–40°. No obvious lithologic break separates these slope elements. In some cases, such as at Canyon de Chelly, Arizona, examples can be found where such bare rock slopes yield laterally to rock slopes of similar gradient but mantled with rockfall debris (Fig. 8.32). One explanation for such features is that they are Richter slopes, forming at the characteristic friction angle of rockfall debris under conditions where removal of rockfall debris (by weathering or erosion at the scarp base) just balances production at the superjacent cliff face (Richter 1901, Bakker and Le Heux 1952, Cotton and Wilson 1971, Selby 1982d, pp. 204–8, Scheidegger 1991, p. 132; also equivalent to the case of Schumm and Chorley's (1966) unity weathering ratio). However, at least in the Colorado Plateau examples, it is difficult to imagine how such a scattered debris cover protects the underlying rock from weathering and erosion. One possible explanation is that the moving talus has erosive capability which planes the rock down to the friction angle of the moving talus. Erosion by rock and ice avalanches has been used to explain spur-and-gully development on steep arctic and alpine mountain slopes (Rapp 1960a,b, Howard 1989, 1990). However, in this case erosion tends to become localized in chutes.

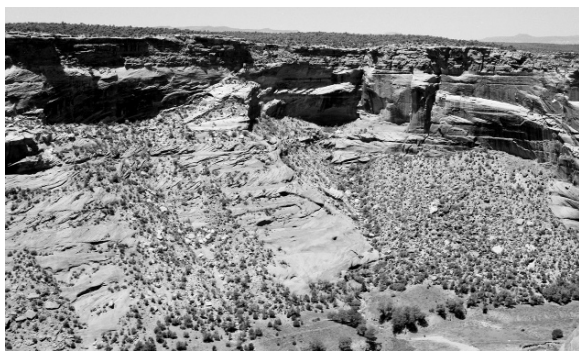


Fig. 8.32 Portion of a canyon wall in a portion of Canyon De Chelly, Arizona, showing prominent vertical cliffs in upper parts of the De Chelly Sandstone with slickrock slope of 30–40° inclination below the cliffs and also development in the De Chelly Sandstone. Note partial mantling of lower slopes with talus and the suggestion that other slickrock slopes were formerly mantled (photo A. Howard)

A more direct explanation is that the bare rock slopes were formerly mantled with rock debris produced by cliff retreat and that the debris mantle protected the underlying rock from weathering and erosion, forming threshold slopes (Carson 1971, Carson and Petley 1970). The cover of debris on such slopes is now obviously inadequate to protect the rock from weathering and erosion with the implication that scarp backwasting is presently largely inactive, permitting the weathering and stripping of the former debris mantle. Somewhat similar forms can be found in the areas of segmented scarps in Entrada Sandstone at Arches National Monument (the area studied by Oberlander 1977) (Fig. 8.21). Oberlander interpreted the low-gradient slickrock slopes to be a normal feature of erosion of segmented scarps (Fig. 8.22), but it is possible that some of the slickrock slopes below vertical cliffs may have formerly been mantled by a debris blanket under a different climate (a scattering of debris can be seen on many of these segmented slopes).

Ahnert's (1960) historical interpretation of scarp morphology contrasts sharply with the 'dynamic equilibrium' interpretations of Schumm and Chorley (1966) and Oberlander (1977). Ahnert suggested that scarp retreat occurs by erosion of shales and other weakly resistant layers by sapping-related processes that have only been active during pluvial episodes characterized by abundant, gentle precipitation. By contrast, he felt that slickrock slopes are eroded largely by sheetwash, which has been active during the sparse but intense rainfall of interpluvials. Ahnert felt that the generally sharp breaks between slickrock slopes and subjacent scarps were evidence of the non-contemporaneous origin of the two features, whereas Oberlander (1977) and Howard and Kochel (1988) felt that such sharp breaks can result from contemporaneous slickrock erosion and scarp retreat. Ahnert's assertion that slickrock slopes are eroded primarily by sheetwash misses the importance of the initial weathering that must occur on these weathering-limited slopes. This weathering, which makes debris capable of slopewash transport, is probably not optimal under present climatic conditions of brief, intense rainstorms, but is favoured by winter precipitation and freeze–thaw.

Despite these objections to Ahnert's interpretations, the examples presented previously suggest that there have been changes throughout the Pleistocene and

Holocene in the relative rates of the major processes producing cuesta landforms: scarp undermining by sapping, undermining, or surface-directed weathering; weathering and erosion of rockfall debris; weathering and erosion of slickrock slopes; erosion of interbedded shale layers; and, locally, slumping and landsliding. Changes in the relative importance of these processes have produced landforms that in many cases can only be understood by knowledge of the temporal process changes (i.e. relict landforms). These examples of remnant effects of past climates add a complicating element in unravelling the geomorphic history of cuesta landforms and interpretation of scarp morphologies which would greatly benefit from application of modern techniques of geomorphic surface dating.

The greater rainfall during the Bull Lake pluvial maximum would suggest greater sapping activity then relative to now, but that same moisture supply would also contribute to surface-directed weathering, particularly freeze–thaw weathering. Also, more abundant groundwater does not necessarily produce more sapping in sandstones because if seepage is sufficient and humidity high enough to permit runoff of the seepage, little salt accumulation will occur. However, other groundwater-related weathering processes would be enhanced by greater available moisture, such as hydration and leaching of shale interbeds. Thus during the pluvial epochs we might expect more debris production by freeze–thaw, less alveolar weathering, but greater back-wasting at the contact between sandstones and underlying shales and along shale interbeds. The present paucity of rockfall debris, rather than indicating nearly complete breakup and rapid weathering of the debris, as suggested by Schumm and Chorley (1966), may simply indicate relative inactivity of many scarps in present climates.

Another prominent feature of the theatre headwalls at seeps is the paucity of rockfall debris. This commonly contrasts with abundant rockfall debris along the sides of the valley downstream from the seep (Fig. 8.26). Laity (1988) suggested that both initial pulverization of rockfall debris and rapid weathering in the moist environment of the headwall account for the paucity of debris. However, recent rockfalls in alcoves generate abundant debris, and the rockfall debris along valley walls also suggests that the backwasting of scarp walls produces a considerable volume of coarse debris that must be further weathered before it is transported from the rampart and backwasting can

continue. The paucity of rockfall debris at headwater scarps is therefore interpreted as further evidence of greater sapping activity during past pluvial periods and relative inactivity during the present climatic regime.

In the western desert of Egypt, particularly the Gilf Kebir, scarp retreat via groundwater sapping during wet episodes of the Pleistocene and present erosional inactivity has been inferred by Peel (1941), Maxwell (1982), and Higgins and Osterkamp (1990).

Quantitative Models of Scarp Profile Evolution

A variety of quantitative models of scarp profiles has been developed over the years. Most notable are the models of scarp backwasting and talus accumulation that predict the evolutionary slope history and the sub-talus bedrock profile (Lehmann 1933, Bakker and Le Heux 1952, Gerber and Scheidegger 1973, and others: see summary in Scheidegger 1991, pp. 130–4). These have been developed for the case of a horizontal, fixed base level with no provision for weathering and removal of talus except through a constant ratio of rock eroded to talus produced.

Other slope evolution models permit resistant layers, but some express resistance solely in terms of resistance to weathering and do not allow for the effects of mass movement and surface wash (e.g. Scheidegger 1991, pp. 144–7; Aronsson and Linde 1982), or they include creep-like mass movement but not the effects of surface wash, rockfall, and talus weathering (Ahnert 1976, Pollack 1969).

Developing of realistic scarp profile models will prove challenging owing to the range of processes and materials that must be considered. The first component must be the accounting of weathering and failure of the caprock, addressing rates of weathering, effects of slope steepness and profile, and predicting size and frequency of rockfall, undermining, or slumps, possibly including the role of groundwater sapping. A second component should follow the distribution, comminution during emplacement, weathering, and erosion of caprock debris, as well as the protective influence of the debris on underlying bedrock. The third component should address erosion of the subcaprock unit where exposed. Finally, initial and boundary conditions must be addressed, including stratigraphy, dip, base level

changes, and possible process variations through time. A final problem with profile evolution on ramparts with rockfall talus is that production and erosion of the talus is inherently three-dimensional (e.g. the development of flatirons).

Evolution of Scarp Planform

The planimetric form of canyons and escarpments is the most obvious signature of the erosional processes involved in scarp retreat in layered rocks. The processes of scarp erosion create characteristic shapes of planform features such as re-entrants, projections, and inset canyons. For example, some scarps have rounded projections, or headlands, whereas others are sharply terminated. Similarly, some scarps are inset with deep, narrow canyons, while others have shallow, broadly rounded reentrants. The scarp form is determined by the spatial distribution of the processes discussed earlier as well as the lithologic and structural influences such as rock thickness and dip. Several generalities about scarp planform have been noted for many years.

- (a) In areas of strong structural dip (more than a few degrees) scarps tend to be oriented updip with a linear to gently curving planform that closely follows the structural strike, locally interrupted by through-flowing streams. At least two factors contribute to the linearity of scarps with steep dip. Drainage divides tend to be at or very close to the scarp crest, so that little fluvial or sapping erosion occurs. In addition, escarpment height is a strong function of crest location, so that portions of the scarp that lag in erosion rapidly become higher than surrounding locations, tending to enhance average rates of retreat. Thus the remarkable linearity of scarps in steeply dipping rocks is *prima facie* evidence for the sensitive dependency of scarp backwasting rates on scarp relief.
- (b) Where the structural dip is slight, scarp planforms are highly textured, with deep embayments and canyons as well as headlands and detached mesas. Scarps may face either updip or downdip, with downdip portions being more deeply embayed because drainage areas on top of such scarps are generally larger.
- (c) Resistant rock units generally become first breached by stream erosion, forming scarps,

where structural elevations are high (e.g. at the top of anticlines) and/or where gradients are steepest (e.g. on dip-slope stripped plains on the resistant unit). Both factors are combined in the common scenario of breaching of caprocks just below the crest of monoclines.

Since caprock erosion occurs dominantly by scarp retreat, the scarp planform reflects areal variations in rate of scarp retreat. One factor controlling scarp retreat rates is the erosional resistance of the caprock and/or subcaprock units. Nicholas and Dixon (1986) examined the role of variations in inherent caprock resistance (e.g. compressive strength, slake durability, cementation) and caprock fabric (fracture orientation and spacing) in controlling relative rates of caprock erosion. They found that embayments in the Organ Rock Formation escarpment, Utah, have a higher density of fractures at the base of the caprock than occurs beneath headlands. However, caprock resistance is nearly equal on headlands and escarpments. Accordingly, they suggest that areal variations in caprock fabric are a dominant control on scarp planform (Fig. 8.33).

The other factor controlling scarp form is areal variations in process rates. Erosion of the scarp face by rockfall, slumping, and undermining of the caprock, weathering of the caprock debris, and erosion of subcaprock units will be termed scarp backwasting. Deep re-entrant canyons are clearly created by rapid erosion along a linear zone, generally by fluvial erosion or groundwater sapping. Structure often plays an indirect

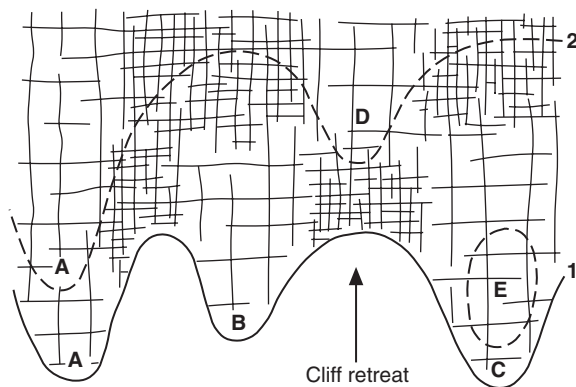


Fig. 8.33 Plan view of scarp retreat, showing variable rates of backwasting governed by differences in fracture density. Particularly unfractured masses may persist as isolated buttes, as at A, B, D and E (from Nicholas and Dixon 1986, Fig. 8)

role by channelling surface or subsurface water along fractures. The role of areal variations in intensity of these three classes of processes and in rock resistance in creating scarp planforms is discussed below based upon both field evidence and theoretical modelling. The discussion will concentrate on the development of Schumm and Chorley's (1964) compound scarps consisting of one scarp-forming unit sandwiched between easily eroded rocks.

Scarp Backwasting

Natural scarps commonly exhibit a planform characterized by sharply pointed or cusped projections (headlands, spurs) and broadly concave re-entrants (embay-

ments). The role of essentially uniform rates of scarp retreat in creating pointed headlands and broad embayments was first discussed by Dutton (1882, pp. 258–9) and Davis (1901, pp. 178–80), who noted the nearly uniform spacing between successively lower scarps on the walls of the Grand Canyon. Attack of an escarpment by areally uniform backwasting gradually makes embayments more shallow and the planform of the scarp face close to linear (Fig. 8.34). Similar observations were made by Howard (1970) and Schipull (1980). Lange (1959) systematically studied the consequences of uniform erosional attack (uniform decrescence) on two- and three-dimensional surfaces, showing that uniform erosion acting on any arbitrary, rough surface produces rounded re-entrants and cusped spurs. He also showed that scarp retreat in the

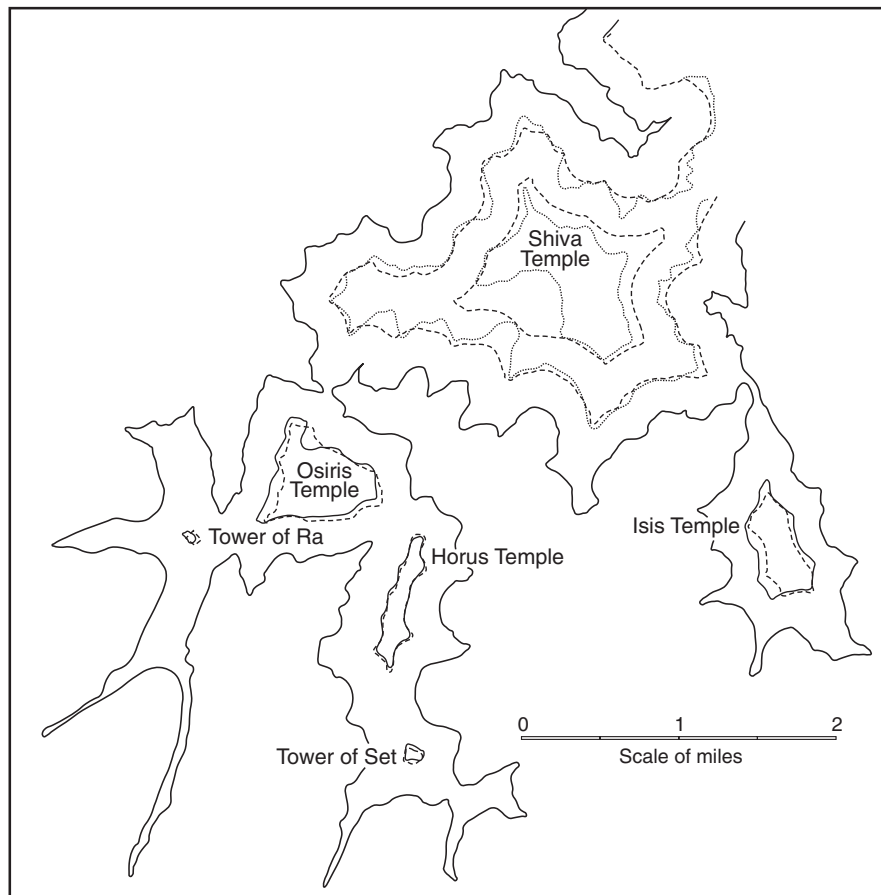


Fig. 8.34 Process of uniform erosional attack illustrated on the Redwall, Coconino, and Kaibab escarpments within the Grand Canyon. The rim outline of both the Coconino and Kaibab escarpments (*dotted lines*) may be approximately duplicated by uniform erosional attack of the Redwall escarpment rim (*dashed*

lines). In a few cases the hollows, or re-entrants, are deeper than would be produced by uniform backwasting of the next lower escarpment; sapping backwasting may be involved (from Lange 1959, Fig. 7.17)

Grand Canyon is uniform to a first approximation, with some tendency towards more rapid erosion of embayments (Fig. 8.34). Furthermore, uniform decrease and uniform addition of material (accrescence) are not reversible (Lange 1959), so that the position of an escarpment rim during the past cannot be uniquely inferred from its present configuration, even if subject to uniform retreat in the past, especially in the case of former spurs that first eroded into outliers and then wasted away.

A simple numerical model of scarp backwasting (Howard 1988, 1995) lumps local variations in process rates and rock resistance into a single 'erodibility', which is assumed to vary randomly from location to location. The erodibility is assumed to be a self-similar fractal with variations at both large and small scales. The spatial variability of erodibility is characterized by three parameters, an average value, a variance, and a rate of change of variance with scale. The rate of lateral backwasting is assumed to be linearly dependent upon the erodibility. Figure 8.35 shows a plan view of successive scarp positions starting from a square 'mesa'. All simulations discussed here assume a scarp such as is shown in Fig. 8.10, with a single caprock over shale and a superjacent stripped plain. As expected, scarps produced by this type of

backwasting are characterized by sharply terminating projections and broad re-entrants (as in the Grand Canyon, Fig. 8.34), with re-entrants produced by erosion through more erodible rocks.

The projections of some natural scarps are rounded rather than sharply pointed. This suggests that the lateral erosion rates on the projections are enhanced relative to straight or concave portions of the scarp. Several mechanisms may account for this. The rate of erosion of many scarps is limited by the rate that debris shed from the cliff face can be weathered and removed from the underlying rampart on the less resistant rocks. Debris from convex portions of scarps (projections) is spread over a larger area of rampart, so that weathering rates may be enhanced. Additionally, headlands commonly stand in higher relief than re-entrants, leading to longer and/or steeper ramparts and enhanced erosion rates. Finally, some scarp caprock may be eroded by undermining due to outward creep or weathering of the rocks beneath the caprock. Both processes are enhanced at convex scarp projections. Figure 8.36 shows a model scarp in which backwasting rates are enhanced by scarp convexity and restricted by concavity, and Fig. 8.37 shows a similar natural scarp.

By contrast, some natural scarps exhibit narrow, talon-like headlands that are unlikely to have resulted

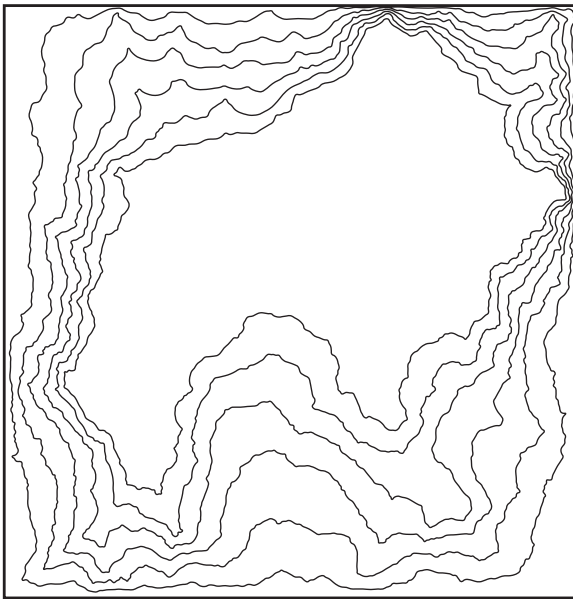


Fig. 8.35 Stages of simulated scarp retreat by uniform erosional attack of caprock with spatially variable resistance. The simulation starts from an assumed square mesa and the lines show successive stages of retreat of the edge of the scarp. Note that the lines are not contours

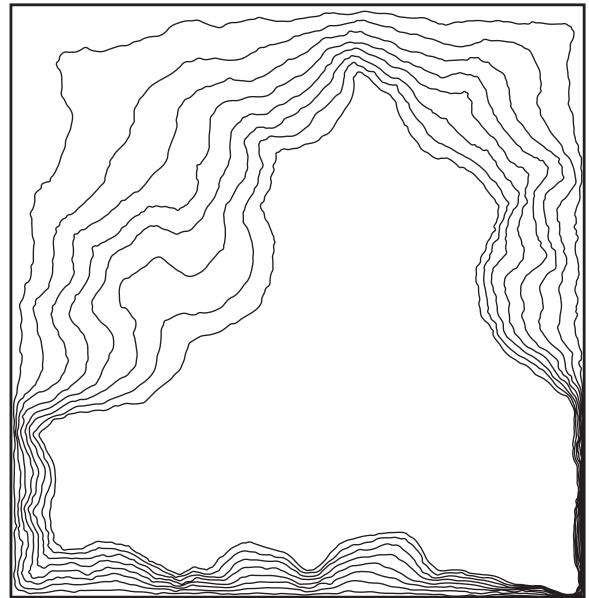


Fig. 8.36 Plan view of simulated scarp evolution by uniform erosional attack with an assumed dependency of retreat rate on planform convexity. Note rounding of scarp headlands

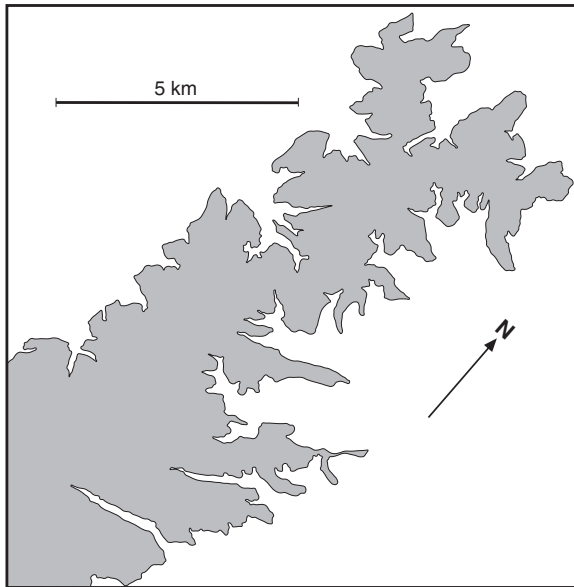


Fig. 8.37 Plan view of a natural scarp with rounded headlands in sandstone underlain by shale in north-eastern New Mexico (1:100000 Capulin Mountain quadrangle). Note that fluvial incision has created numerous sharply terminated embayments. The uneroded portion of the scarp is patterned

from pure uniform retreat (Figs. 8.34 and 8.38). Occasional thin headlands may result from narrowing of divides between canyons by uniform back-wasting, but where such sharp cusps are frequent it is likely that they result from slowed erosion of narrow projections. Impeding of erosion on such headlands may be due to a lack of uplands contributing surface or subsurface drainage to the scarp face, thus inhibiting chemical and physical weathering of the cliff face or its undermining. Scarps that have been broken up into numerous small buttes, such as occurs at Monument Valley, Arizona, may be further examples of inhibition of scarp retreat where upland areas are small. An alternative explanation may be that these headlands occur in locations with minimal caprock jointing, in the manner suggested by Nicholas and Dixon (1986).

Almost all escarpments in gently dipping rocks exhibit deep re-entrants and, as erosion progresses, break up into isolated mesas and buttes. This indicates erosion concentrated along generally linear zones of either structural weakness, fluvial erosion, or sapping (often acting in combination). One indicator of the role of either or both fluvial erosion and groundwater sapping is the asymmetry of scarps in gently tilted rocks. The planform of segments where the scarp faces up-



Fig. 8.38 Natural scarp with sharply pointed projections and sharply pointed, gradually narrowing box canyon embayments due to stream incision (scarp of the Redwall limestone in Grand Canyon National Park: Grand Canyon 1:100000 quadrangle). The uneroded portion of the scarp is patterned

dip is generally similar to that expected by uniform decrease, since little drainage passes over the scarp, but segments facing down-dip ('back scarps' of Ahnert 1960) are deeply indented as a result of fluvial or sapping erosion (Fig. 8.39) (Ahnert 1960, Laity and Malin 1985).

Nicholas and Dixon (1986) emphasized the role of variable density of fracturing of incompetent and caprock layers in controlling the rate of scarp retreat and the development of re-entrants, projections, and isolated buttes. Their evidence suggests that variable density of fracturing is important on certain scarps, but it probably controls primarily the small-scale platform features. Variable density of fracturing cannot account for the remarkable asymmetry of gently dipping scarps (Fig. 8.39) and the general association of embayments with the drainage basins on the overlying stripped surface of the caprock.

Fluvial Downcutting

One process creating deep re-entrants is downcutting by streams which originate on the top of the escarpment and pass over its front. Such caprock erosion is localized to the width of the stream, which is generally very small compared with overall scarp dimensions, so that the heads of fluvially eroded canyons are quite narrow, but increase gradually in width downstream by scarp backwasting, as noted by Ahnert (1960) and Laity and Malin (1985). Examples of such gradually

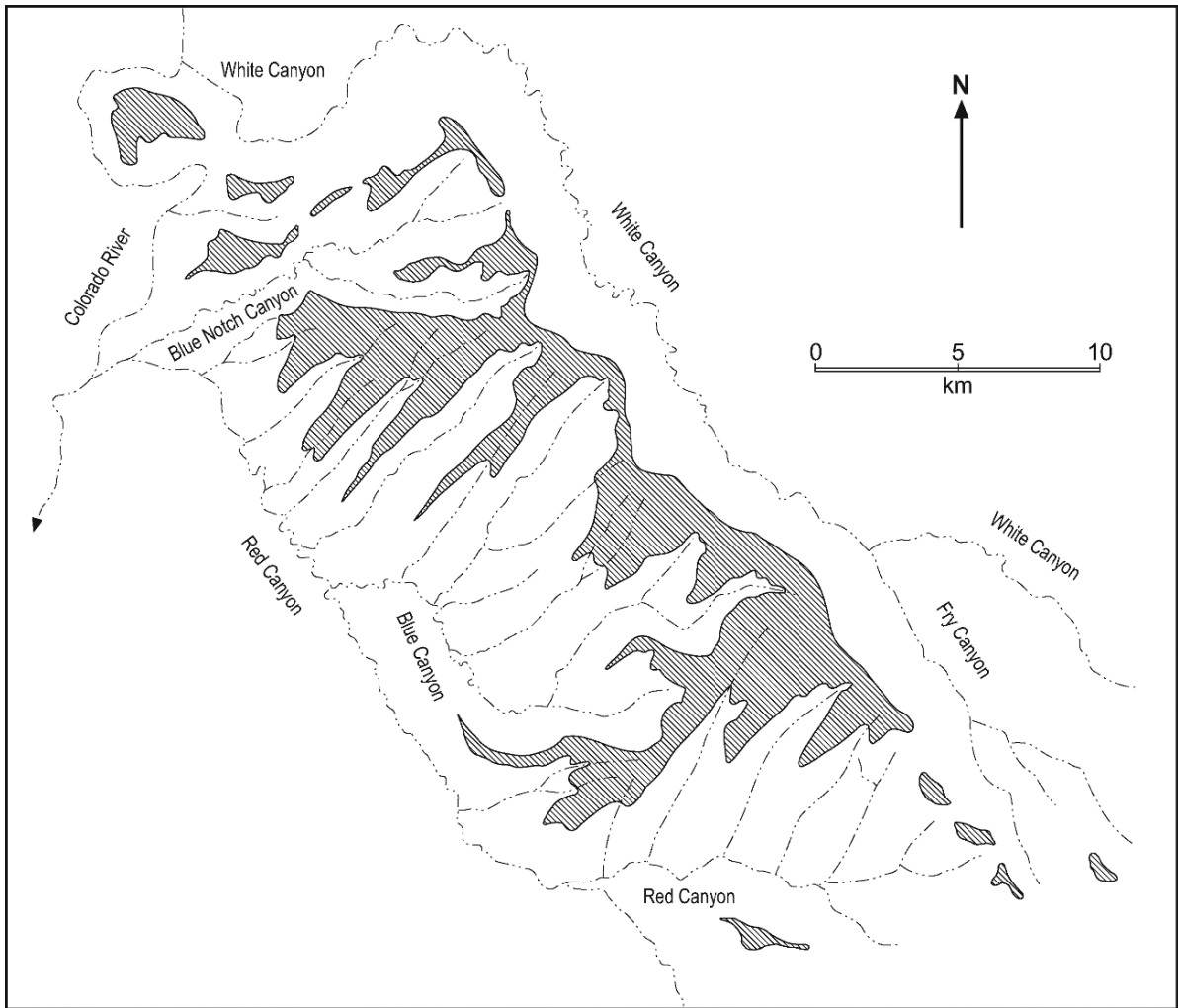


Fig. 8.39 Plan view of an escarpment of gently dipping Wingate Sandstone showing smooth, nearly straight planform on up-dip side and deeply embayed down-dip side. Strata dip about 1.5° due

west. Mesa is located to south-west of Utah Highway 95 between the Colorado River and Natural Bridges National Monument

narrowing, deep re-entrants are common on the Colorado Plateau (Figs. 8.37 and 8.38).

Scarp erosion by combined fluvial erosion and scarp backwasting can be numerically modelled by superimposing a stream system on to a simulated scarp and allowing it to downcut through time. If a resistant layer is present, the stream develops a profile characterized by a low gradient section on top of the resistant layer and a very steep section (rapids or falls in natural streams) which rapidly cuts headward. In the model shown here, erosion is assumed to be proportional to the product of the bed erodibility and the average channel bed shear stress (Howard

and Kerby 1983). Between streams, scarps are assumed to erode by the backwasting process described above. Figure 8.40 shows a simulated scarp eroded by both fluvial downcutting and scarp backwasting, and Fig. 8.41 is similar but with convexity-enhanced scarp backwasting. Figures 8.37 and 8.38 are natural scarps similar to Figs. 8.41 and 8.40, respectively. Scarp morphology is similar to that produced by scarp backwasting, with the addition of several deep re-entrants characterized by a gradually narrowing width and projections that are sharply pointed in the linear case and rounded in the convexity-enhanced case.

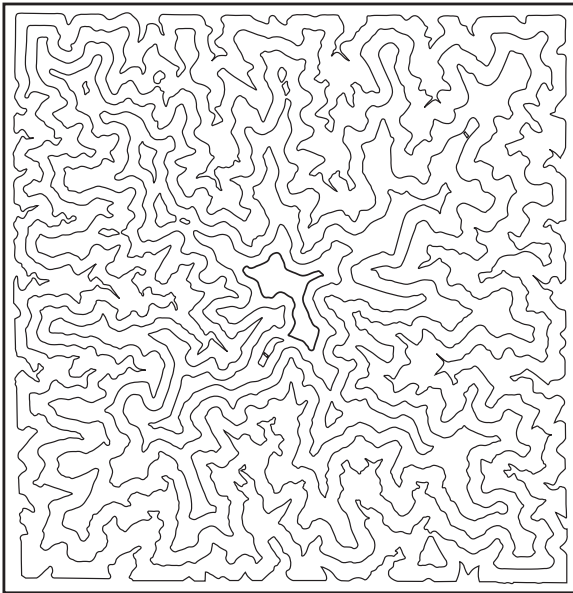


Fig. 8.40 Plan view of successive stages of simulated erosion of an initially square mesa by combined action of fluvial downcutting and uniform scarp retreat

All of the fluvial models combine scarp backwasting with downcutting and scarp retreat due to fluvial erosion, with valleys gradually enlarging downstream due to longer duration since fluvial backwasting passed

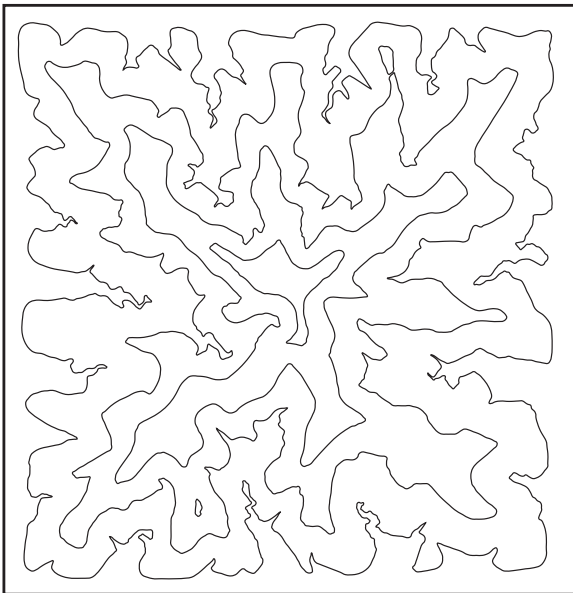


Fig. 8.41 Plan view of successive stages of simulated erosion of an initially square mesa by combined action of fluvial downcutting and uniform erosion with a curvature rate dependence

through that location. The angle subtended by the valley walls ψ is related to the ratio of the scarp backwasting rate R_b to the stream headcutting rate R_s by the expression

$$R_b/R_s = \sin \psi/2. \quad (8.6)$$

The rate of downstream increase of valley width has been used by Schmidt (1980, 1989b) to estimate scarp recession rates.

Groundwater Sapping

Groundwater sapping is an important process in scarp retreat throughout the Colorado Plateau (Ahnert 1960) and probably contributes to the scarp backwasting process discussed above. However, the deep, narrow, bluntly terminated canyon networks of the type discussed by Laity and Malin (1985) and Howard and Kochel (1988) are the scarp form most clearly dominated by sapping processes.

Ahnert (1960) and Laity and Malin (1985) suggested that while fluvial erosion produces V-shaped canyon heads that widen consistently downstream, sapping produces U-shaped, theatre-headed canyons of relatively constant width downstream (the terms U- and V-shaped refer here to valley planform, not valley cross-section). Other planform features diagnostic of sapping-dominated canyon extension and widening include theatre-shaped heads of first-order tributaries with active seeps (Figs. 8.26 and 8.42), relatively constant valley width from source to outlet (Figs. 8.42 and 8.43), high and steep valley sidewalls, pervasive structure control, and frequent hanging valleys. The most direct evidence of sapping processes are the numerous alcoves, both in valley heads and along sidewalls, and the many springs on the valley walls and bottoms. Although many valley headwalls occur as deeply undercut alcoves, some terminate in V or half-U shapes with obvious extension along major fractures (Fig. 8.43). Although sapping action is fairly evident where alcoves occur at valley heads, sapping may also occur in fracture-controlled headwalls. Evidence for erosion by streams passing over the valley headwalls is slight, and plunge pools are not obvious. The valley extension along fracture traces suggests control by groundwater flowing along the fractures, and a preponderance of the major valley heads are

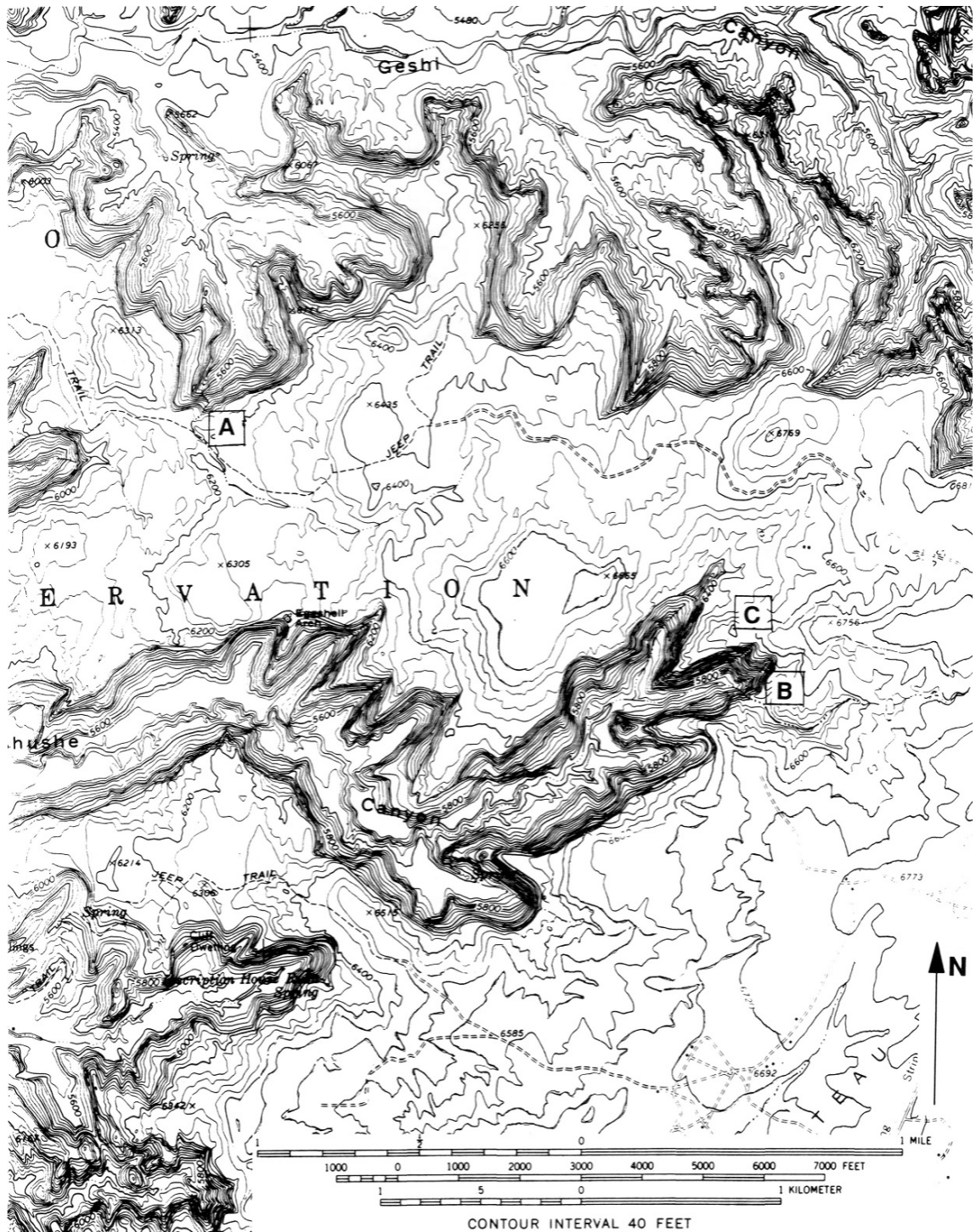


Fig. 8.42 Topographic map of box canyons eroded into Navajo Sandstone in the Inscription House area of the Navajo Indian reservation, Arizona (part of the Inscription House Ruin 1:24000 quadrangle). Note that the backslope drainage at B does not enter the canyon at its head and that the canyon headwalls at C extend

nearly to the backslope divide with little contributing drainage area. The extensive upslope drainage enters the large alcove at A at a minor niche. Also note the rounded, theatre-like canyon headwalls, most of which have active seeps

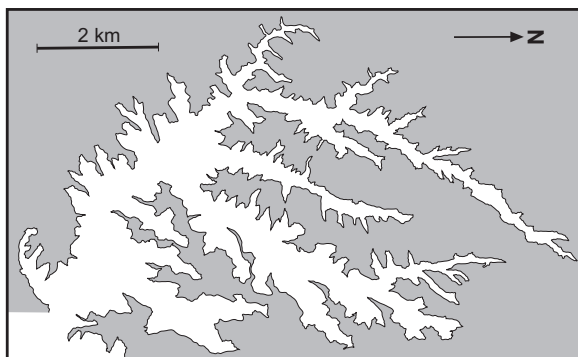


Fig. 8.43 Planform of cuesta in Navajo Sandstone at Betatakin National Monument, Arizona, showing rounded canyon headwalls, weak branching, and nearly uniform canyon width. Re-entrants locally show strong control of orientation by fractures (Kayenta 1:100000 quadrangle). The uneroded portion of the scarp is patterned

oriented updip, which is consistent with sapping by a regional downdip groundwater flow. One striking feature of these canyon networks is frequent discrepancy between the extent of headward canyon growth and the relative upland area contributing drainage to the canyon head, including situations where the upland drainage enters the side rather than the end of the canyon (Fig. 8.42). Such circumstances suggest that groundwater flow rather than surface runoff controls headward extension of the canyons. Another indication of sapping control is the extension of canyons right up to major topographic divides (Fig. 8.42), sometimes causing the divides to be displaced. In contrast to surface runoff, groundwater flow can locally cross topographic divides. Divide migration can also occur in surface runoff drainage basins (see discussion of badlands), but is associated with the presence of slopes that are well graded from stream to divide. In the case of headward migration of canyons, streams above the scarp are perched on the upper slickrock slopes, so that base-level control is very indirect.

The pattern of valley development and the relative contribution of sapping processes versus fluvial erosion is influenced by many structural, stratigraphic, and physiographic features. Infiltration may be restricted where overlying aquicludes were not stripped from the sandstone (Laity and Malin 1985, Laity 1988). The permeability of the sandstone is influenced not only by primary minerals but by diagenetic cements and overgrowths. These secondary minerals vary considerably from layer to layer and location to

location (Laity 1983, 1988). Kochel and Riley (1988) and Laity (1988) discussed the important role that thin aquicludes (generally interdune deposits) and bedding strikes and dips have in controlling groundwater flow through the sandstone. Primary and off-loading (sheeting) fractures are important avenues of groundwater migration (Laity and Malin 1985, Laity 1988, Kochel and Riley 1988). On the other hand, dense primary fracturing, such as occurs in the Wingate Sandstone, restricts development of alcoves and probably diminishes weathering by cement solution and salt fretting (Laity 1988), although sapping processes may still be important but lack the spectacular alcoves commonly developed in the Navajo Sandstone. Relatively flat uplands on sandstones should encourage infiltration, whereas steep slickrock slopes probably lose much precipitation to runoff. Weathering pits and thin covers of windblown sand may also encourage infiltration.

A simulation model of scarp backwasting by groundwater sapping assumes a planar caprock aquifer (that may be tilted) with uniform areal recharge and a self-similar areal variation in permeability. The rate of sapping backwasting E is assumed to depend upon the amount by which the linear discharge rate at the scarp face q exceeds a critical discharge q_c : $E = k(q - q_c)^a$, where k and a are coefficients. Scarp erosion is simulated by solving for the groundwater flow, eroding the scarp by a small amount, with repeated iterations. Figure 8.44 shows a simulated scarp with $q_c = 0$ and $a = 1$. Figure 8.45 shows a case with $q_c = 0$ and $a = 2$. Figure 8.46 shows a case with $q_c = 0$ and $a = 2$ and superimposed scarp backwasting. Figure 8.43 is a natural scarp in Navajo Sandstone formed largely by sapping. Valleys developed by groundwater sapping tend to be linear or crudely dendritic, with nearly constant width and rounded rather than sharp terminations, in accord with the predictions of Ahnert (1960) and Laity and Malin (1985).

Deep re-entrants that are essentially constant in width with rounded ends imply either that almost all valley enlargement occurs at the headward end of the valley (see, for example, the simulations in Figs. 8.44 and 8.45) or that a period of rapid headward canyon growth (either by sapping or fluvial incision) is succeeded by a period of uniform scarp retreat. This latter case is exemplified by the simulations in Fig. 8.46 combining groundwater sapping and uniform backwasting in which rapid valley erosion by sapping

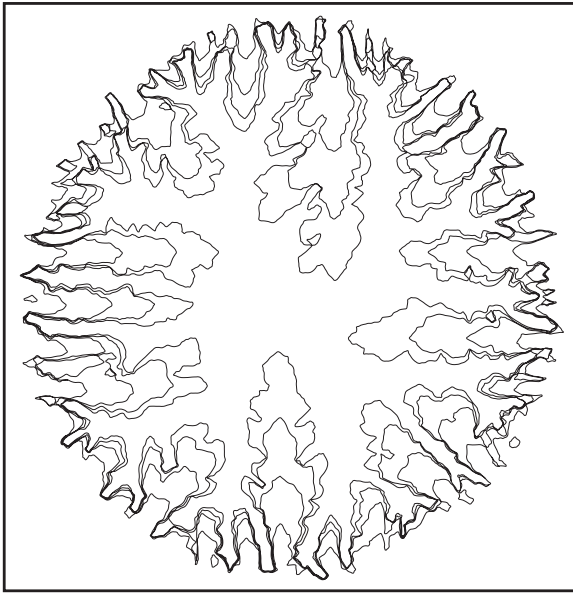


Fig. 8.44 Plan view of successive stages of simulated erosion of an initially round mesa by groundwater sapping. Caprock aquifer is assumed to undergo uniform recharge from above and water flows along the base of the caprock at the contact with underlying shale until it emerges at the edges of the mesa. Erosion rate is proportional to seepage discharge rate per unit width of scarp face

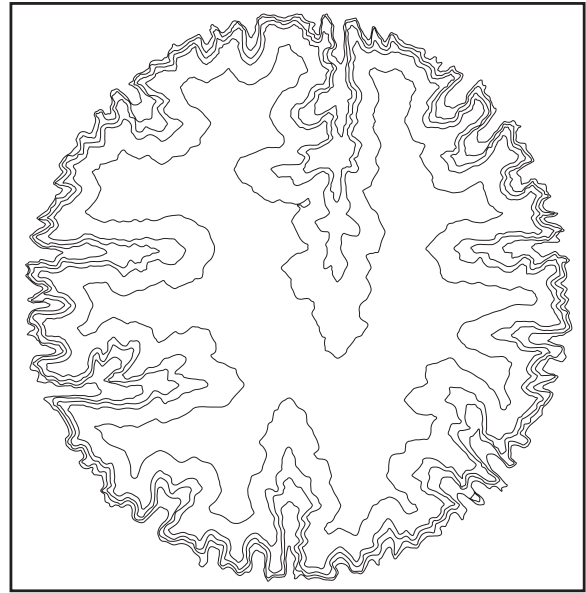


Fig. 8.46 Simulated sapping erosion of an initially round mesa as in Fig. 8.43, but with superimposed uniform scarp retreat

early in the simulation is succeeded by a dominance of uniform backwasting in later stages of erosion as upland areas contributing to groundwater sapping diminish in size.

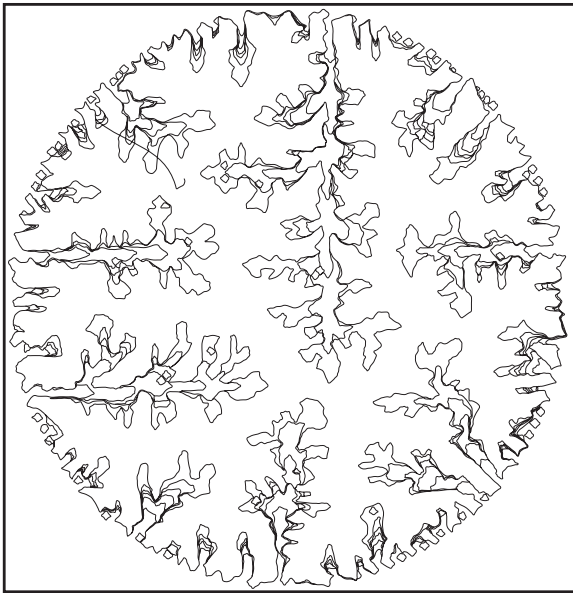


Fig. 8.45 Simulated sapping erosion of an initially round mesa as in Fig. 8.42, but with erosion rate proportional to the square of seepage discharge rate

Summary of Scarp Modelling

These simulations illustrate a common evolutionary sequence in scarp erosion. Early stages are characterized by a planform determined largely by the agency first exposing the scarp, generally fluvial downcutting. Thus scarp walls parallel the stream course with small re-entrants along tributaries draining over the scarp. Continuing tributary erosion by fluvial or sapping processes with modest rates of scarp backwasting between streams eventually creates a maximally crenulate scarp planform. As erosion continues and upland areas diminish, rates of stream or sapping backcutting diminish relative to more uniform scarp backwasting (which may accelerate as relief increases) so that the scarp planform becomes characterized by broad re-entrants, sharp cusps, and often isolated small buttes during the last stages before complete caprock removal. This is illustrated in Figs. 8.40 and 8.46. This sequence can occur sequentially on a single large

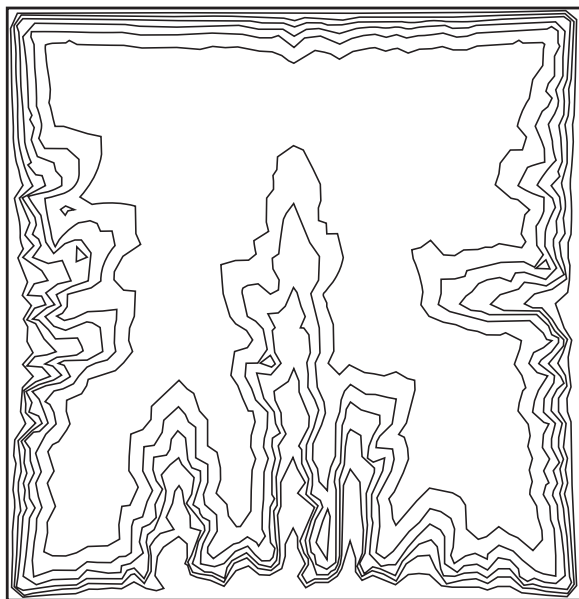


Fig. 8.47 Simulated sapping erosion of a slightly tilted square mesa by groundwater sapping. The caprock tilts to the bottom. Note the strongly asymmetrical development of re-entrant canyons

mesa, as takes place in the simulations, but it is also illustrated by spatial sequences during long-continued erosion of a given caprock from newly exposed scarps in structurally low areas to the last caprock remnants in structurally high areas.

Even slight structural dips create a strong asymmetry of rates of fluvial and sapping headcutting on down- and up-dip facing portions of the scarp, since drainage divides tend to be located close to the up-dip crest. General scarp backwasting is less affected by dip, so that up-dip scarp segments are nearly linear, whereas down-dip sections are highly crenulate. This is illustrated in the natural scarp in Fig. 8.39 and the simulated sapping erosion of a tilted scarp in Fig. 8.47.

The planform of escarpments gradually develops through scarp retreat of hundreds to thousands of metres. Thus it is unlikely that planforms reflect to any great extent Quaternary climatic oscillations. Longer-term climatic changes may have influenced scarp planforms, but such influences would probably be hard to detect.

Conclusions

Weathering and erosional processes on exposed bedrock are generally either very slow-acting or occur

as large and often rapid mass wasting events. As a result, process studies of the sort that have been conducted on badland and rock-mantled slopes are rare. Furthermore, the slow development of bedrock landforms often implies a strong influence of climate changes on slope form. Among the issues that need further study are (a) the relative roles of stress-release fracturing and contemporary gravitational, neotectonic, or weathering-induced stresses in generating broadly rounded bedrock exposures, (b) weathering processes involved in groundwater sapping, (c) controls on rates of scarp backwasting and scarp height and steepness, and (d) processes of talus weathering and removal on scarp ramparts. We anticipate that the study of landform development on bedrock will benefit greatly from application of modern techniques for dating of geomorphic surfaces, including use of rock varnish (Chapters 7 and 21) and thermoluminescence, fission-track, and radiogenic or cosmogenic isotopes.

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