Chapter 9

Rock-Mantled Slopes

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Introduction

Desert hillslopes below the angle of repose are dominated by the weathering characteristics of the underlying lithology, and specifically by the rate of production of fine material compared to the rate of removal. The previous chapter considered hillslopes underlain by massive rocks, or those in layered rocks dominated by outcropping resistant layers. These lithologies are weathering limited, and give rise to hillslopes where a surficial layer of weathered material is thin or absent. On more readily weathered lithologies a more-or-less continuous layer of debris is found. This layer of debris is subject to pedogenic processes. This chapter deals with such hillslopes.

The size distribution of particles forming this layer is a function of the composition and weathering characteristics of the underlying lithology. Consequently, rock-mantled slopes lie along a continuum. At one end of the continuum they grade into rock slopes. At the other end of the continuum, where the underlying lithology is, itself, dominated by fine-grained materials, they may grade into badlands (Chapter 10). On both massive rocks and on badlands vegetation in sparse or absent. In contrast, the layer of weathered material on rock-mantled slopes provides a substrate for vegetation to grow. Consequently, these hillslopes often have a vegetation cover and the processes acting upon them are affected by this vegetation (see Chapter 3).

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Slope Form and Adjustment

Although rock-mantled slopes may exist at any gradient below the angle of repose, a distinction is often drawn between those with gradients in excess of 10° and those below this gradient. This distinction arises because of the sharp boundary between upland areas and piedmonts – the piedmont junction – that is characteristic of desert environments and which usually occurs at about this angle. Above this gradient, rock-mantled slopes have been termed boulder-controlled slopes (Bryan 1922), debris-covered slopes (Melton 1965) or debris slopes (Abrahams et al. 1985). However, many characteristics of these slopes persist through the piedmont junction, so here we retain the more general term rock-mantled slopes and consider hillslopes both steeper and gentler than 10°.

Typically the profiles of rock-mantled slopes are convex-rectilinear–concave, though either the rectilinear or concave elements may be missing. Generally the upper convexity is narrow, and the profile is dominated by either the rectilinear or concave element. The rectilinear element tends to dominate on slopes affected by stream undercutting (Strahler 1950), and the concave element on slopes unaffected by this process. However, even in the latter circumstances, a rectilinear element may be present and occupy a significant proportion of the profile, especially where the slope is long or steep.

The coarse debris mantling these slopes is often embedded within and/or resting upon a matrix of fines, particularly toward the base of slopes or where gradients are gentle. These fines are produced by the chemical and physical breakdown of the coarse debris. As the weathering particles become finer, they are preferentially transported downslope by hydraulic processes,

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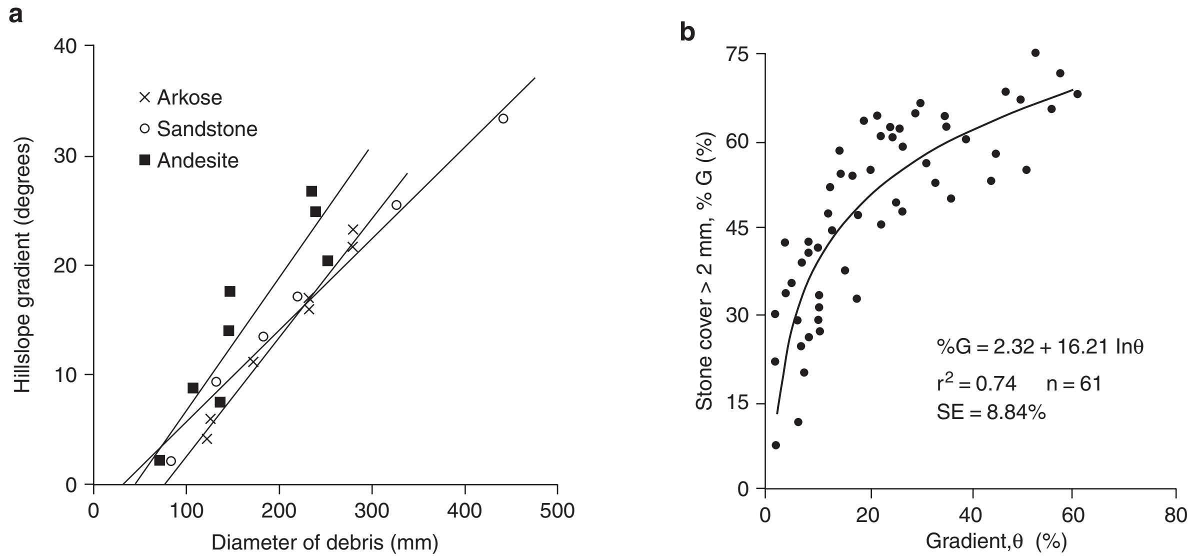

Fig. 9.1 Graphs of hillslope gradient against various measures of particle size. (**a**) Graph of hillslope gradient against the mean size of the ten largest debris particles for three debris slopes on three rock types in southern Arizona (after Akagi 1980). (**b**) Graph of stone cover against hillslope gradient for 12 debris slopes in Walnut Gulch Experimental Watershed, Arizona

so that usually the proportion of fines increases and the proportion of coarse debris decreases in this direction. Inasmuch as gradient also tends to decrease down-slope, positive correlations between gradient and various measures of particle size often obtain, particularly on weak to moderately resistant rocks (e.g., Cooke and Reeves 1972, Kirkby and Kirkby 1974, Akagi 1980, Abrahams et al. 1985, Simanton et al. 1994) (Fig. 9.1).

A more detailed picture of the variation in mean particle (fines plus debris) size with gradient down a slope profile is presented in Fig. 9.2. This profile is located in the Mojave Desert, California, and is underlain by closely jointed latitic porphyry (Fig. 9.3). Beginning

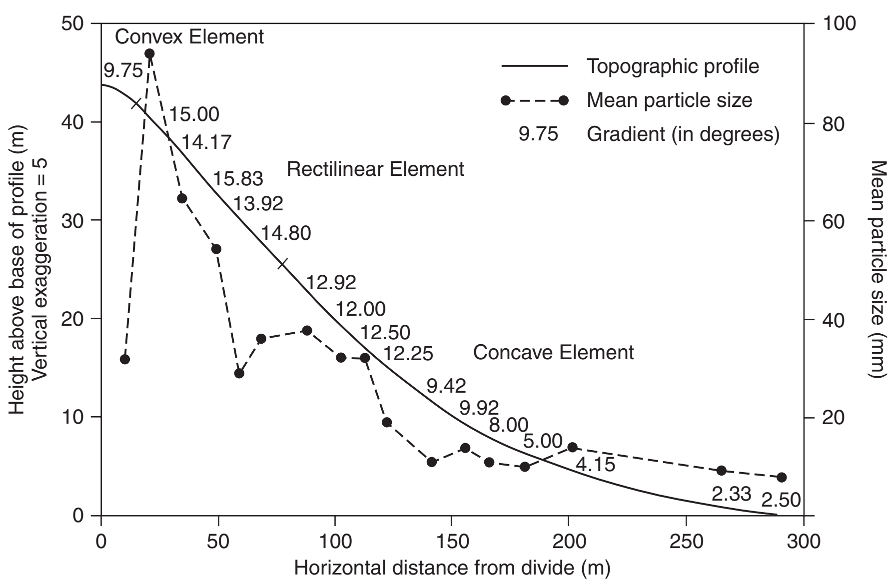

Fig. 9.2 Debris slope profile showing the downslope variation in mean particle size (sample size 100) and gradient (measured length 5 m). The debris slope is depicted in Fig. 9.3

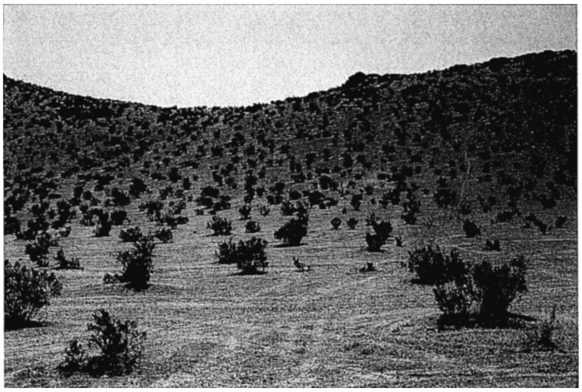

Fig. 9.3 Photograph of well-adjusted debris slope underlain by latitic porphyry in Turtle Valley, Mojave Desert, California

at the divide, mean particle size increases with gradient down the upper convexity. Note, however, that the particles are much larger than at comparable gradients on the basal concavity. This is because the weathering mantle is thinner and bedrock outcrops are more common on the convexity. Downslope from the convexity is a substantial rectilinear element. Mean particle size is at a maximum at the top of this element and decreases down the element. The decrease continues down the long concave element. This downslope pattern of change in particle size is representative of many, if not most, rock-mantled slopes without basal streams, including those that are both much steeper and much gentler than this example, and it appears to be primarily due to the selective transport of fine sediment by hydraulic processes.

That hydraulic processes play a dominant role in removing sediment from and fashioning many rock-mantled slopes is suggested by a study of the relation between gradient S and mean particle size $\overline{D}$ for slopes underlain by weak to moderately resistant rocks in the Mojave Desert, California. In this study, Abrahams et al. (1985) found that plan-planar slopes on different rocks have S–$\overline{D}$ relations with similar slope coefficients but different intercepts (Fig. 9.4). However, on a given rock the slope coefficient varies with slope planform, being greater for plan-concave slopes than for plan-convex ones (Fig. 9.5).

To explain their findings Abrahams et al. assumed that sediment transport by hydraulic processes can be characterized by an equation of the form

$$G \propto X^m S^n / \overline{D}^p \qquad (9.1)$$

where G is sediment transport rate, X is horizontal distance from the divide, and m, n, and p are positive

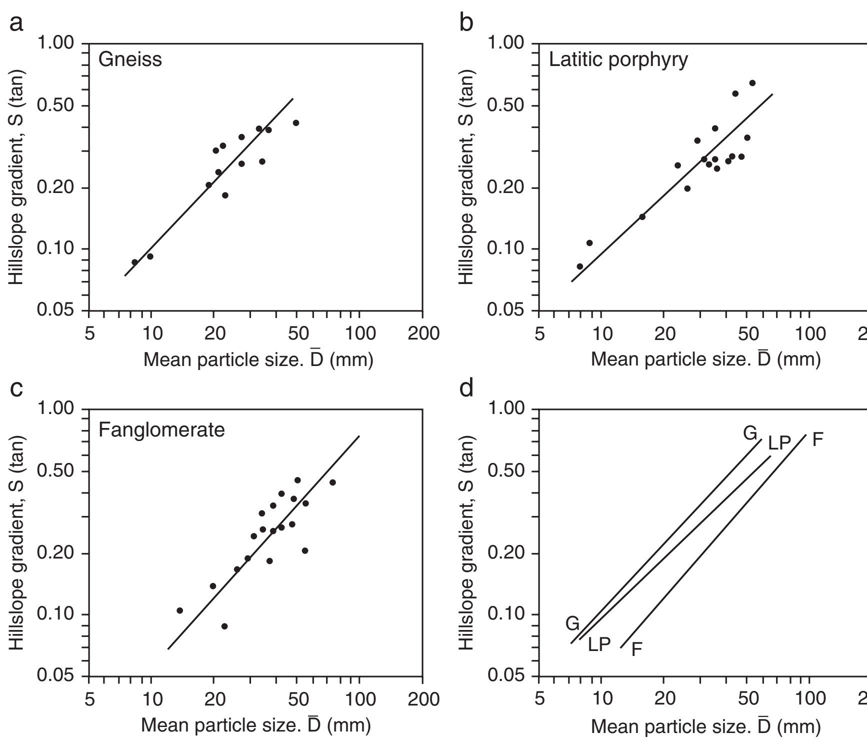

Fig. 9.4 Graphs of hillslope gradient against mean particle size for plan-planar debris slopes underlain by (**a**) gneiss, (**b**) latitic porphyry, and (**c**) fanglomerate. The fitted lines in (**a**), (**b**), and (**c**) are reproduced in (**d**) for comparative purposes (after Abrahams et al. 1985)

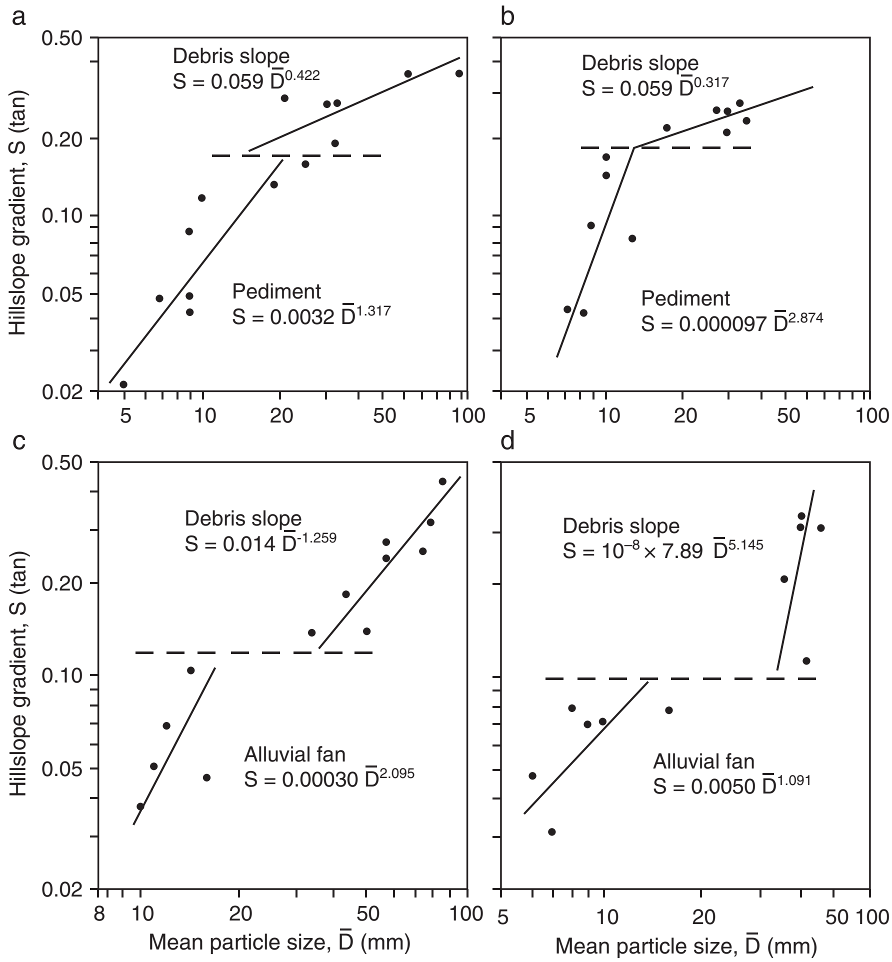

Fig. 9.5 Graphs of hillslope gradient against mean particle size for two (**a**, **b**) plan-convex debris slopes and their basal pediments and two (**c**, **d**) plan-concave slopes and their basal alluvial fans (after Abrahams 1987). Note that the S–$\overline{D}$ relations are much steeper for the plan-concave slopes than for the plan-convex ones

coefficients. Now if slopes are formed by and adjusted to hydraulic processes, Equation (9.1) may be manipulated to ascertain how the S–$\overline{D}$ relation varies with slope planform. Because $\overline{D} \propto X^q$, where $q < 0$,

$$G \propto \overline{D}^{[(m/n)-p]} S^n \tag{9.2}$$

Rearranging Equation (9.2), one obtains

$$S \propto G^{1/n} / \overline{D}^{[(m/q)-p]/n} \tag{9.3}$$

From Equation (9.1) it can be seen that m is larger for plan-concave slopes, where overland flow converges, than for plan-convex slopes, where overland flow diverges. The larger the value of m, the larger is the exponent (i.e. slope coefficient) of $\overline{D}$ in Equation (9.3), and the steeper is the $S - \overline{D}$ relation. Thus the analysis predicts that plan-concave slopes have steeper $S - \overline{D}$ relations than plan-convex ones. The agreement between the analysis and observed variation in the $S - \overline{D}$ relation with planform implies that the slopes are formed by and adjusted to hydraulic processes.

In the above sediment transport equation (Equation 9.1), the arithmetic mean particle size $\overline{D}$ was used as the measure of particle size because, of the several measures of particle size tested, it correlated most highly with gradient. Variable $\overline{D}$ worked best in this instance possibly because it represented resistance to flow, and the contribution by each piece of rock to flow resistance was additive rather than multiplicative. However, there are other measures of particle size, and in different circumstances they may be better predictors of G than is $\overline{D}$. For example, $\overline{D}$ is very insensitive to size of fines. Therefore where this property has a significant effect on G, a more sensitive sediment size variable should be used.

Laboratory experiments by Poesen and Lavee (1991) showed that the proportion of the surface covered with coarse debris (i.e. percentage stone cover) and the size of debris (stones) have an important influence on G (Fig. 9.6). Usually G decreases as stone cover increases due to increased resistance to flow and increased protection of the underlying fines. However, where stones are larger than about 50 mm and cover less than 70% of the surface, the opposite is true because the stones tend to concentrate the flow. Most interesting is the fact that for a given stone cover, G consistently increases with stone size, again because the stones tend to concentrate the flow. These findings by Poesen and Lavee suggest that although Equation (9.1) may be a useful start to the modelling of slope form, the situation on actual slopes is probably far more complex, and that a great deal more work is required to elucidate the effect of rock-fragment size and cover on sediment transport rate.

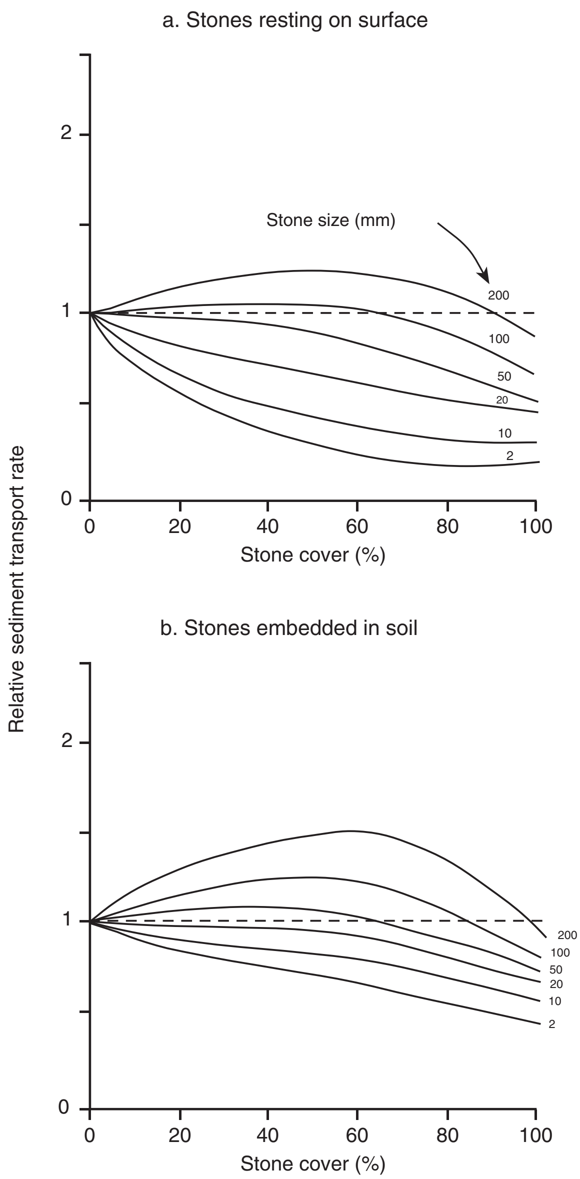

Fig. 9.6 Graphs showing relations between sediment transport rate and stone cover for different stone sizes: (**a**) stones resting on the soil surface, and (**b**) stones partially embedded in the soil. The graphs are generalizations of the experimental results of Poesen and Lavee (1991)

Two studies have provided direct evidence of the movement of stones on rock-mantled hillslopes in the American South-west. Kirkby and Kirkby (1974) painted lines across 12 hillslopes with gradient up to 20° in the Sonoran Desert of southern Arizona. During a two-month period they measured after each rainstorm the movement of all particles with diameters $\geq$1 mm. Field observations confirmed that the processes moving these particles were rainsplash and unconcentrated overland flow, and statistical analyses indicated that the distance moved was directly related to hillslope gradient and inversely related to grain size.

Abrahams et al. (1984) analysed 16 years of stone movement on two hillslopes with gradients up to 24° in the Mojave Desert, California. They found that the distance each particle moved was directly related to both length of overland flow (a surrogate for overland flow discharge) and hillslope gradient and inversely related to particle size. These results were interpreted as indicating that the stones, which ranged in size up to 65 mm, were moved mainly by hydraulic action. Citing Kirkby and Kirkby's findings as well as their own, Abrahams et al. (1984, p. 369) concluded 'that hydraulic action is probably the dominant process transporting coarse debris down hillslopes with gradients up to at least 24° over most of the Mojave and Sonoran Deserts'.

In a further study of the character and likely process of movement of debris on rock-mantled hillslopes, Abrahams et al. (1990) investigated the fabric of coarse particles mantling a debris slope on Bell Mountain in the Mojave Desert. The slope is typical of debris slopes in the Mojave Desert underlain by closely jointed or mechanically weak rocks. Samples of rod- and disc-shaped particles from five sites ranging in gradient from 11.7° to 33.17° were found to display essentially the same fabric: particles tend to be aligned downslope and to lie flat on the ground surface. There is no evidence of imbrication signifying sliding or creep nor of transverse modes indicating rolling. Abrahams et al. concluded that the fabric is probably produced by hydraulic action, and that this process is mainly responsible for moving coarse particles on gradients up to 33° on these debris slopes. Cumulative size distributions of the particles sampled at the two sites with gradients greater than 28° reveal that about 25% of the particles are larger than 64 mm and that the largest particle in each sample has a diameter in excess of 300 mm. It is difficult to imagine particles of this size being entrained by overland flow a few millimetres deep and transported as bed load. Abrahams et al. (1990) suggested that such particles may be moved downslope by a process termed runoff creep. De Ploey and Moeyersons (1975) observed this process on steep hillslopes in Nigeria and then replicated it in a laboratory flume. Their flume experiments disclosed that under the influence of overland flow (a) blocks shifted and tilted downslope when smaller gravel particles on which they were resting became wet and collapsed; (b) pebbles moved forward and tilted downslope during liquefaction of the underlying soil layer; (c) erosion of underlying finer material caused pebbles to settle downslope; and (d) scour on the upslope side of pebbles resulted in their being drawn into the holes and tilted upslope.

Piedmont Junctions

The sharp transition zone between upland areas and the piedmont characteristic of deserts has been variously referred to as the transition slope (Fair 1948), the nickpoint (Rahn 1966), the break in slope (Kirkby and Kirkby 1974), the piedmont angle (Twidale 1967, Young 1972, pp. 204–8, Cooke and Warren 1973, p. 199), and the piedmont junction (Mabbutt 1977, p. 82, Parsons and Abrahams 1984). In this chapter we use the term piedmont junction. At many locations the piedmont junction marks the boundary between the operation of different processes: for example, where an alluvial fan abuts against a hillslope. At other locations, the morphology of the piedmont junction is manifestly influenced by geological structure (e.g. Twidale 1967) or subsurface weathering (e.g. Twidale 1962, Mabbutt 1966). We are concerned with none of these situations here. Rather we focus on piedmont junctions that are simply concavities in slope profiles. These piedmont junctions occur at the transition between a pediment and its backing hillslope and may be defined as extending from 15° on the lower part of the backing hillslope to 5° on the upper part of the pediment (Kirkby and Kirkby 1974).

Piedmont junctions vary greatly in concavity. At one extreme are features that are so concave that they take the form of a true break in slope and can be identified only as a point on the hillslope profile (Fig. 9.7a). At the other extreme are features whose concavity is so slight that they can reach lengths of 750 m (Fig. 9.7b) (Kirkby and Kirkby 1974). Numerous workers have noted that piedmont junctions in many locations conform to two general tendencies. First, under a given climate they tend to vary in concavity from one rock type to another (e.g. Kirkby and Kirkby 1974, Mabbutt 1977, pp. 85–7). Second, on a given rock type they tend to decrease in concavity as precipitation increases (e.g. Bryan 1940, Fair 1947, 1948, Young 1972, p. 208, Mabbutt 1977, p. 85). The latter

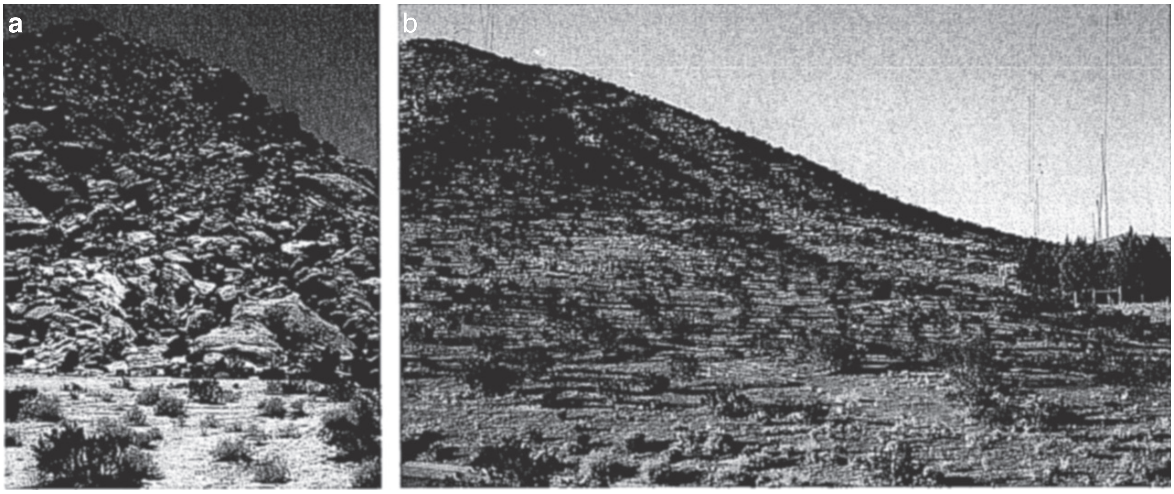

Fig. 9.7 Photographs of piedmont junctions, showing (**a**) a narrow, highly concave one formed on widely jointed quartz monzonite, Mojave Desert, California, and (**b**) a broad, gently concave one developed on gneiss, Mojave Desert, California

tendency is of course implicit in the fact that piedmont junctions (i.e. pronounced concavities) are generally associated with desert landscapes and not humid ones.

Both these tendencies derive from the fact that on slopes adjusted to present-day processes of sediment transport, gradient varies directly with particle size, which in turn varies inversely with distance downslope at different rates on different rock types and in different climates. Early workers claimed that gradient was related to particle size (e.g. Lawson 1915, Bryan 1922, Gilluly 1937), and in the latter part of the twentieth century this relation was verified quantitatively (e.g. Kirkby and Kirkby 1974, Abrahams et al. 1985). M.J. Kirkby (Carson and Kirkby 1972, pp. 346–7, Kirkby and Kirkby 1974) was perhaps the first to point out that different rock types have different comminution sequences and that, because hydraulic processes selectively transport finer particles further downslope than coarser ones, the comminution sequence is reflected in the downslope rate of change in particle size and, hence, gradient. At one extreme are rocks, such as basalts and schists, that have fairly continuous comminution sequences from boulder- to silt-sized particles. The slopes that form in these rocks exhibit a progressive decrease in particle size accompanied by a steady decline in gradient downslope, forming a broad and gently curving piedmont junction. At the other extreme are rocks, such as widely jointed granites, that are characterized by markedly discontinuous comminution sequences in which boulders disintegrate directly into granules and sands. On these rocks, steep backing slopes mantled with boulders give way abruptly downslope to gentle pediments covered with granules and sands, and an extremely narrow, almost angular piedmont junction is produced. It is therefore evident that given the relationship between hillslope gradient and particle size, the concavity of piedmont junctions in a desert climate depends on the comminution sequence of the underlying rock.

The same line of reasoning may be applied to explaining the variation in piedmont junction concavity with climate. In desert climates, particle size decreases across piedmont junctions in accordance with the comminution sequence of the underlying rock, as explained above. In humid climates, on the other hand, soils with fine-grained A horizons are developed on both the backing hillslopes and the footslopes, and the size of surface particles decreases very little, if at all, downslope (e.g. Furley 1968, Birkeland 1974, p. 186). Because the decrease in particle size across the piedmont junction is more pronounced in desert climates than in humid ones, the piedmont junctions are typically narrower and more concave.

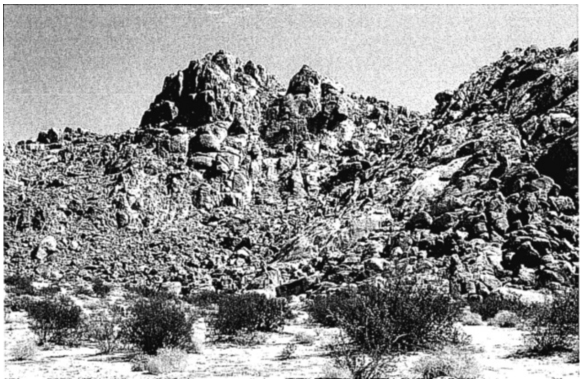

Fig. 9.8 Photograph of a steep, poorly adjusted debris slope developed on widely jointed quartz monzonite in the Mojave Desert, California

The foregoing discussion applies to slopes that are adjusted to present-day processes, at least in the vicinity of the piedmont junction. However, not all slopes are so adjusted. Where they are not, particle size may be unrelated to gradient, and the preceding analysis is irrelevant. The situation most commonly encountered is where a pediment covered with fines and presumably adjusted to contemporary processes is backed by a weathering-limited slope that is clearly not adjusted to current transport processes (Fig. 9.8). The form of the backing slope might be controlled by rock mass strength (Selby 1980, 1982a, b, pp. 199–203) or rock structure (Oberlander 1972) or inherited from a previous climate. In such circumstances, the concavity of the piedmont junction cannot be understood in terms of contemporary hydraulic processes. About all that can be said about piedmont junctions of this type is that they tend to be more concave than most. The reason for this is that in a given (desert) climate, steep backing slopes are more likely to become weathering-limited than are gentle ones, and piedmont junctions with steep backing slopes are likely to be more concave than those with gentle backing slopes.

Processes on Rock-Mantled Slopes

Hydraulic Processes

Virtually all runoff from desert hillslopes occurs in the form of overland flow that is generated when the rainfall intensity exceeds the surface infiltration rate. Such rainfall-excess overland flow is widely termed Hortonian overland flow (Horton 1933). Because hydraulic processes are, therefore dependent on infiltration rates, understanding infiltration is central to understanding runoff and erosion on rock-mantled slopes.

Infiltration on Rock-Mantled Slopes

As in all environments, infiltration through the surface layer of rock-mantled hillslopes is controlled by its physical and chemical properties (see, for example, Mills et al., 2006) However, what is particularly important for hydraulic processes on desert hillslopes is the great variation in infiltration that they exhibit, both spatial and temporal. This variation may reflect variation in surface or subsurface properties. Although much less is known about the role of the latter than the former, Perrolf and Sandstrom (1995), in a study undertaken in Botswana and Tanzania, showed that variations in subsoil conditions were responsible for only a fivefold variability in infiltration, compared to twentyfold differences due to variations in surface conditions. Among the properties controlling infiltration and runoff are the ratio of bedrock to soil, surface and subsurface stone size, stone cover, vegetation and surface sealing.

Given the widespread occurrence of bedrock outcrops on many desert hillslopes, an important control of infiltration and runoff is the ratio of bedrock to soil. Figure 9.9 shows the infiltration curves for rocky and soil-covered surfaces at Sede Boqer and the Hovav Plateau in the northern Negev, Israel (Yair 1987). The infiltration capacities are lower for the bedrock than for the soil-covered surfaces at both sites. The difference is especially pronounced for the Sede Boqer site because the rock is a smooth, massive crystalline limestone, whereas at the Hovav Plateau it is densely jointed and chalky. Data from natural rainfall events at Sede Boqer (Yair 1983) indicate that the threshold level of daily rainfall necessary to generate runoff in the rocky areas is 1–3 mm, whereas it is 3–5 mm for the colluvial soils. As rain showers of less than 3 mm represent 60% of the rain events, the frequency and magnitude of runoff events are both much greater on the rocky than on the soil-covered areas.

Even where bedrock outcrops are absent, desert soils are typically stony. The effect of surface stones on runoff is quite complex and has been the subject of numerous field and laboratory studies (e.g. Jung 1960, Seginer et al. 1962, Epstein et al. 1966, Yair and Klein 1973, Yair and Lavee 1976, Box 1981, Poesen et al. 1990, Abrahams and Parsons 1991a, Lavee and Poesen 1991, Poesen and Lavee 1991). Figure 9.10, which is based on laboratory experiments by Poesen and Lavee (1991, Fig. 3), summarizes the state of knowledge for surfaces devoid of vegetation. Basically, surface stones affect runoff by two groups of mechanisms. First, increasing stone size and stone cover increasingly protect the soil surface from raindrop impact and thereby inhibit surface sealing and reduce runoff. Increasing stone size and stone cover also increase depression storage which promotes infiltration. Second, increasing stone size and stone cover result in greater quantities of water being shed by the stones (stone flow) and concentrated in the interstone areas, where the water overwhelms the ability of the underlying soil to absorb it and runs off in increasing amounts. Both groups of mechanisms operate simultaneously. In general, it appears that as stone size increases the second group dominates. As a result, runoff increases with stone size irrespective of stone cover. The relation between runoff and stone cover is less straightforward. Where stone sizes and stone covers are small, the

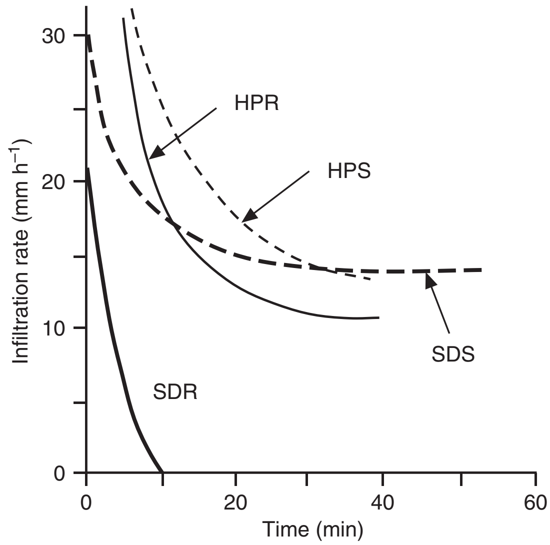

Fig. 9.9 Infiltration curves for rocky and soil-covered surfaces in the northern Negev, Israel: Sede Boqer (rainfall intensity $36\,\mathrm{mm\,h^{-1}}$) massive limestone (SDR), Sede Boqer stony colluvium soil (SDS), Hovav Plateau (rain intensity $33\,\mathrm{mm\,h^{-1}}$) densely jointed and chalky limestone (HPR), and Hovav Plateau stoneless colluvial soil (HPS) (after Yair 1987)

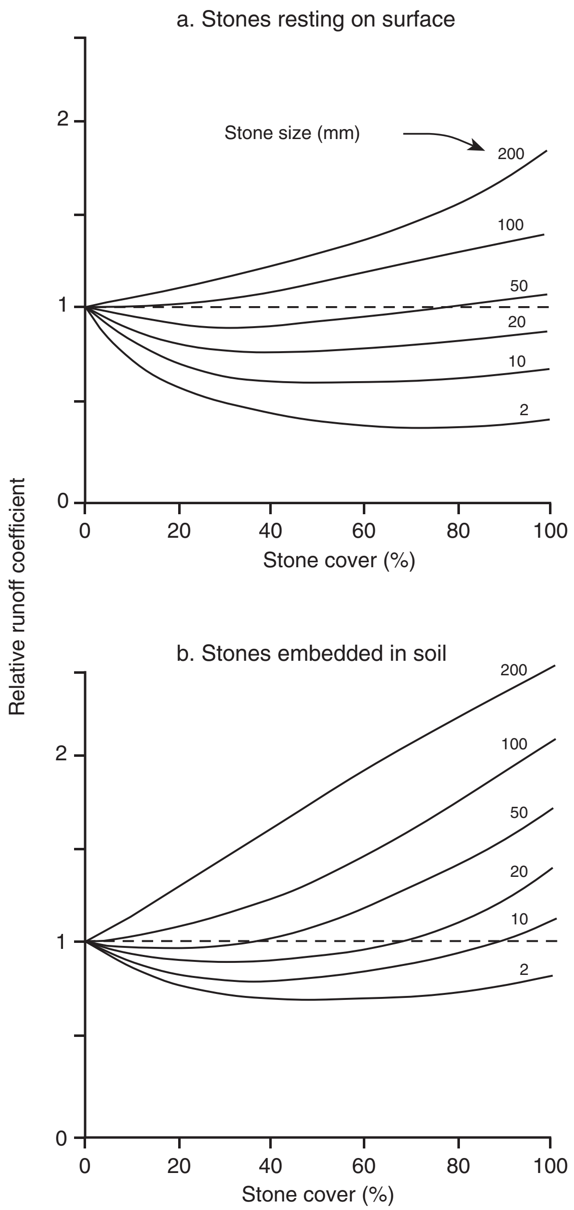

Fig. 9.10 Graphs showing relations between runoff coefficient and stone cover for different stone sizes: (**a**) stones resting on the soil surface and (**b**) stones partially embedded in the soil. The graphs are generalizations of the experimental results of Poesen and Lavee (1991)

first group of mechanisms dominates, and runoff is negatively related to stone cover. However, for other combinations of stone size and stone cover the second group dominates, and runoff is positively related to stone cover. Stone position also affects runoff. A comparison of Figs. 9.10a, b reveals that where stones are embedded in the soil runoff rates are higher than where they are resting on the surface. Interestingly, for intermediate (mean) stone sizes (i.e. 20–50 mm), the sign of the relation between runoff and stone cover may actually change from negative for stones resting on the surface to positive for stones embedded in the soil. (Poesen 1990, Poesen et al. 1990).

Figure 9.10 applies to areas devoid of vegetation. Where there is a significant vegetation cover, particularly of shrubs, the controls of infiltration and runoff are quite different. This is reflected in the correlation between infiltration and stone cover. Abrahams and Parsons (1991b) noted that both positive and negative correlations between infiltration and stone cover have been reported for semiarid hillslopes in the American South-west. They observed that positive correlations (Tromble 1976, Abrahams and Parsons 1991a) have been obtained when infiltration measurements were confined to (bare) stone-covered areas between shrubs (lower curves in Fig. 9.10), and they attributed these correlations to increasing stone cover progressively impeding surface sealing. In contrast, negative correlations have been found when infiltration was measured in shrub as well as intershrub areas (e.g. Tromble et al. 1974, Simanton and Renard 1982, Wilcox et al. 1988, Abrahams and Parsons 1991b). Abrahams and Parsons ascribed these correlations to infiltration rates under shrubs being greater than those between shrubs (Lyford and Qashu 1969), and percentage stone cover being negatively correlated to percentage shrub canopy (Wilcox et al. 1988). The mechanisms giving rise to higher infiltration rates under shrubs than between them are summarized in Fig. 9.11. As might be expected, positive correlations have been recorded between infiltration rate and percentage plant canopy (Kincaid et al. 1964, Simanton et al. 1973, Tromble et al. 1974).

In recent years, the role of surface crusts in controlling infiltration has achieved greater recognition. These crusts are of two types: mechanical and biological. Mechanical crusts are formed by one or more of raindrop impact, trapped gas bubbles (forming a vesicular crust) and evaporation (forming chemical crusts, e.g. Romao and Escudero 2005). Biological crusts are created by an association between soil particles and cyanobacteria, lichens and mosses. Airborne silts and clays are trapped by sticky cyanobacterial sheaths, resulting in a thin surface layer of silts and clays that are often lacking where biological crusts are absent. In

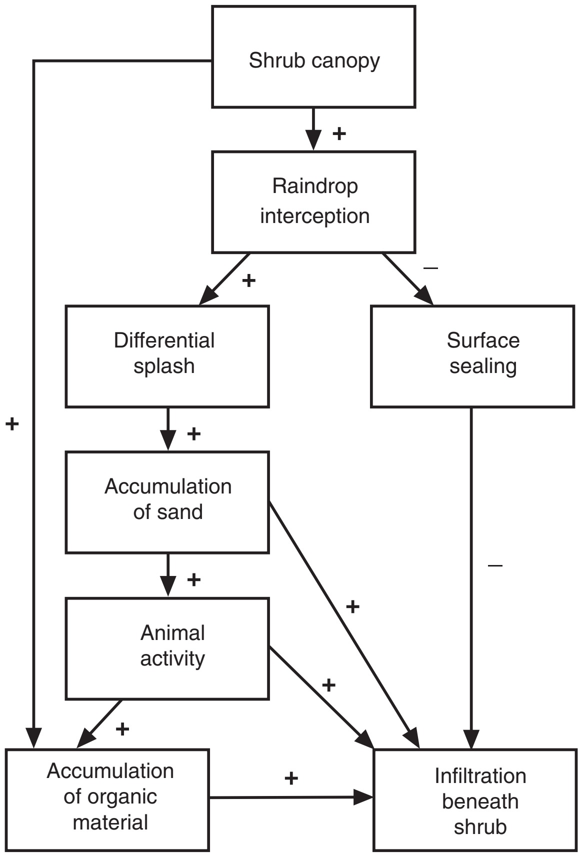

Fig. 9.11 Causal diagram showing the mechanisms whereby a shrub's canopy promotes infiltration under the shrub at Walnut Gulch, Arizona (after Abrahams and Parsons 1991a)

general, it is believed that the existence of crusts decreases surface permeability and infiltration (e.g. Wang et al. 2007), but debate still exists on the effects of biological crusts (see Belnap 2006, for a review). All crusts are fragile and may be disturbed by faunal activity, vehicles and penetration by large raindrops. Quantification the effects of crusts on infiltration relative to other factors remains unresolved (Belnap 2006).

In addition to the effects of crusts, many dryland soils are hydrophobic. This hydrophobicity may arise from the effects of fires, where spatial variability in fire intensity results in spatial variability in hydrophobicity (e.g. Ferreira et al. 2005), or from chemicals exuded by desert plants and the effects of organic debris (Cammeraat et al. 2002).

Finally, spatial variability in infiltration may result from faunal activity. Animals affect infiltration in many ways from grazing and consequent reduction in vegetation cover to disturbance of soil crusts, but probably their greatest influence on infiltration is in their creation of macropores through burrowing. In deserts, termites are probably the most significant producers of macropores. Mando and Miedema (1997), in an experimental study in Burkina Faso to assess the effects of termites on degraded soils, found 60% of macropores to be due to termite activity, accounting for 38.6% of total soil porosity in the top 7 cm of the soil. Similarly, working in Senegal, Sarr et al. (2001) showed that infiltration rates were about 80% lower in plots from which termites had been excluded. In contrast, however, Debruyn and Conacher (1994) found that micropores produced by ants only affected infiltration rates when the soil is saturated and water is ponded on the surface.

Not only do desert hillslopes exhibit spatial variability in infiltration, but they also exhibit temporal variability. At a seasonal scale variations in antecedent soil mositure may affect infiltration (Fitzjohn et al. 1998), and this effect may vary with rainfall intensity (Castillo et al. 2003). Simanton and Renard (1982) identified differences between spring and autumn infiltration which they attributed to the effects of frost and wetting and drying in the winter and rainfall in the summer. Processes in the winter loosened the soil surface, whereas those in the summer compacted it. Over longer timespans fire-induced hydrophobicity, for example, changes. Although it is generally found to decline, Cerda and Doerr (2005) found topsoil hydrophobicity actually increased with time under *Pinus halepensis*.

The spatial variability in the factors controlling point infiltration rates means that overland flow is not generated uniformly over a desert hillslope but preferentially from those parts of the hillslope where the infiltration rates are lowest. Overland flow generated then travels downslope where it may encounter other areas whose infiltration capacity remains higher than the rainfall intensity. Some or all of the flow may infiltrate into these areas (Smith and Hebbert 1979, Hawkins and Cundy 1987). As a result of this runon infiltration, runoff per unit area may decrease with the size of the area being investigated. This phenomenon is illustrated in Fig. 9.12 for a small piedmont watershed at Walnut Gulch, Arizona (Kincaid et al. 1966).

Not only may spatial variability in point infiltration cause a decrease in runoff per unit area as scale increases so, too, may temporal variability in rainfall input. Most studies of infiltration have used ring infiltrometers or constant-intensity simulated rainfall and have thus neglected the impact of temporal variability

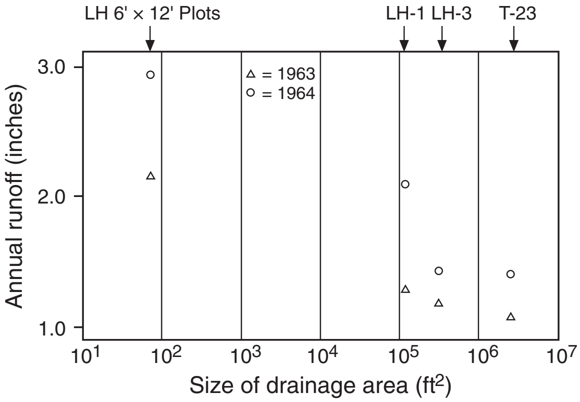

Fig. 9.12 Graph of annual runoff against size of drainage area for runoff plots and very small watersheds ($>10^5$ ft^2) at Walnut Gulch, Arizona (after Kincaid et al. 1966)

of rainfall input on runon infiltration. Wainwright and Parsons (2002) showed in a modelling study that temporal variability in rainfall input is likely to be a larger contributor to runon infiltration than is spatial variability in point infiltration. Similarly, De Lima and Singh (2002) and Reaney et al. (2007) demonstrated that storms delivering the same rainfall amount could yield different amounts of runoff depending on the temporal distribution of the rainfall.

Unconcentrated Runoff

Runoff on desert hillslopes generally first appears as an unconcentrated sheet of water with threads of deeper, faster flow diverging and converging around surface protuberances, rocks, and vegetation. As a result of these diverging and converging threads, flow depth and velocity may vary markedly over short distances, giving rise to changes in the state of flow. Thus over a small area the flow may be wholly laminar, wholly turbulent, wholly transitional, or consist of patches of any of these three flow states. Much research on the hydraulics of unconcentrated overland flow has focused on the controls on resistance offered to the flow by the rough surface that characterises rock-mantled hillslopes. Resistance to overland flow may be quantified by the dimensionless Darcy–Weisbach friction factor

$$f = 8ghS/V^2 \tag{9.4}$$

where g is the acceleration of gravity, h the mean depth of flow, S the energy slope, and V the mean flow velocity. Flow resistance f consists of grain resistance, form resistance, rain resistance and sediment-transport resistance. Grain resistance f_g is imparted by soil particles and microaggregates that protrude into the flow less than about ten times the thickness of the viscous sublayer (Yen 1965). Form resistance f_f is exerted by microtopographic protuberances, stones, and vegetation that protrude further into the flow and control the shape of the flow cross-sections (Sadeghian and Mitchell 1990). Rain resistance f_r is due to velocity retardation as flow momentum is transferred to accelerate the raindrop mass from zero velocity to the velocity of the flow (Yoon and Wenzel 1971). Finally, sediment-transport resistance f_t is due to velocity retardation as flow momentum is transferred to accelerate sediment mass from zero velocity to its transport velocity (Abrahams and Li 1998). For laminar flow on gentle slopes f_r may attain 20% of f (Savat 1977). However, generally it is a much smaller proportion, and the proportion becomes still smaller as the state of flow changes from transitional to turbulent (Yoon and Wenzel 1971, Shen and Li 1973). Because f_r is typically several orders of magnitude less than f on desert hillslopes (Dunne and Dietrich 1980), and similarly f_t is likely to be small on rough hillslopes (Abrahams and Li 1998), the following discussion will focus on f_g and f_f.

Resistance to flow generally varies with the intensity of flow, which is represented by the dimensionless Reynolds Number

$$R_e = 4Vh/v \tag{9.5}$$

where v is the kinematic fluid viscosity. Laboratory experiments and theoretical analyses since the 1930s have established that where f is due entirely to grain resistance the power relation between f and R_e for shallow flow over a plane bed is a function of the state of flow. The relation has a slope of -1.0 where the flow is laminar and a slope close to -0.25 where it is turbulent. This relation between f and R_e (or surrogates thereof) for plane beds has been widely used in models of hillslope runoff. However, the surfaces of desert hillslopes are rarely, if ever, planar, and the anastomosing pattern of overland flow around microtopographic protuberances, rocks, and vegetation attests to the importance of form resistance. If form resistance is important, its influence might be expected to be reflected in the shape of the f–R_e relation.

This was first recognized in a set of field experiments conducted by Abrahams et al. (1986) on small runoff plots located in intershrub areas on piedmont hillslopes at Walnut Gulch, southern Arizona. Although the plot surfaces were mantled with gravel, clipped plant stems occupied as much as 10% of their area, and the steeper plots had quite irregular surfaces. Analyses of 14 cross-sections yielded f–R_e relations that were positively sloping, negatively sloping, and convex-apward (Fig. 9.13). These shapes were attributed to the progressive inundation of the roughness elements (i.e. gravel, plant stems, and microtopographic protuberances) that impart form resistance. So long as these elements are emergent from the flow, f increases with R_e as the upstream wetted projected area of the elements increases. However, once the elements become submerged, f decreases as R_e increases and the ability of the elements to retard the flow progressively decreases.

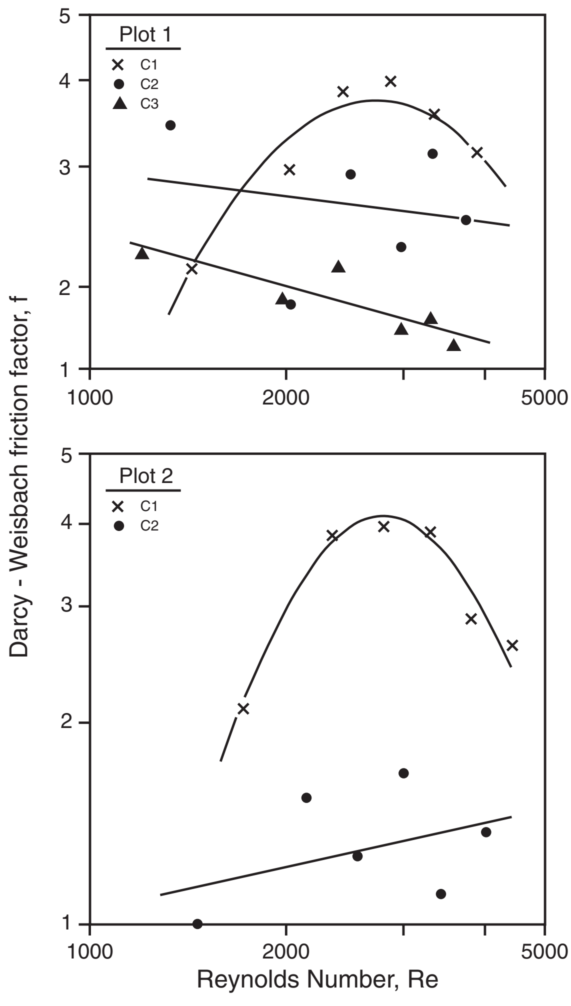

Fig. 9.13 Graphs of Darcy–Weisbach friction factor against Reynolds Number for five cross-sections on two runoff plots at Walnut Gulch, Arizona. The cross-sections are denoted by C1, C2, etc. (after Abrahams et al. 1986)

In a second set of field experiments on small plots sited in gravel-covered intershrub areas at Walnut Gulch, Abrahams and Parsons (1991c) obtained the regression equation

$$\log f = -5.960 - 0.306 \log R_e + 3.481 \log \%G + 0.998 \log D_g \tag{9.6}$$

where $\%G$ is the percentage of the surface covered with gravel, D_g is the mean size of the gravel (mm), and $R^2 = 0.61$. Of the independent variables in Equation (8.4), $\%G$ was by far the single best predictor of f, explaining 50.1% of the variance. The dominance of $\%G$ implies that $f_f >> f_g$ on these gravel-covered hillslopes. This was confirmed using a procedure developed by Govers and Rauws (1986) for calculating the relative magnitudes of f_g and f_f in overland flow. For the 73 experiments performed on the small plots the modal and median values of $\% f_g$, which denotes grain resistance expressed as a percentage of total resistance, were 4.55% and 4.53% (Fig. 9.14). Thus on these gravel-covered hillslopes, f_g is typically about one-twentieth of f_f. This conclusion has important implications for sediment transport which will be explored below.

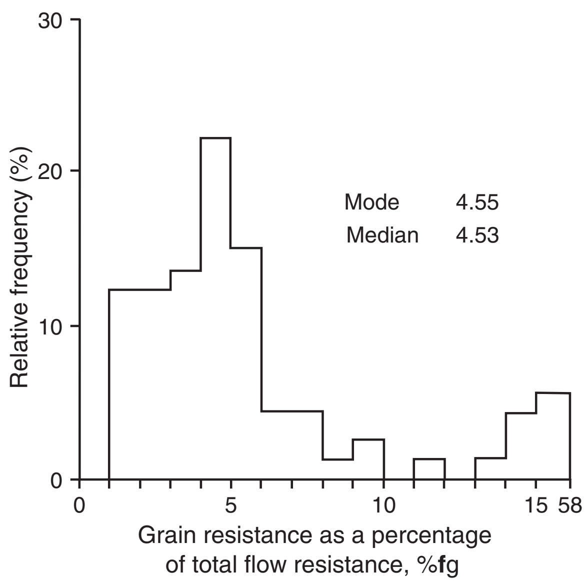

Fig. 9.14 Relative frequency distribution of grain resistance expressed as a percentage of total flow resistance for 73 experiments on 8 runoff plots at Walnut Gulch, Arizona

The findings of Abrahams et al. (1986) and Abrahams and Parsons (1991c) are supported by laboratory experiments by Gilley et al. (1992) in which varying rates of flow were introduced into a flume covered with different concentrations and sizes of gravel. Gilley et al. also recorded positively sloping, negatively sloping, and convex-upward f–R_e relations which they attributed to the progressive inundation of the gravel. In addition, they obtained regression equations of the form

$$\log f = \log a_1 - a_2 \log R_e + a_3 \log \%G \qquad (9.7)$$

for each gravel size class. The R^2 value for each regression exceeded 0.94. These flume experiments confirm the important role of gravel cover in controlling resistance to overland flow through its influence on form resistance. The importance of rock-fragment size was shown in a study by Bunte and Poesen (1994). These authors found significant differences between pebbles (with mean b-axis of 15 mm) and cobbles (mean b-axis 86 mm) in the changes to flow hydraulics with percentage cover, particularly at low cover percentages (Fig. 9.15). They attributed these differences to the more reticular flow around the smaller pebbles.

Because rock-mantled hillslopes are also typically vegetated, rock fragments are not the only contributor to form roughness. In a laboratory experiment, Dunkerley et al. (2001) showed that for a given percentage cover, plant litter increases resistance to flow more than do rock fragments. Dunkerley (2003) used a similar argument to that used by Bunte and Poesen (1994) to explain this effect, and it is noteworthy that the rock-fragment cover percentages used in the experiments are relatively low, so the results are consistent with the differences found by Bunte and Poesen. Furthermore, Dunkerley et al. (2001) showed that for flows wholly within the laminar range f consistently declined with Re, in contrast to observations of more complex relationships, and argued that flow regime is a significant control on the form of the relationship. Away from the laboratory, flow on natural hillslopes is almost always a mixture of wholly laminar, wholly turbulent and wholly transitional, termed by Abrahams et al. (1986) *composite flow*. In more recent work, Dunkerley (2004) has argued that deriving a simple, unweighted average value for f for such composite flow biases the value towards the high-resistance shallow laminar flow, and that estimates of average f should be weighted according to discharge in the different elements of composite flow.

The hydraulics of unconcentrated overland flow over entire hillslopes were investigated by Parsons et al. (1990, 1996) using simulated rainfall on plots 18 m wide and approximately 30 m long located on shrub-covered and grass-covered piedmont hillslopes at Walnut Gulch. The h, V, indundated width w and discharge Q values were computed for two measured sections situated 12.5 and 21 m from the top of the plot on the shrub-covered plot, and 6, 12 and 20.5 m from the top of the plot on the grassland plot. (Figures 9.16 and 9.17, respectively) At-a-section h–Q w-Q and V–Q relations show that increases in Q are accommodated by increases in h and w. On both plots, increases in discharge result in minimal change to mean velocity. Downslope hydraulic relations differ strikingly from at-a-section relations and between the two vegetation types. Under equilibrium (steady state) runoff conditions on the shrubland, f decreases rapidly as Q increases, permitting increases in Q to be accommodated almost entirely by increases in V. The decrease in f is due to the progressive downslope concentration of flow into fewer, larger threads. Under non-equilibrium conditions, downslope hydraulic relations are different from those at equilibrium, but f always decreases downslope. This is the result of low flows following pathways formed by higher flows that concentrate downslope. On the grassland downslope increases in Q are accommodated more-or-less equally by increases in V and h because flow does not concentrate downslope to the degree that it does on the shrubland.

Concentrated Runoff

The tendency for threads within unconcentrated overland flow to increase in depth and velocity downslope coupled with the convergence (and divergence) of these threads around obstructions may lead to the formation of small channels (rills). Such features are very common on desert hillslopes, particularly where the underlying material is easily eroded. However, in contrast to their study on agricultural land, there are few studies of rills on desert hillslopes. Abrahams et al. (1996) undertook a study of rills on a shrub-covered hillslope in southern Arizona. Although they

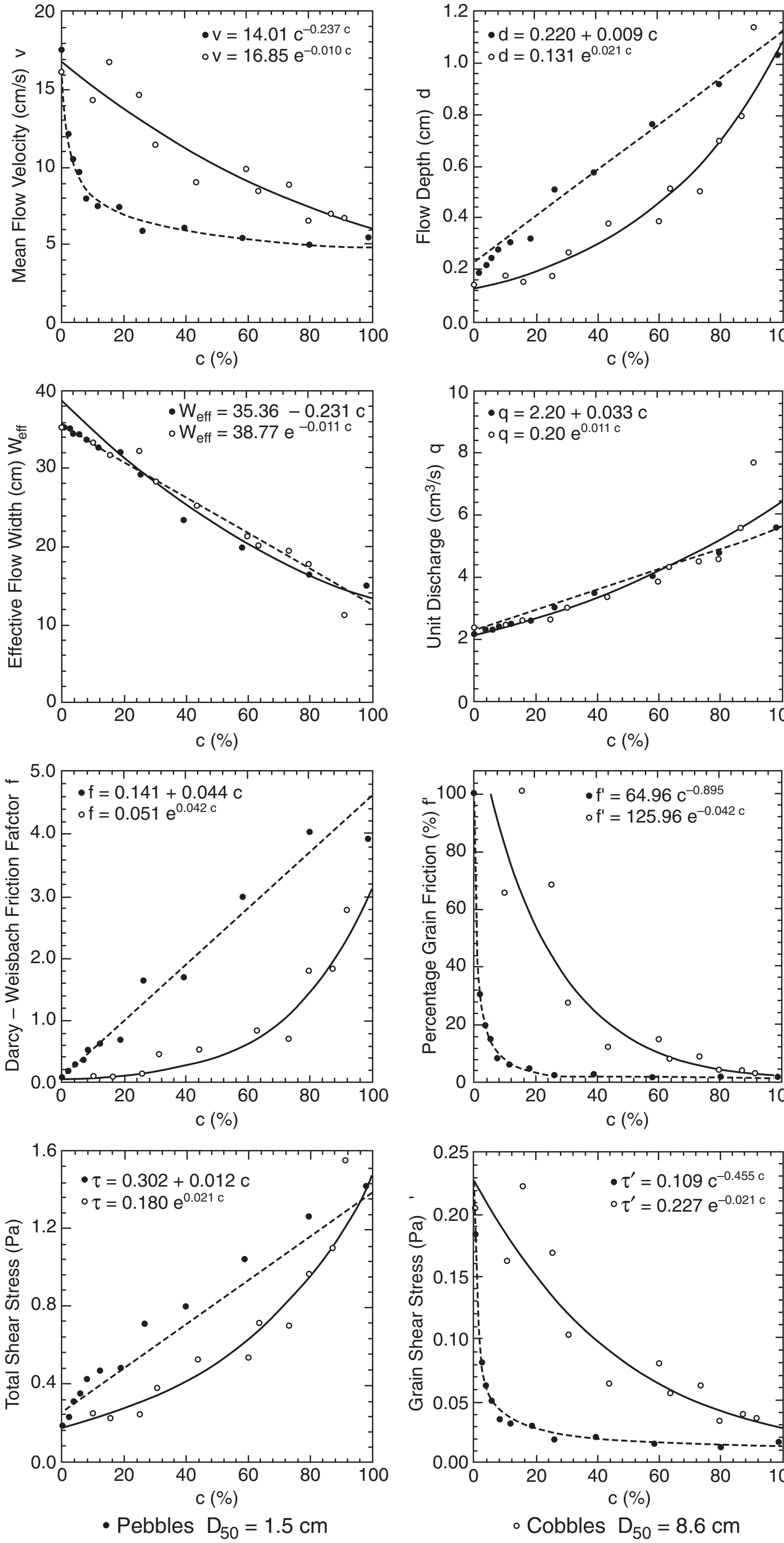

Fig. 9.15 Differences in flow hydraulics for surfaces partially covered with pebbles and cobbles (after Bunte and Poesen 1994)

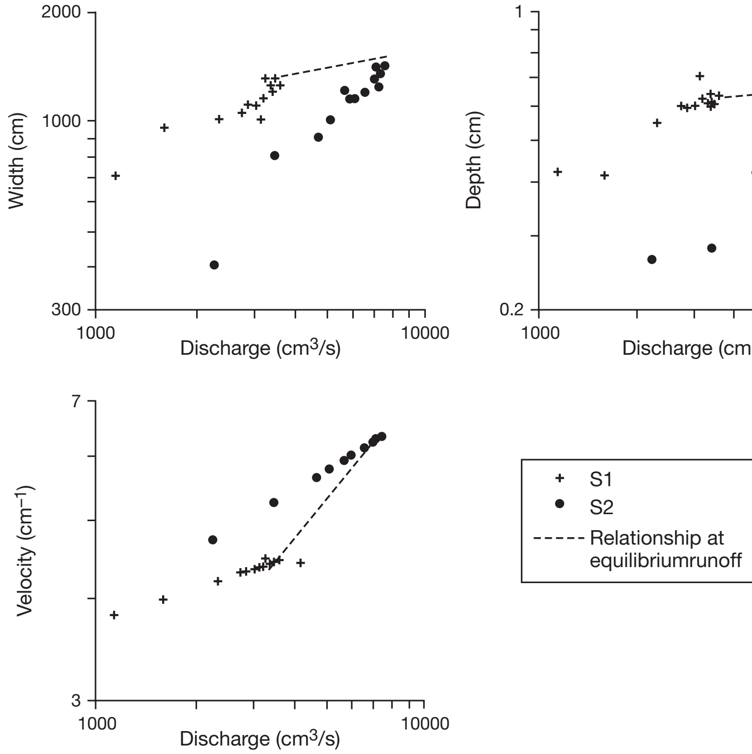

Fig. 9.16 Flow hydraulics at two cross sections on a large shrubland plot at Walnut Gulch, Arizona (after Parsons et al. 1996)

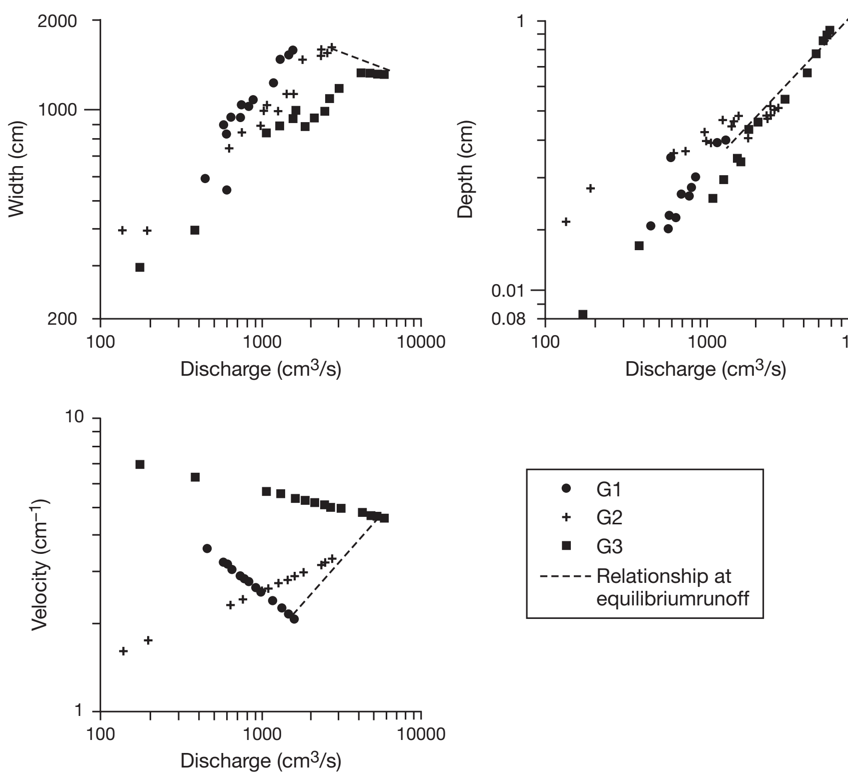

Fig. 9.17 Flow hydraulics at three cross sections on a large grassland plot at Walnut Gulch, Arizona (after Parsons et al. 1996)

found only a small difference in at-a-station hydraulic geometry between these rills and their agricultural counterparts that was not statistically significant, they argued that the difference was real and reflected the tendency for these rills to be wider and shallower than those on agricultural land. Furthermore, whereas Govers (1992) argued that a general relationship existed for rills in which flow velocity u depended on discharge Q and was unaffected by slope S or soil materials, the rills studied by Abrahams et al. showed a consistently lower velocity for a given discharge than the data presented by Govers (Fig. 9.18), and, in a multiple regression equation to predict flow velocity, Abrahams et al. found that both slope and percentage of gravel-sized particles (≥ 2 mm) $\%G$ were significant independent variables in an equation to predict rill flow velocity

$$\log u = 0.672 + 0.330 \log Q - 0.00415\%G + 0.0664 \log S, \quad (9.8)$$

which had an R^2 of 0.859. In a more recent laboratory experiment, Rieke-Zapp et al. (2007) also found increasing resistance to flow with increasing percentage of gravel cover. However this effect was less evident at higher discharges and at the higher of the two slopes at which their experiments were conducted.

Erosion by Hydraulic Processes

Rates

There are very few data on rates of erosion by hydraulic processes on desert hillslopes. A survey by Saunders and Young (1983) indicated that rates exceed 1 mm y^{-1} on normal rocks in semi-arid climates but are less than 0.01 mm y^{-1} in arid climates. These rates of erosion for semi-arid climates are amongst the highest in the world. Although debris flows may be an important agent of erosion on slopes steeper than 30°, Young and Saunders (1986) concluded that hydraulic action is the predominant denudational process in semi-arid climates, and probably in arid ones as well. Within a given climate, however, there is considerable variability in rates of hydraulic erosion, even over a single hillslope. This variability is largely the result of differences in surface properties affecting runoff generation and sediment supply. Among these properties are stone size, stone cover, vegetation cover, and biotic activity.

Erosion by Unconcentrated Runoff

Controlling Factors

Abrahams and Parsons (1991a) investigated the relation between hydraulic erosion and gradient at Walnut

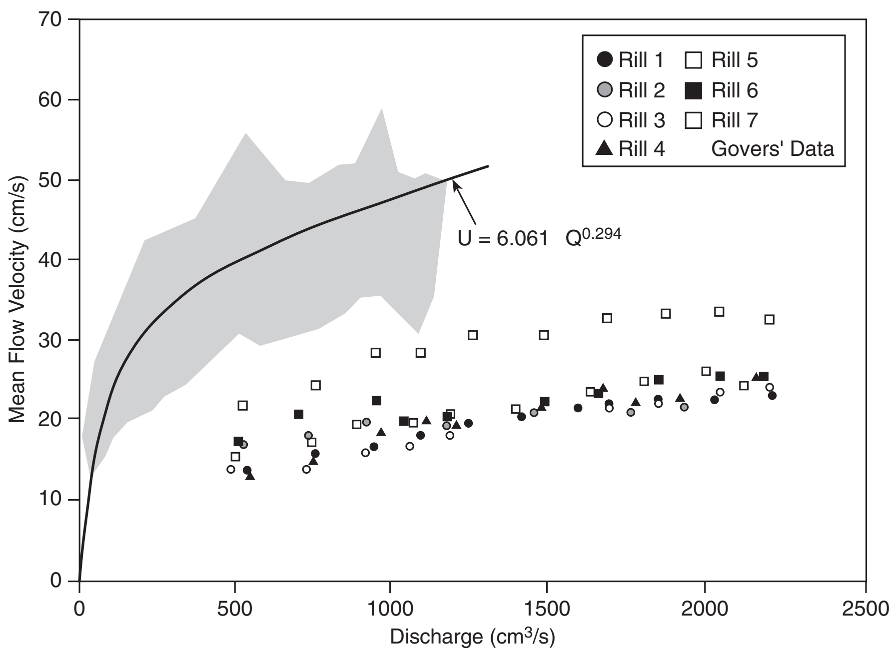

Fig. 9.18 Graph of mean flow velocity against discharge for seven rills in shrubland at Walnut Gulch, Arizona compared to Govers' (1992) data (shaded) and best-fit equation (after Abrahams et al. 1996)

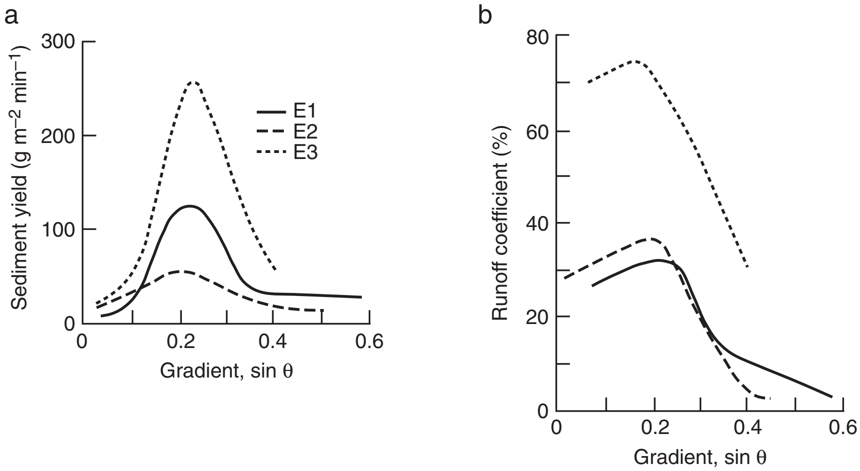

Fig. 9.19 Curves fitted to graphs of (**a**) sediment yield and (**b**) runoff coefficient against gradient for three sets of experiments denoted by E1, E2, and E3 at Walnut Gulch, Arizona. Experiments E1 and E2 were conducted on plots underlain by Quaternary alluvium, with the ground vegetation being clipped for E1 but not for E2. Experiment E3 was performed on plots underlain by the Bisbee Formation

Gulch by conducting three sets of field experiments on small runoff plots under simulated rainfall on two different substrates. Each set of experiments yielded a convex-upward sediment-yield–gradient relation with a vertex at about 12° (Fig. 9.19). The key to understanding this relation is the relation between runoff and gradient. On slopes less than 12° runoff increases very slowly with gradient, so sediment yield increases with gradient mainly in response to the increase in the downslope component of gravity. On slopes steeper than 12° runoff decreases rapidly as gradient increases. This decrease in runoff outweighs the increase in the downslope component of gravity and causes sediment yield to decrease.

Although sediment yield is curvilinearly related to gradient, it is actually controlled in a complex way by a combination of stone size, surface roughness, and gradient. The nature of this control is outlined in Fig. 9.20 (Abrahams et al. 1988). Where gradients exceed 12° sediment yield is positively correlated with runoff which, in turn, is negatively correlated with gradient, stone size, and surface roughness (Yair and Klein 1973). Where gradients are less than 12° runoff is almost constant, and sediment yield is positively correlated with these variables. Thus the controls of sediment yield depend on the range of gradient being considered. Where gradients exceed 12° stone size and surface roughness have a strong influence on runoff and, through runoff, affect sediment yield. On the other hand, where gradients are less than 12°, stone size and surface roughness have little effect on runoff. However, they are correlated with gradient, and gradient determines sediment yield. The interesting question raised by these results for slopes steeper than 12° is if stoniness increases with gradient causing runoff and erosion to decrease, how does one explain the increase of stoniness with gradient? The most likely explanation is that the small plot experiments that produced the above results do not take into account overland flow from upslope which would presumably be highly effective in eroding the steeper portions of desert hillslopes.

The relation between stone cover and sediment yield on semi-arid hillslopes has been investigated by Iverson (1980) and Simanton et al. (1984) using simulated rainfall. For 21 plots in the Mojave Desert, California, Iverson obtained a correlation of −0.56 between sediment yield and percentage stones (>2 mm) in the surface soil. However, these plots ranged in gradient from

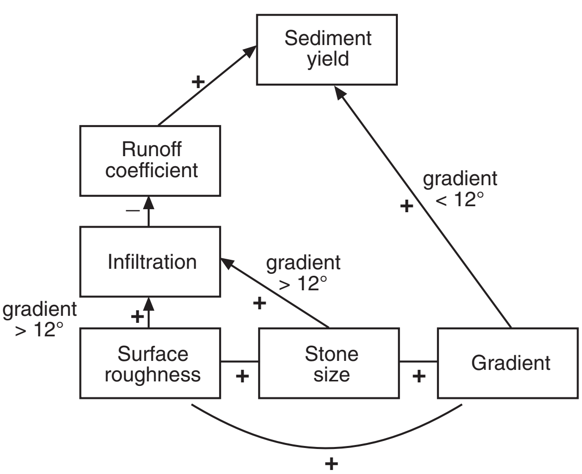

Fig. 9.20 Causal diagram showing the factors controlling the runoff coefficient and sediment yield on desert hillslopes

4° to 25°, which contributed greatly to the scatter. In a better controlled study in which all the plots had similar gradients (5.1–6.8°), Simanton et al. obtained a correlation of −0.98 between sediment yield and percentage stone cover (>5 mm) for eight plots at Walnut Gulch, Arizona. These negative correlations can be attributed to several factors: the stones protect the soil structure against aggregate breakdown and surface sealing by raindrop impact; enhance infiltration and diminish runoff; increase surface roughness which decreases overland flow velocities; and reduce soil detachment and, hence, interrill erosion rates (Poesen 1990).

Laboratory experiments by Poesen and Lavee (1991), however, suggest that the correlation between sediment yield and stone cover is not always negative. Figure 9.6, which is a generalization of Poesen and Lavee's results, indicates that the correlation becomes positive where the stones are embedded in the soil and are larger than 50 mm. In these circumstances, the increasing stone-flow effect outweighs the increasing protection-from-raindrop-impact and flow-retardation effects as stone cover increases, and the increasing concentration of water between the stones results in greater flow detachment and transport of soil particles. However, once stone cover increases above about 70%, sediment yield begins to decline toward a minimum at 100%, when the stone cover affords complete protection of the soil beneath. Poesen and Lavee's experiments also show that for a given stone cover, sediment yield consistently increases with stone size due to increasing stone flow.

Simanton and Renard (1982) used simulated rainfall to examine seasonal variations in the erosion of three soils at Walnut Gulch. In the spring the soil surface is loose due mainly to freeze–thaw during the preceding winter, whereas in the autumn it is compacted as a result of summer thunderstorms. Nevertheless, sediment yields in the spring are not always greater than those in the autumn. Figure 9.21a shows that the change in sediment yield is closely related to the change in runoff, which is inversely related to the percentage of the surface covered with stones (>2 mm). This relation can be attributed at least in part to an increase in stone cover inhibiting surface sealing. However, between spring and autumn there is also an increase in vegetation cover in response to the summer rains. This increase is negatively correlated with the change in runoff and sediment yield (Fig. 9.21b), suggesting that the change in sediment yield is also a function of differences in summer vegetation growth.

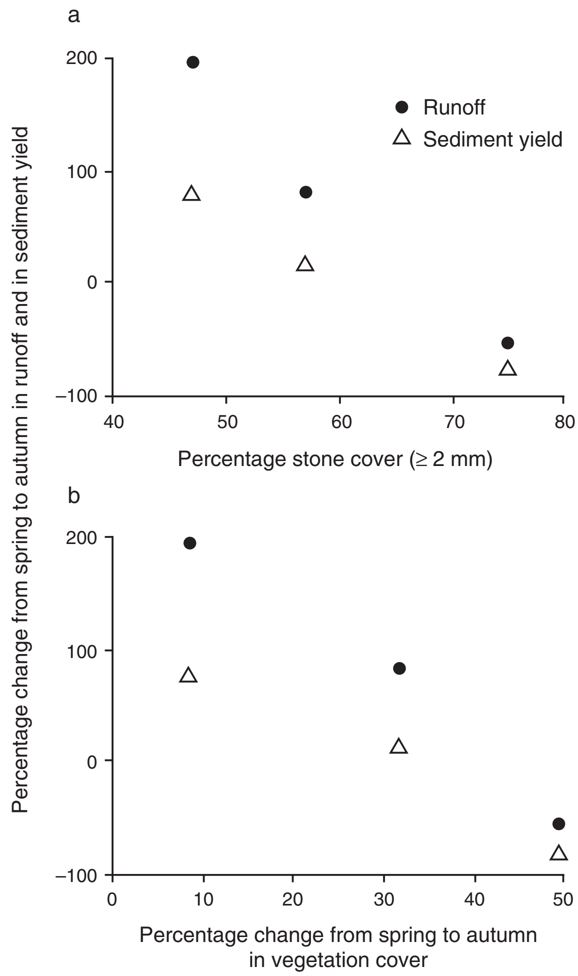

Fig. 9.21 Graphs of percentage change from spring to autumn in runoff and sediment yield against (**a**) percentage stone cover and (**b**) percentage change from spring to autumn in vegetation cover for three soils at Walnut Gulch, Arizona

The bulk of desert flora consists of ephemerals and annuals that germinate in response to rainfall events (Thomas 1988). This is very significant geomorphologically, as ephemerals typically appear 2–3 days after a rainfall event and annuals a few days later. Thus major rainfalls that trigger growth at the end of dry periods are erosionally very effective. Conversely, erosion rates at the end of wet periods, during which the plant canopy has thickened, are generally much lower than at other times. The plant canopy impedes soil erosion in a variety of ways, including protecting the ground surface from raindrop impact, which promotes infiltration and reduces soil detachment, and slowing overland flow. However, given that interrill erosion is governed by soil detachment rates (even if it is not detachment-limited) and that detachment on most desert hillslopes is accomplished mainly by raindrop impact, the principal mechanism whereby the plant canopy reduces erosion is probably through its influence on soil detachment.

Semi-arid ecosystems are dominated by either shrubs or grasses. Although semi-arid grasses are often clumped, they are more effective than shrubs as interceptors of rainfall (Thomas 1988). As a consequence, erosion rates at Walnut Gulch are two to three times greater for watersheds with predominantly shrub cover than for those with predominantly grass cover, even though runoff rates are similar (Kincaid et al. 1966). In general, erosion rates in semi-arid environments are inversely related to plant canopy or biomass. Kincaid et al. (1966) provide an example of such a relation for grass-covered watersheds at Walnut Gulch, whereas Johnson and Blackburn (1989) offer one for sagebrush-dominated sites in Idaho.

On some desert hillslopes, biological activity, in the way of digging and burrowing by animals or insects, plays a significant part in determining spatial and temporal variations in erosion rates. In a study conducted at the Sede Boqer experimental site, northern Negev, Israel, Yair and Lavee (1981) recorded intense digging and burrowing by porcupines and isopods (woodlice). Porcupines seeking bulbs for food break the soil crust which otherwise, due to its mechanical properties and cover of soil lichens and algae, inhibits soil erosion. Thus fine soil particles and loose aggregates are made available for transport by overland flow. Similarly, burrowing isopods produce small faeces which disintegrate easily under the impact of raindrops. Measurements of sediment produced by this biological activity on different plots revealed amounts that were of the same order of magnitude as eroded from the plots during a single rainy season (Fig. 9.22). Erosion rates were greater on the Shivta than on the Drorim and Netzer Formations (a) because of the proximity of biotic sediment to the measuring stations at the slope base, and (b) because of the higher magnitude and frequency of overland flow on the massive Shivta Formation. Yair and Lavee (1981) investigated the availability of biotic sediment across the northern Negev and found that it increased from 3 to 70 g m^{-1} as mean annual precipitation increased from 65 to 310 mm. These authors also

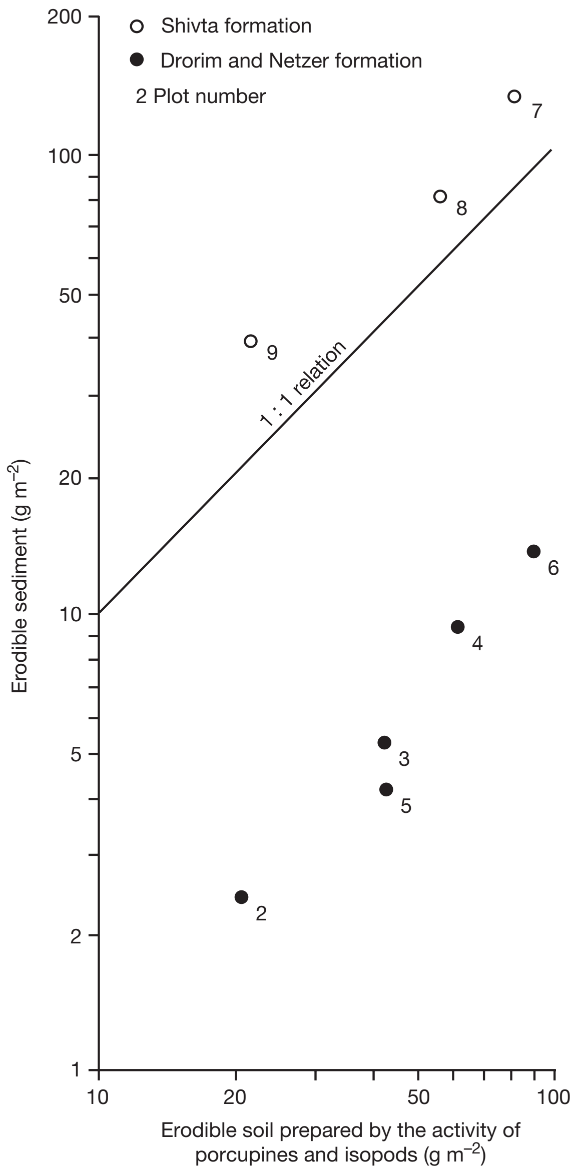

Fig. 9.22 Graph of eroded sediment against erodible soil prepared by porcupines and isopods for the Sede Boqer experimental site, northern Negev, Israel. Data from Yair and Shachak (1987, Table 10.4)

noted that biotic sediment may be produced in desert environments by a variety of animals and insects other than porcupines and isopods, including moles, prairie dogs, and ants (Table 9.1). In a study of the effects of small mammels on sediment yield in the Chihuahuan Desert, Neave and Abrahams (2001) showed that, under simulated rainfall, sediment concentration in runoff from small plots established on degraded grassland and in intershrub spaces correlated with the mean diameter

Table 9.1 Biological activity and sediment production in northern Negev, Israel (data from Yair and Lavee 1981)

Plot name	Annual rainfall (mm)	Sediment production ($g\,m^{-2}$)			
		Isopods	Porcupines	Moles	Total
Yattir	310	41.3	30.2	100	171.6
Dimona	110	4.5	3.5	0	8.0
Shivta	100	5.3	3.3	0	8.6
Sede Boqer	93	11.9	8.7	0	20.6
Mount Nafha	85	2.1	6.6	0	8.7
Tamar-Zafit	65	0	2.9	0	2.9

of animal disturbances. They argued that small mammels disturb surface crusts on soils and scatter sediment that is then available for entrainment into runoff.

Raindrop Detachment and Erosion

On desert hillslopes where the vegetation cover is generally sparse (Thomas 1988), the impact of rain drops is an important mechanism in the erosion process. Raindrop impact gives rise to rainsplash and rain dislodgement. Each of these processes will be discussed in turn.

Rainsplash occurs when raindrops strike the ground surface or a thin layer of water covering the ground and rebound carrying small particles of soil in the splash droplets. On a horizontal surface the mass of material splashed decreases exponentially with distance from the point of impact (Savat and Poesen 1981, Torri et al. 1987). The presence of a thin film of water appears to promote splash. Although Palmer (1963) reported that maximum splash occurs when the ratio of water depth to drop diameter is approximately 1, other workers have found that the maximum occurs at much smaller ratios (Ellison 1944, Mutchler and Larson 1971). Mass of soil splashed then decreases as the ratio increases (Poesen and Savat 1981, Park et al. 1982, Torri et al. 1987).

The most complete model currently available for predicting the net downslope splash transport rate for vertical rainfall has been proposed by Poesen (1985):

$$Q_{\mathrm{rs}} = \frac{E\cos\theta}{R\gamma_{\mathrm{b}}}[0.301\sin\theta + 0.019D_{50}^{-0.220}(1-\exp^{-2.42\sin\theta})] \qquad (9.9)$$

where Q_{rs} is net downslope splash transport rate ($m^3\,m^{-1}\,y^{-1}$), E is kinetic rainfall energy ($J\,m^{-2}\,y^{-1}$), R is resistance of the soil to splash ($J\,kg^{-1}$), γ_{b}, is bulk density of the soil ($kg\,m^{-3}$), θ is slope gradient (degrees of arc), and D_{50} is median grain size (mm).

This model indicates that Q_{rs} is positively related to rainfall kinetic energy corrected for surface gradient and negatively related to the bulk density of the soil and its resistance to splash. Resistance is, in turn, a function of D_{50} with a minimum at about 100 μm (Fig. 9.23). Coarser particles are more difficult to splash by virtue of their greater mass, while finer particles are more susceptible to compaction, are more cohesive, and promote the formation of a water layer that impedes splash

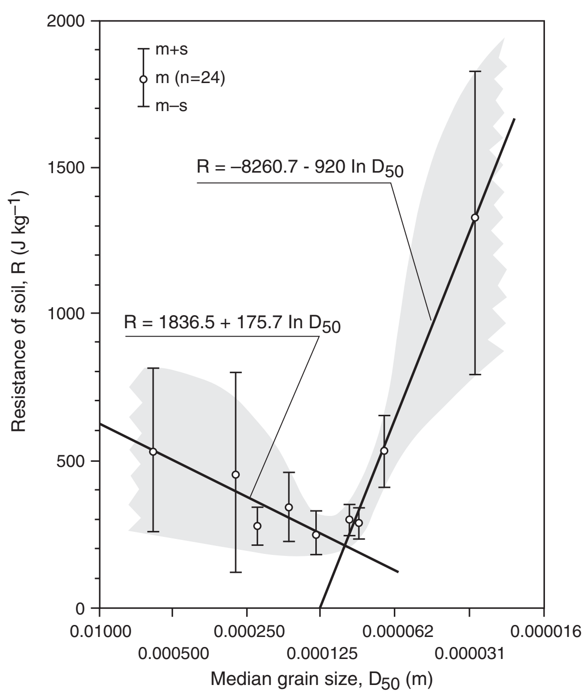

Fig. 9.23 Relation between the mean resistance of loose sediments to rainsplash and their median grain size. The standard deviation (s) of experimental values is shown with each mean (m)

(Poesen and Savat 1981). The first two terms inside the brackets respectively represent the effects of gradient and particle size on the mean splash distance, whereas the expression inside the parentheses reflects the influence of gradient on the difference between the volumes of soil splashed upslope and downslope. The model is based on laboratory experiments. However, Poesen (1986) assembled field data from a number of sources suggesting that it produces order-of-magnitude estimates of splash transport on bare slopes.

There have been few field studies of rainsplash in desert environments. Kirkby and Kirkby (1974) monitored painted stone lines near Tucson, Arizona, over a two-month period during the summer thunderstorm season. They found that mean travel distance due to rainsplash and unconcentrated overland flow increases with gradient and decreases with particle size (Fig. 9.24). By multiplying the travel distances in Fig. 9.25 by the grain diameters, they obtained the mass transport for each grain size per unit area. Then combining these data with data on storm frequency and assuming that rainsplash is completely suppressed under vegetation, Kirkby (1969) produced Fig. 9.25, which shows that erosion by rainsplash and unconcentrated overland flow reaches a maximum at annual precipitations of 300–400 mm. In other parts of the world, the maximum may occur at somewhat different precipitations, reflecting differences in the distribution of intense storms, seasonality of rainfall, and vegetation characteristics.

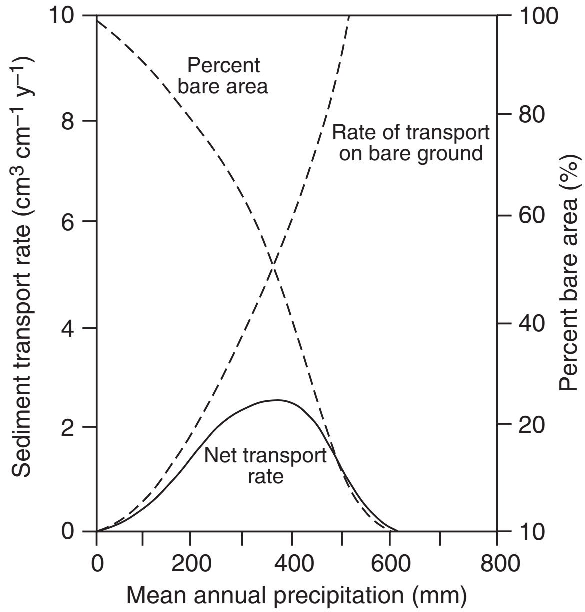

Fig. 9.25 Generalized relations of sediment transport rate, percentage bare area, and net transport rate (calculated as the product of the former two variables) to mean annual precipitation for the southern United States (after Kirkby 1969)

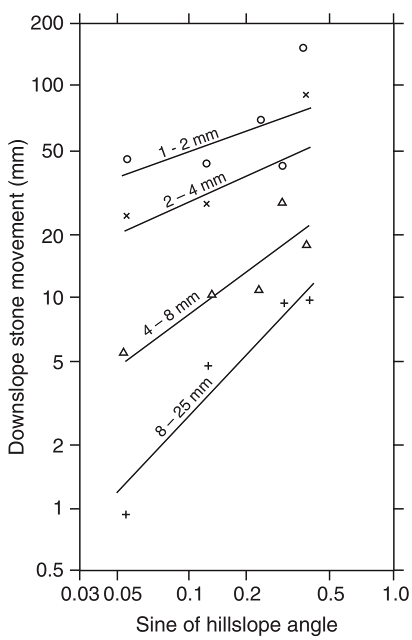

Fig. 9.24 Graph of downslope stone movement by rain-splash and unconcentrated overland flow against hillslope gradient for different stone sizes. Data were collected by Kirkby and Kirkby (1974) from painted stone lines on hillslopes near Tucson, Arizona

Kotarba (1980) monitored splash transport on two plots with gradients of 12° and 15° on the Mongolian steppe. During a single summer he found that length of transportation was closely related to rainfall intensity and grain diameter for particles coarser than 2.5 mm but weakly related to these variables for particles 2.0–2.5 mm in size. He attributed these weak correlations to the finer particles being transported by wind as well as splash. Kotarba pointed out that although the region has an annual precipitation of about 250 mm, 80% of which occurs as rainfall during June, July, and August, overland flow is confined to very limited areas, and rainsplash is the dominant erosional agent. Stones as large as 12 mm were moved by this process. The average displacement of stones during a single summer was 2–4 cm, with some stones travelling as far as 50 cm.

Martinez et al. (1979) measured rainsplash under simulated rainfall at six sites in southern Arizona. They found that mass of splashed material decreases as the proportion and size of stones in the surface pavement increase. Moreover, the presence of an undisturbed stone pavement seems to dampen the effect of increasing rainfall intensity, causing mass of splashed material to increase with rainfall intensity to the 0.48 power, whereas for bare agricultural fields the exponent is usually in the range of 1.5–2.5 (Meyer 1981, Watson and Laflen 1986). Finally, these authors noted that rainsplash is greatest for particles with diameters between 100 and 300 μm (i.e. fine sand), consistent with the laboratory findings of Poesen and Savat (1981).

Parsons et al. (1991b) pointed out that on many semi-arid hillslopes, shrubs are located atop small mounds of fine material, whereas the intervening intershrub areas are swales with a desert pavement surface. Applying simulated rainfall to seven shrubs at Walnut Gulch, Arizona, they showed that these mounds were formed largely by differential rainsplash – that is, to more sediment being splashed into the areas beneath shrubs than is splashed outward. Parsons et al. also demonstrated that both the splashed sediment and the sediment forming the mounds were richer in sand than the matrix soil in the intershrub areas, reflecting the tendency of rainsplash to preferentially transport sand-sized particles.

Rain dislodgement refers to the movement of soil particles by raindrops where the particles are not transported in splash droplets. Ghadiri and Payne (1988) showed that a large proportion of the splash corona (and hence detached sediment) fails to separate into droplets and falls back into the impact area. This proportion increases as the layer of water covering the surface becomes deeper. Moeyersons (1975) coined the term splash creep for the lateral movement of gravel by raindrop impact. He observed that stones as large as 20 mm could be moved in this manner, and using simulated rainfall he demonstrated that splash creep rate increases with gradient and rainfall intensity (Fig. 9.26).

Raindrop-detached sediment includes that which is dislodged by raindrops as well as that carried in splash droplets. In laboratory experiments, Schultz et al. (1985) found that the total weight of detached sediment was 14–20 times greater than that transported by splash. This finding underscores the fact that the most important role of raindrop impact is not in directly transporting sediment but in detaching soil particles from the surface prior to their removal by overland flow.

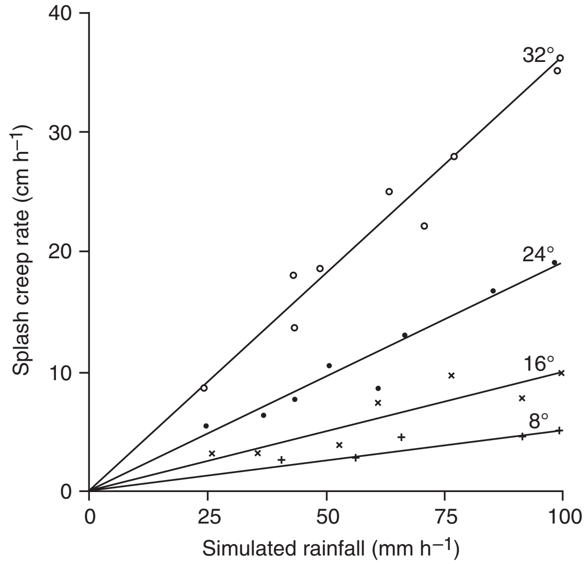

Fig. 9.26 Graph of splash creep rate measured in the laboratory against rainfall intensity for different slope gradients (after Moeyersons 1975)

Detachment of soil particles from the soil mass may be due to raindrop impact or flowing water. Numerous studies have established that on agricultural lands under unconcentrated runoff detachment occurs chiefly by raindrop impact (Borst and Woodburn 1942, Ellison 1945, Woodruff 1947, Young and Wiersma 1973, Lattanzi et al. 1974, Quansah 1985). To ascertain whether the same is true on undisturbed semi-arid hillslopes, five paired runoff plots, one covered with an insect screen to absorb the kinetic energy of the falling raindrops and the other uncovered, were established at Walnut Gulch, Arizona. Assuming that there is no raindrop detachment on the covered plots, the proportion of the sediment load detached by flowing water may be estimated by dividing the sediment load from the covered plot by that from the uncovered plot. As can be seen in Table 9.2, the proportions for all the plots are less than or equal to 0.25. These results support the proposition that raindrop impact is the dominant mode of detachment.

Work by Parsons and Abrahams (1992), however, has indicated that on a semi-arid hillslope at Walnut Gulch, the erosion rate in areas of unconcentrated runoff is less than the detachment capacity. Two types of evidence support this contention. First, particle size analyses reveal that splashed sediment is coarser than sediment being transported by overland flow, signify-

Table 9.2 Sediment yields for covered and uncovered runoff plots, Walnut Gulch, Arizona

Plot number	Status	Gradient degrees	Percentage vegetation	Percentage stones	Sediment yield, G ($gm^{-2}min^{-1}$)	G for covered plot / G for uncovered plot
1	Covered	7.7	38.1	43.8	6.0	0.19
1	Uncovered	7.5	44.8	33.3	31.9	
2	Covered	11.7	15.2	41.9	6.4	0.10
2	Uncovered	11.5	19.1	44.8	61.0	
3	Covered	16.0	27.6	48.6	1.2	0.086
3	Uncovered	17.5	18.1	55.2	13.4	
4	Covered	14.0	25.7	54.3	2.3	0.087
4	Uncovered	14.0	35.2	38.1	26.2	
5	Covered	17.0	18.1	48.6	8.9	0.25
5	Uncovered	17.7	34.3	44.8	36.3	

ing that the coarser detached particles are not eroded from the hillslope (Parsons et al. 1991a). Second, although raindrop detachment occurs over the entire hillslope except where overland flow is too deep, the detached sediment is transported downslope only where it is splashed into or dislodged within overland flow competent to transport it. Using a simulation model and detailed measurements of the cross-slope variation in overland flow depth and velocity, Abrahams et al. (1991) showed that there are significant proportions of the hillslope where soil is detached but there is no flow competent to transport it downslope. Inasmuch as overland flow on all desert hillslopes displays across-slope variations in depth and velocity, soil detachment rates probably always exceed actual sediment transport rates. Thus the notion that interrill erosion is detachment-limited appears to be an oversimplification. In reality the erosion rate will always be smaller than the detachment capacity. However, the magnitude of the disparity is difficult to estimate and is probably highly variable over both time and space.

Detachment and Erosion Under Concentrated Runoff

Although soil detachment in areas of unconcentrated runoff on desert hillslopes appears to occur mainly by raindrop impact, as flow paths become longer and threads of flow deeper, flow detachment may come to dominate in these threads both because critical flow shear stresses are exceeded and because the deeper water protects the soil surface from raindrop impact. This line of reasoning is supported by Roels' (1984a, b) findings on a rangeland hillslope in the Ardeche drainage basin, France. Roels observed that natural irregularities in the ground surface cause runoff to concentrate into interrill flow paths, which range in length up to 20 m, and he termed the longest of these flow paths prerills. He derived separate regression models for soil loss from the prerill and non-prerill interrill areas. In prerill sites 87% of the variation in soil loss was accounted for by a runoff erosivity factor $REF = Q \times Q_{\mathrm{p}}^{0.33}/A$, where Q is the runoff volume, Q_{p} the peak discharge, and A the drainage area. In contrast, in non-prerill interrill areas, 85% of the variation in soil loss was explained by a rainfall erosivity factor $EA \times IM$, where *EA* is the excess rainfall amount and *IM* is the maximum 5-minute rainfall intensity. These results imply that flow detachment dominates in the prerill portions of these areas, whereas raindrop detachment dominates elsewhere. Unless the prerills migrate laterally or are periodically infilled by other processes, flow detachment will inevitably cause them to evolve into rills.

Understanding how and when flow detachment becomes effective has, however, proved less than straightforward. In simple terms, it can be defined as when shear stress exerted by the flowing water exceeds the shear strength of the soil. Several authors have attempted to identify critical values for threshold mean shear stresses or shear velocities (e.g., Rauws and Govers, 1988; Slattery and Bryan, 1992), but Nearing (1994) has pointed out that mean shear stress exerted by shallow flow is of the order of a few pascals, whereas mean shear strength of soils is typically measured in the order of kilopascals. Nearing proposed that the solution to this apparent conundrum lay in the overlapping distributions of the two values (Nearing 1991). Based on this approach, Parsons and Wainwright (2006) undertook an analysis of the probability of incision on shrubland and grassland hillslopes at Walnut Gulch, southern Arizona (Fig. 9.27). They demonstrated that probabilities for incision on both shrubland and grassland were similar for similar discharges and similar degrees of soil moisture, and that the explanation for incision on shrubland hillslopes but not on grassland that is apparent at Walnut Gulch lies in the more frequent higher discharges on the shrubland compared to the grassland. These more frequent higher discharges on the shrubland compared to the grassland are a result of soil and microtopographic differences between the two vegetation communities (Parsons et al., 1996).

Gravitational Proceses

The movement of weathered detritus under the influence of gravity encompasses a very wide range of phenomena differing in depth and mass of material being mobilized, rates of motion, transport mechanisms, and relative volumes of debris, water, ice and air. Only a limited number of these phenomena are common on rock-mantled hilllslopes in arid environments. The most important of these are debris flows, but a limited amount of movement of dry debris also occurs.

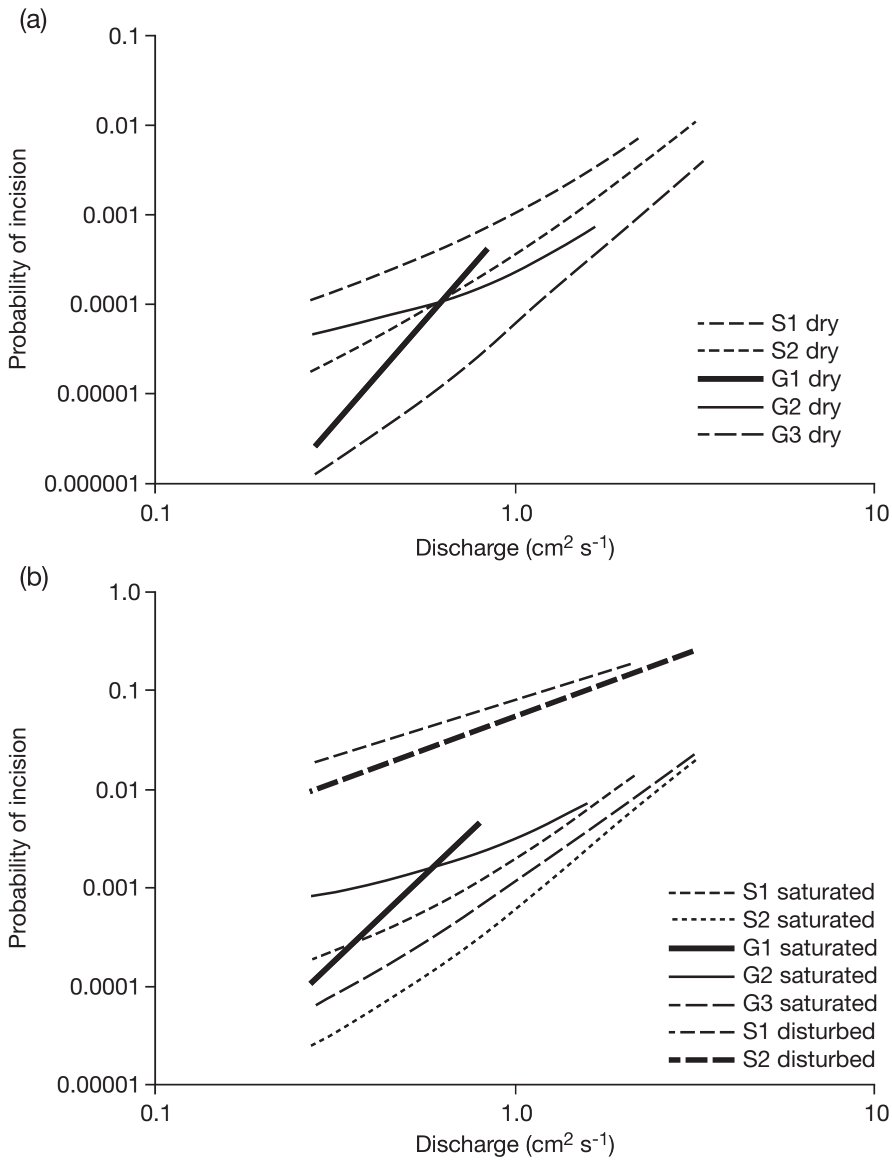

Fig. 9.27 Probabilities of incision at cross section on grassland and shrubland plots at Walnut Gulch, Arizona under dry, saturated and disturbed soil conditions (after Parsons and Wainwright 2006). See also Figs. 9.16 and 9.17

Movement of Dry Debris

The movement of dry debris as particle-by-particle sliding under gravity is known as dry ravel and has been claimed to be a dominant mechanism of sediment transport on steep hillslopes in arid and semi-arid environments (Gabet 2003). Inasmuch as most coarse particles on debris slopes are weathered in-situ from the underlying bedrock the question of the triggering mechanism to make previously stable particles become unstable naturally arises. The most commonly cited cause for the initiation of dry ravel on debris slopes is fire (e.g. Florsheim et al. 1991, Cannon et al. 1998, Roering and Gerber 2005). The destruction of vegetation removes the support for fine material that often accumulates behind plants and it is argued that this initiates instability in the coarse fragments.

Debris Flows

The commonest phenomena due to gravitational processes on rock-mantled hillslopes, certainly in terms of amount of sediment moved, are debris flows. A full discussion of the mobilization of debris flows is beyond the scope of this chapter, which will focus on conditions on desert rock-manted hillslopes that influence this mobilization. For further discussion of debris-flow processes, the reader is referred to the review by Iverson et al. (1997). Debris flows generally exhibit a consistent range of behaviour (Blackwelder 1928, Johnson 1970, Fisher 1971, Costa 1984, Johnson and Rodine 1984). The flows occur as a series of blunt-nosed pulses with the first pulse commonly being the largest. The maximum depth of each pulse occurs near the nose, with a long tailing flow that is commonly more fluid than the nose, often changing to hyper-concentrated or water flood flows during the waning stages. What characterizes debris flows from either of these types of flow is the synergistic transfer of momentum by both solids and fluids (Iverson et al. 1997). Debris flows are noted for the wide range of grain sizes transported and the tendency for large boulders to be concentrated near the flow surface and at the leading edge of the flow. Deposits from debris flows often show inverse grading. Characteristics of debris flow deposits are considered in Chapter 14. The general rule for debris flows is that failure will occur in the layer of weathered material on rock-mantled hillslopes if stresses obey Coulomb's (1773) rule

$$\tau > \sigma' \tan \varphi + c \qquad (9.10)$$

where τ is the mean shear stress acting on the failure surface, σ' is the mean effective normal stress acting on the failure surface, φ is the angle of internal friction of the weathered layer, and c is the cohesion of the weathered layer. Cohesion mainly depends upon electrostatic forces between clay particles and secondary mineralisation in the weathered layer. The latter can be significant, especially where petrocalcic horizons have developed, but, in the absence of such horizons, many rock mantles are effectively almost cohesionless because of their low clay content. The angle φ is a function of the weathered layer, and hence the weathering characteristics of the underlying lithology. Consequently, the frequency of debris flows on these hillslopes is strongly dependent on the underlying lithology. Effective normal stress accounts for pore-fluid pressure p such that

$$\sigma' = \sigma + p \qquad (9.11)$$

where σ is the total normal stress.

Although it cannot be assumed that ϕ and c are fixed quantities for a given weathered mantle (see Iverson et al. 1997), the key requirement for mobilization of debris flows is sufficient water to result in high pore pressures. On desert hillslopes this requirement is almost certain to be met as a result of rainfall; groundwater inflow is unlikely. However, as Iverson et al. (1997, p.101) point out, mobilization of debris flows by rainfall on steep hillslopes 'presents a mechanical difficulty'. How do the slopes remain stable for long enough to become nearly saturated? The solution proposed by Iverson et al. is to suggest that prolonged rainfall at an intensity greater than the saturated hydraulic conductivity of the mantle layer will create a saturated zone at the ground surface that will propagate downward. This layer may remain tension-saturated after rainfall ceases so that a subsequent burst of high-intensity rainfall may cause positive pore pressure to develop almost instantaneously. Although heavy rainfall is widely associated with reports of debris flows on desert hillslopes (e.g. Coe et al. 1997, Cannon et al. 2001) data to support this hypothesized mechanism are lacking. An alternative supply of large amounts of water may be from upslope where percentages of bedrock may be higher and infiltration rates lower. Such differences in infiltration between uplsope and downslope portions of desert hillslopes are not uncommon (Yair 1987). An alternative triggering mechanism is proposed by Blijenberg et al. (1996) who postulate that microscale mass movements that occur in response to rainfall events could play a significant role in triggering mass movements.

Three other factors may contribute to the occurrence of debris flows on rock-mantled hillslopes. First, the sparse vegetation of desert environments limits the cohesive strength that is imparted to soil layers by plant roots. However, although the sparseness of vegetation makes rock-manted hillslopes inherently unstable, it also removes one of the triggers to debris-flow mobilization, namely the sudden loss of strength as cohesion is suddenly and dramatically lost when roots are broken. Secondly, marked differences in infiltration capacity often exist on debris slopes, particularly where petrocalcic horizons exist. The existence of such hoizons may provide temporary perched water tables during storm events. Finally, debris flows in deserts commonly occur shortly after fire has destroyed the natural vegetation cover (Sidle et al. 1985, Wohl and Pearthree 1991, Cannon et al. 1998). Fire may contribute to failure by reducing evapotranspiration while creating a hydrophobic fire-sealed soil layer that promotes surface soil saturation and runoff (Wells 1981, 1987, Laird and Harvey 1986, Campbell et al. 1987, Wells et al. 1987). However, Florsheim et al. (1991) noted that wildfire often is not followed by large debris flows even though sediment yield is increased, and they suggest that rainfall intensity and duration is much more important in triggering debris flow than is wildfire.

The mathematical analysis of debris-flow mobilization is typically undertaken using the infinite-slope model (see Iverson et al. 1997, Fig. 9). The insights that the model provide are important because they indicate the relationships that exist among the forces controlling the mobilization of debris flows. However, this model, as Iverson et al. point out, is not amenable to testing because it makes predictions that cannot be applied to any naturally occurring hillslope. Consequently, many gaps still remain in our understanding of the conditions under which debris flows are triggered on rock-mantled slopes.

Climate Change

A discussion of rock-mantled slopes would be incomplete if it did not consider the effects of climatic change. Major climatic fluctuations have probably occurred in every desert during the Cenozoic (Chapter 28) and have strongly influenced the form of many debris slopes (Chapter 22). The imprint of these former climates appears to be most pronounced where rock resistance is greatest. This is well illustrated by Oberlander's (1972) classic study of boulder-covered slopes on resistant quartz monzonite in the Mojave Desert, California. These slopes consist of a 'jumble of subangular to spheroidal boulders of a variety of shapes and sizes clearly derived from plane-faced blocks bounded by intersecting joints' (Oberlander 1972, p. 4) (Fig. 9.6). Oberlander argued that the boulders formed as corestones within a deep weathering profile under a wetter climate, and that these corestones became stranded on bedrock slopes as the supporting matrices of fines were removed under the more arid climate of the late Tertiary and the Quaternary. Not surprisingly, Oberlander could find no correlation between hillslope gradient and boulder size. Other investigators too have reported an absence of any relation between gradient and debris size on slopes underlain by resistant rocks (e.g. Melton 1965, Cooke and Reeves 1972, Kesel 1977), suggesting that these slopes also owe much of their form and sedimentology to climatic change.

Legacies from past climates are probably more prevalent on rock-manted hillslopes than is generally realized. Certainly, Oberlander's description of boulder-clad slopes in the Mojave Desert applies to similar hillslopes in most granitic terranes. Erosion on such slopes is often characterized as weathering-limited (e.g. Young 1972, p. 206, Mabbutt 1977, p. 41). However, if this were wholly the case, such slopes would be more or less devoid of fine material because, by definition, such material should be removed as rapidly as it is produced. Instead, what we often find are boulders or bedrock outcrops protruding from a matrix of fines that becomes progressively more extensive downslope. Parsons and Abrahams (1987) investigated this phenomenon in the Mojave Desert and concluded that the presence of the fines indicates adjustment by the hillslope to extant hydraulic processes, and that the degree of

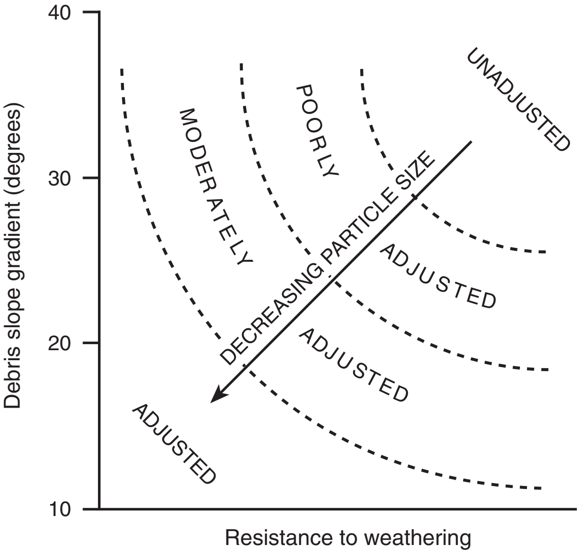

Fig. 9.28 Graph of debris slope gradient against resistance to weathering, showing how the degree of debris slope adjustment varies with these variables (after Parsons and Abrahams 1987)

adjustment is inversely related to slope gradient and rock resistance (Fig. 9.28). It is interesting to note that inasmuch as particle size decreases as degree of adjustment increases, the hillslopes studied by Parsons and Abrahams display strong correlations between gradient and particle size, even though they are far from being adjusted to contemporary hydraulic processes, especially in their steeper parts.

Conclusion

As with most of the landforms and processes in deserts, our knowledge of rock-mantled slopes is deficient and there are many opportunities for further field observations and experimentation, laboratory investigations, and theoretical modelling. Our understanding of weathering processes of regolith generation is poor, and the relative roles of present and past climates is uncertain. Mechanisms of debris mobilization, deposition, and further weathering are reasonably well understood in a qualitative sense, but the long-term interaction of processes and materials to create specific types of desert slopes is poorly characterized.

The deficiency in our understanding of rock-mantled hillslopes is important because the majority of desert hillslopes fall into this category. Understanding the form and processes of these hillslopes is, therefore, important for understanding a significant component of desert landscapes. Furthermore, they give rise to a variety of hydrologic and geomorphic phenomena such as flash flooding, extreme soil erosion, and hazardous debris flows. More generally, they exert a major control over the flux of water and sediment that passes through desert river systems, across active piedmonts, and into closed lake basins. Thus an understanding of many, if not most, desert geomorphic systems must begin with a comprehension of the processes operating on these desert hillslopes.

Acknowledgments The experiments reported in Table 9.2 were funded by NSF grant SES-8812587.

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