ROCK-MANTLED SLOPES 8

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INTRODUCTION

This chapter will consider hillslopes with gradients less than 40° that are neither active, undissected piedmonts (i.e. pediments and alluvial fans, which are dealt with in Chapters 13 and 14) nor are developed in highly erodible, fine-grained sedimentary rocks (i.e. badlands slopes, which are treated in Chapter 9). Such hillslopes in desert (arid and semi-arid) climates are usually mantled with a thin layer of coarse debris that has either weathered out of the underlying substrate or fallen from a superincumbent cliff. The former hillslopes are herein referred to as debris slopes and the latter as talus slopes. In this chapter we shall first consider the processes acting on these rock-mantled slopes and then consider the forms they produce. These processes are not only important in terms of understanding hillslope form but because 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 desert hillslopes.

HYDRAULIC PROCESSES

INFILTRATION AND RUNOFF

Model

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 Horton overland flow (Horton 1933). The infiltration process may be modelled by the Richards (1931) equation:

$$\frac{\partial\Theta}{\partial t} = \frac{\partial}{\partial z}\left[K_s k_r(\Psi)\frac{\partial\Psi}{\partial z}\right] - \frac{\partial}{\partial z}\left[K_s k_r(\Psi)\right] \quad (8.1)$$

where Θ is the volumetric water content, Ψ is the pressure head, z is the distance (positive) below the surface, t is time, K_s is the saturated hydraulic conductivity, and $k_r(\Psi)$ is the relative hydraulic conductivity as a function of Ψ (Smith and Woolhiser 1971). Smith (1972) and Mein and Larson (1973) developed solutions to this equation which enable infiltration curves to be predicted for particular soils for different rainfall rates and initial soil-water contents (Fig. 8.1). The Richards equation is especially useful where the rainfall intensity initially does not exceed the surface infiltration rate.

Although the Richards equation is a powerful analytical tool, its application to desert hillslopes is complicated by the widespread occurrence of surface sealing (Moore 1981, Romkens *et al.* 1990). Surface sealing may be caused by swelling of soil aggregates or soil structure breakdown due to physico-chemical processes, clogging of soil pores as a result of the deposition of fine particles in ponded water, or aggregate breakdown and compaction by raindrops (Romkens *et al.* 1990). The last mechanism is particularly important on desert hillslopes where the effectiveness of raindrops in sealing the surface appears to be contingent on the extent to which the surface is covered with stones and vegetation. Simanton and Renard (1982) conducted rainfall simulation experiments on three different soils on semi-arid hillslopes at Walnut Gulch, Arizona, and compared runoff rates in the spring and autumn. In the spring the soil surface was loosened by frost action and wetting and drying during the preceding winter, whereas in the autumn the surface was compacted by the

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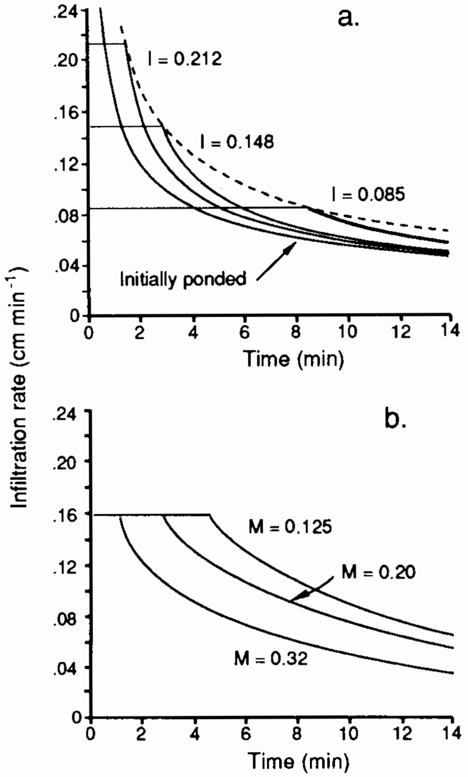

Figure 8.1 Graphs of infiltration rate against time since the start of rainfall obtained by solving the Richards equation. These graphs show how the infiltration curve varies with (a) rainfall rate I and (b) initial moisture content M (volume per unit volume).

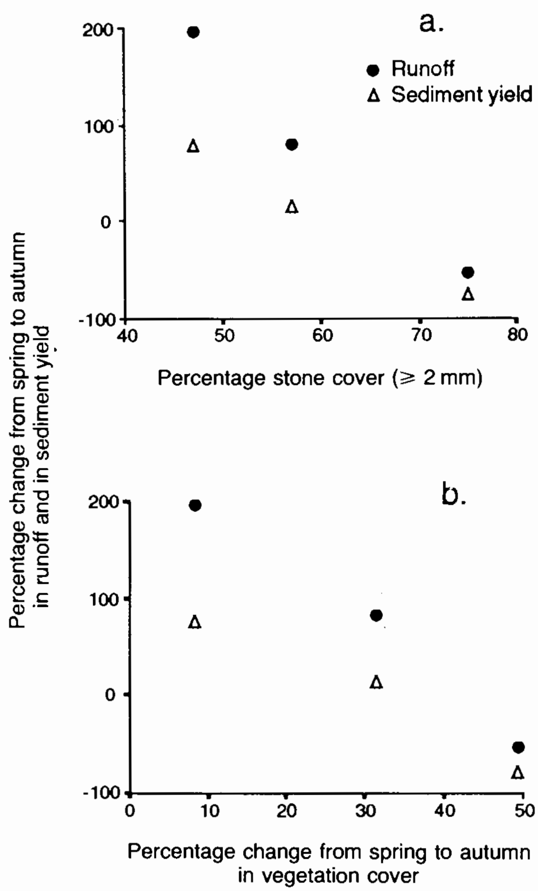

Figure 8.2 Graphs of percentage change from spring to autumn in runoff and sediment yield against (a) percentage stone cover and (b) percentage change from spring to autumn in vegetation cover for three soils at Walnut Gulch, Arizona.

previous summer's thunderstorm rainfall. Figure 8.2 shows that the change in runoff from spring to autumn is inversely related to the proportion of the surface covered with stones (≥2 mm) and the percentage change in vegetation from spring to autumn. This is because higher stone and vegetation covers afford the soil surface greater protection from raindrop impact and, hence, inhibit surface sealing by summer rainfall.

Controlling Factors

Infiltration rates on desert hillslopes vary greatly over short distances (e.g. Blackburn 1975, Sharma *et al.* 1980, Yair and Lavee 1985, Berndtsson and Larson 1987, Johnson and Gordon 1988). This variation may reflect variation in surface or subsurface properties. However, much less is known about the role of the latter than the former. Among the surface properties controlling infiltration and runoff are the ratio of bedrock to soil, surface and subsurface stone size, stone cover, and vegetation.

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 8.3 shows the infiltration curves for rocky and soil-covered surfaces at Sede Boquer 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 Boquer 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 Boquer

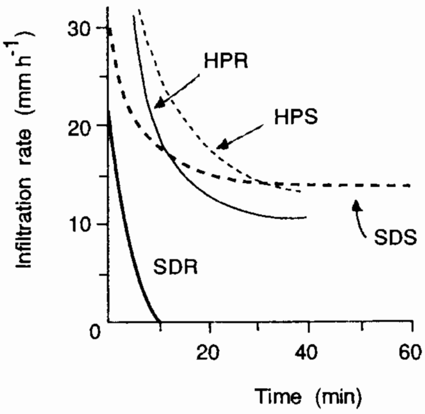

Figure 8.3 Infiltration curves for rocky and soil-covered surfaces in the northern Negev, Israel: Sede Boquer (rainfall intensity $36\,\mathrm{mm\,h^{-1}}$) massive limestone (SDR), Sede Boquer stony colluvium soil (SDS), Hovav Plateau (rain intensity $33\,\mathrm{mm\,h^{-1}}$) densely jointed and chalky limestone (HPR), and Hovav Plateau stoneless colluvial soil (HPS) (after Yair 1987).

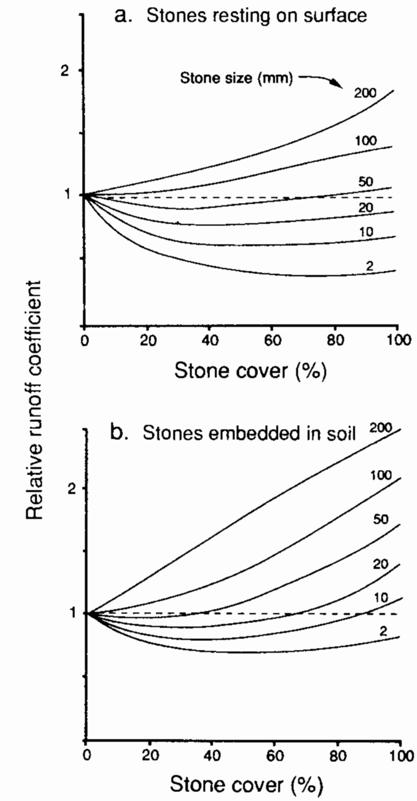

Figure 8.4 Graphs showing relations between runoff coefficient and stone cover for different stone sizes: (a) stones resting on the soil surface and (b) stones partially embedded in the soil. The graphs are generalizations of the experimental results of Poesen and Lavee (1991).

(Yair 1983) indicate that the threshold level of daily rainfall necessary to generate runoff in the rocky areas is 1 to 3 mm, whereas it is 3 to 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.

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 1991b, Lavee and Poesen 1991, Poesen and Lavee 1991). Figure 8.4, which is based on laboratory experiments by Poesen and Lavee (1991, fig. 3), is an attempt to summarize the present 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 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 Figures 8.4a and 8.4b 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 to 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 8.4 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 (1991a) noted that both positive and negative correlations between infiltration and stone cover have been reported for semi-arid hillslopes in the American South-west. They observed that positive correlations (Tromble 1976, Abrahams and Parsons 1991b) have been obtained when infiltration measurements were confined to (bare) stone-covered areas between shrubs (lower curves in Fig. 8.4), 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 1991a). 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 Figure 8.5. 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).

Finally, an oft-neglected factor influencing infiltration and runoff on desert hillslopes is the presence of cryptogamic crusts. Such crusts may be composed of algae, lichen, or moss. They are surprisingly common on desert surfaces and help to stabilize such surfaces. They also have a water-repellent property that results in their generating runoff during relatively light showers of short duration. For example, in the Nizzana sand field, southern Israel, Yair (1990a) found that less than 1 mm of rainfall was sufficient to initiate runoff at a relatively low rainfall intensity of 18.4 mm h^{-1}. He concluded that some runoff could be expected over the crusted surface during any rainstorm exceeding 2 to 3 mm.

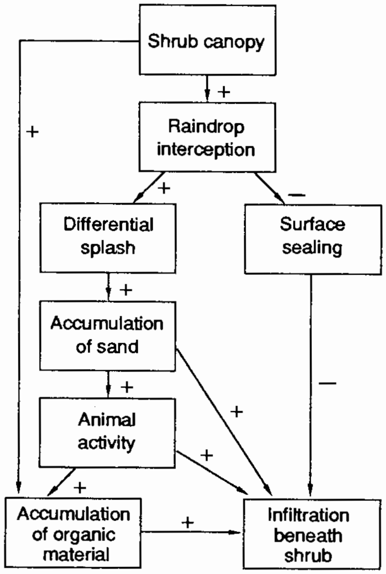

Figure 8.5 Causal diagram showing the mechanisms whereby a shrub's canopy promotes infiltration under the shrub at Walnut Gulch, Arizona (after Abrahams and Parsons 1991a).

Partial Areas

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, termed partial areas, where the infiltration rates are lowest. Overland flow generated by partial areas 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 downslope. This phenomenon is illustrated in Figure 8.6 for a small piedmont watershed at Walnut Gulch, Arizona (Kincaid *et al.* 1966).

Another example of runon infiltration has been documented at the Sede Boquer experimental site studied by Yair and his colleagues in the northern Negev, Israel. The upper part of the hillslope consists largely of exposed bedrock, whereas the lower part is covered by stony colluvium. A series of sprinkling experiments showed that runoff generated on the upper part of the hillslope usually infiltrates into the colluvium. Only when both the antecedent moisture and rainfall intensity are exceptionally high, does overland flow cross the colluvial footslope and reach the stream channel (Fig. 8.7). Runoff data for natural rainfall events collected over a ten-year period revealed that continuous flow down the 60-m hillslope occurs in only about 5% of the rainstorms, thus representing rather extreme conditions (Yair and Lavee 1985).

Yair and Shachak (1987) suggested that desert hillslopes are often divided into a rocky upper slope and a soil-covered lower slope. Certainly, in the

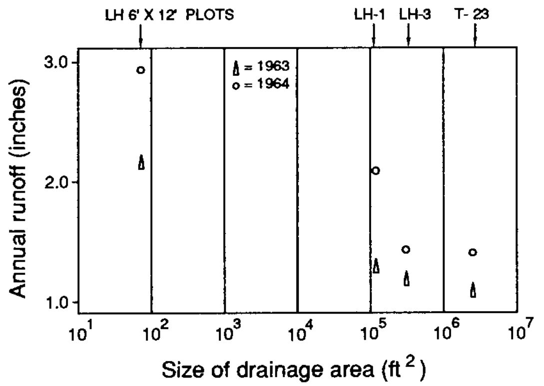

Figure 8.6 Graph of annual runoff against size of drainage area for runoff plots and very small watersheds ($>10^5$ ft^2) at Walnut Gulch, Arizona (after Kincaid *et al.* 1966).

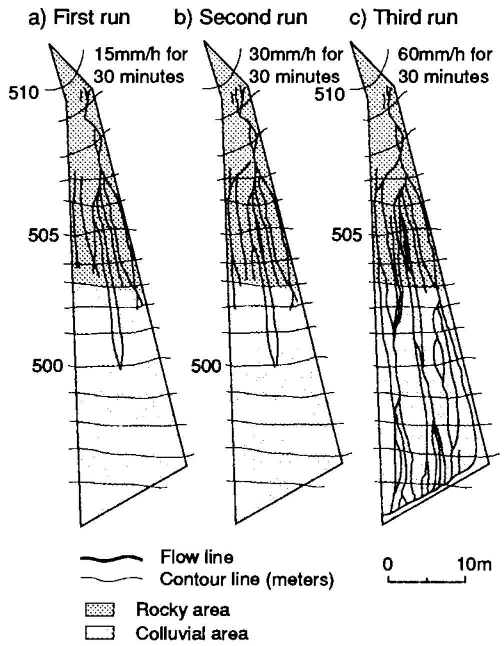

Figure 8.7 Flowlines during three consecutive rainfall simulation experiments on a runoff plot at Sede Boquer, northern Negev, Israel (after Yair 1990b).

Mojave Desert, California, there is a tendency for bedrock to crop out more widely on the upper part of hillslopes (i.e. the upper convexity and rectilinear elements) than on the lower part (i.e. the lower concavity) (Parsons and Abrahams 1987). Inasmuch as this pattern of a rocky upper slope and colluvial lower slope appears to be common and runoff coefficients for bedrock are generally higher than those for colluvium, runon infiltration is likely to be an important phenomenon on desert hillslopes and must be taken into account in modelling and predicting hillslope hydrographs.

OVERLAND FLOW HYDRAULICS

On desert hillslopes virtually all runoff occurs in the form of overland flow. Although there is pedogenic evidence of throughflow on some hillslopes (Wieder *et al.* 1985, Lavee *et al.* 1989), this process appears to contribute little water to stream channels and to play a minor role in hillslope erosion. Consequently, the process of throughflow will not be discussed further.

Interrill Areas

Interrill overland flow on desert hillslopes generally appears as a 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. Resistance to overland flow may be quantified by the dimensionless Darcy–Weisbach friction factor

$$f = 8ghS/V^2 \tag{8.2}$$

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, and rain 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). Finally, 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). 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), 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

$$R_e = 4Vh/\nu \qquad (8.3)$$

where ν 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. Virtually all models of hillslope runoff have employed this relation between f and R_e (or surrogates thereof) for plane beds. 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-upward (Fig. 8.8). 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.

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

$$\log f = -5.960 - 0.306 \log R_e + 3.481 \log \%G + 0.998 \log D_g \qquad (8.4)$$

where $\%G$ is the percentage of the surface covered

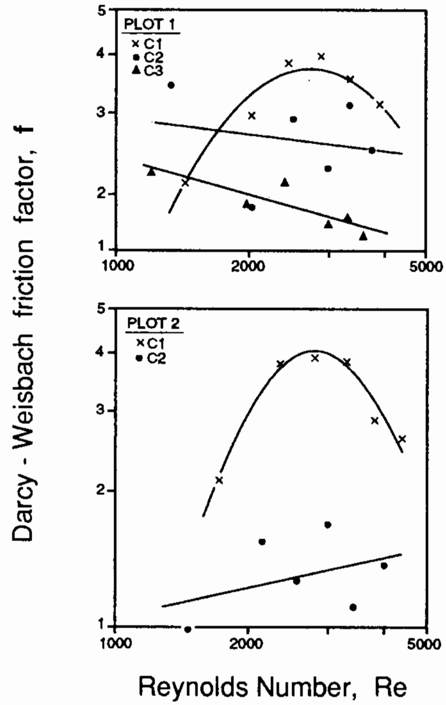

Figure 8.8 Graphs of Darcy–Weisbach friction factor against Reynolds Number for five cross-sections on two runoff plots at Walnut Gulch, Arizona. The cross-sections are denoted by C1, C2, etc. (after Abrahams *et al.* 1986).

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. 8.9). 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.

The findings of Abrahams *et al.* (1986) and Abrahams and Parsons (1991c) are supported by some recent 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

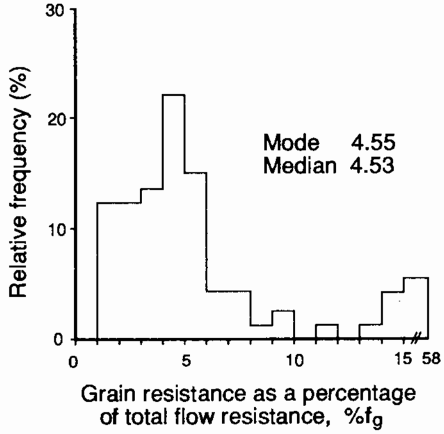

Figure 8.9 Relative frequency distribution of grain resistance expressed as a percentage of total flow resistance for 73 experiments on 8 runoff plots at Walnut Gulch, Arizona.

inundation of the gravel. In addition, they obtained regression equations of the form

$$\log f = \log a_1 - a_2 \log R_e + a_3 \log \%G \qquad (8.5)$$

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 hydraulics of overland flow over an entire interrill area were investigated by Parsons *et al.* (1990) using simulated rainfall on a plot 18 m wide and 35 m long also located on a shrub-covered piedmont hillslope at Walnut Gulch. These shrubs sit atop low mounds formed largely by differential splash (see Rainsplash below). Between the shrubs are gravel-covered surfaces where overland flow is generated and travels downslope. The h, V, and R_e values were computed for two measured sections situated 12.5 and 21 m from the top of the plot. At-a-section h–R_e and V-R_e relations show that increases in R_e are accommodated largely by increases in h. These hydraulic relations are principally the result of f increasing with R_e as overland flow spreads laterally into new areas. Downslope hydraulic relations differ strikingly from at-a-section relations. Under equilibrium (steady state) runoff conditions, f decreases rapidly as R_e increases, permitting increases in R_e to be accommodated 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.

Rills

The tendency for threads of 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 or rills. Such features are very common on desert hillslopes, particularly where the underlying material is easily eroded. However, to our knowledge, there have been no quantitative studies of either the form or hydraulics of rills on such hillslopes. All the data available on rills come from agricultural fields in more humid environments. The following discussion, therefore, is based on these data on the assumption that the mechanics of rill formation are similar in desert and humid settings and that, consequently, the findings for agricultural fields can with appropriate caution be generalized to desert hillslopes.

Studies of rill morphology suggest that rill width is a function of soil erodibility, whereas rill spacing is related to hillslope gradient (Parsons 1987) (Fig. 8.10). Rill depth usually reflects the degree of rill development or the depth to a resistant layer in the soil. Small rills may exist for only short intervals before being obliterated by other slope processes, such as frost action, soil creep, or hydraulic deposition. Large rills, on the other hand, may persist for decades. Rills grade into gullies. The boundary is necessarily arbitrary, but one that has been widely adopted specifies that gullies are wider than 0.33 m and deeper than 0.67 m (Brice 1966). Gully processes and forms are discussed in Chapter 12.

Because rills are irregular in profile and cross-section, rill flow varies markedly in depth and velocity over short distances. The flow may be laminar, transitional, or turbulent (Roels 1984c, Gilley *et al.* 1990, Sadeghian and Mitchell 1990), but it is more often turbulent than is interrill flow. Field and laboratory studies of rill flow reveal that f–R_e relations are consistently inverse with slopes ranging from 0 to −1.85 and averaging about −1.0 (Roels 1984c, Foster *et al.* 1984a, Gilley *et al.* 1990, Sadeghian and Mitchell 1990). These slopes reflect the dominant influence of rill form on f. Specifically, as R_e increases f decreases because any roughness elements on the bed are covered by progressively greater depths of water and because the flow better follows the waviness of the bed (Foster *et al.* 1984a, Abrahams *et al.* 1986). In a study of flow through a

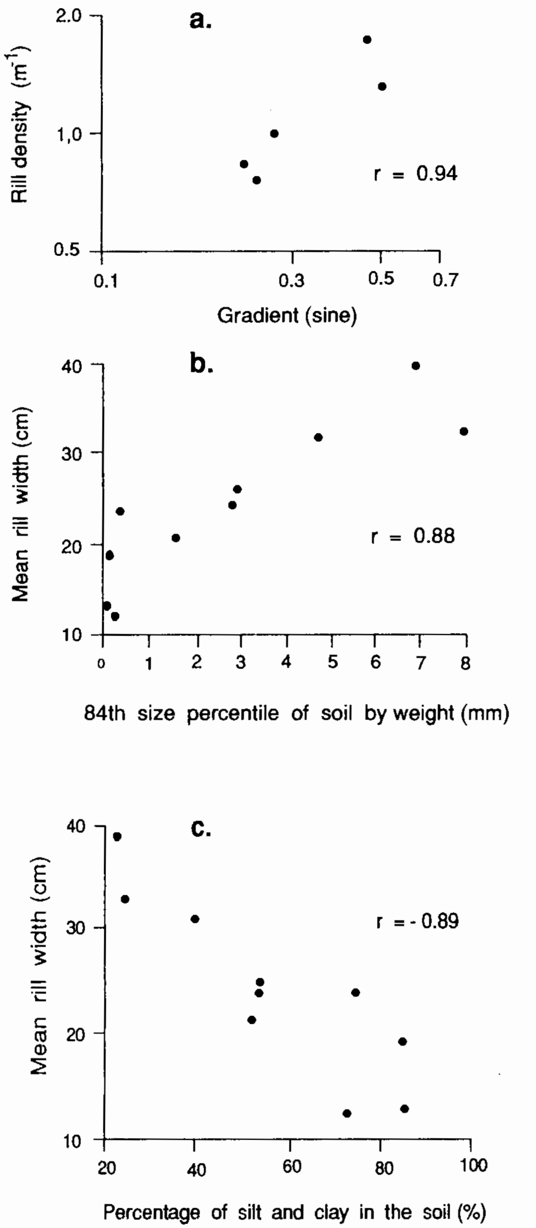

Figure 8.10 Graphs of (a) rill density against hillslope gradient, (b) mean rill width against the 84th size percentile of the soil, and (c) mean rill width against percentage of silt and clay in the soil.

fibreglass replica of a rill, Foster *et al.* (1984b) found that $\% f_g$ increased from 8.6 to 38.5% as R_e increased from 16 000 to 80 000. As in interrill flow, form resistance dominates, but it becomes less dominant as R_e increases.

Using hot-film anemometry, Foster *et al.* (1984a) also investigated velocity profiles in their rill replica. They found that the profiles were well described by the Prandtl–von Karman velocity distribution law:

$$v = (v_*/\kappa) \ln(z/z_0) \qquad (8.6)$$

where v is point velocity, κ is von Karman's constand, v_* is shear velocity, z is height above the bed, and z_0 is roughness length. However, κ, which is normally assumed to equal 0.4, took values between 0.2 and 1.3. In general, κ varied with position along the rill and increased with discharge. Because rill flows are deeper and more turbulent than interrill flows, rainfall has less effect on flow velocity. Foster *et al.* (1984a) showed that mean flow velocities were 2 to 10% less with a rainfall intensity of 127 mm h^{-1} than with no rain.

To our knowledge, the hydraulic geometry of rills on desert hillslopes has never been investigated. Studies have been made by agricultural engineers of the hydraulic geometry of rills on croplands, but with one exception these studies have examined rills developed on preformed (furrowed) surfaces. Consequently, their hydraulic geometry may not be representative of natural rills. The exception is a recent study by Gilley *et al.* (1990), who analysed the at-a-section relation between rill width and discharge for ten soils scattered across the eastern and mid-western United States. The exponent of the power relation ranged from 0.144 to 0.467 and averaged 0.303. This average exponent is comparable to that for river channels (Ferguson 1986), suggesting that rills and river channels are hydraulically similar (Moore and Foster 1990). However, whereas the hydraulic geometry of individual rivers fluctuates about some equilibrium, that of individual rills is in a state of constant evolution. Moreover, although average exponents may be similar, there is considerable variation about these averages especially in response to differences in bank strength (Ferguson 1986). Thus, caution is recommended in drawing conclusions from the above similarity.

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.

Controlling Factors

Abrahams and Parsons (1991b) investigated the relation between hydraulic erosion and gradient at Walnut 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. 8.11). 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 Figure 8.12 (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.

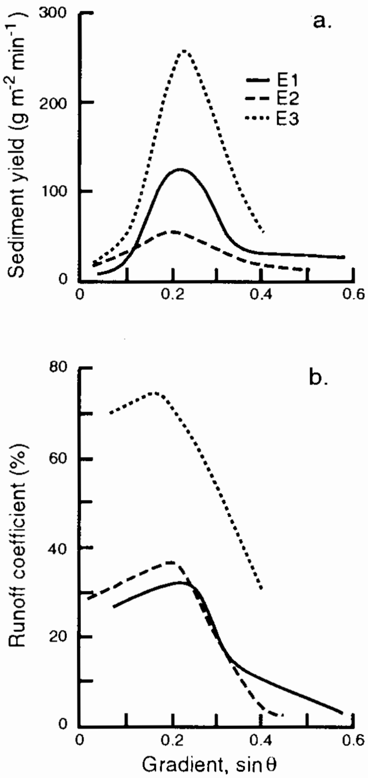

Figure 8.11 Curves fitted to graphs of (a) sediment yield and (b) runoff coefficient against gradient for three sets of experiments denoted by E1, E2, and E3 at Walnut Gulch, Arizona. Experiments E1 and E2 were conducted on plots underlain by Quaternary alluvium, with the ground vegetation being clipped for E1 but not for E2. Experiment E3 was performed on plots underlain by the Bisbee Formation.

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 4° to 25°, which contributed greatly to the scatter. In a better controlled study in which all the plots had similar gradients (5.1° to

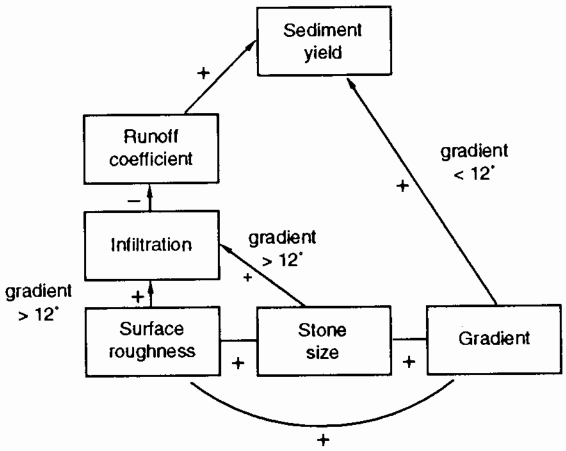

Figure 8.12 Causal diagram showing the factors controlling the runoff coefficient and sediment yield on desert hillslopes.

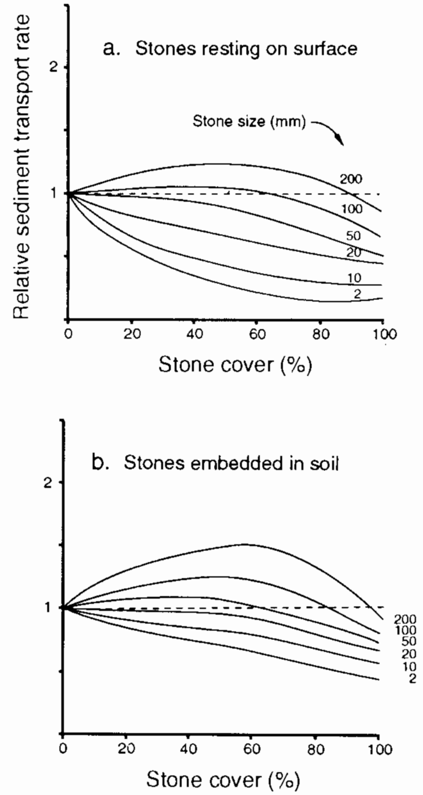

Figure 8.13 Graphs showing relations between sediment transport rate and stone cover for different stone sizes: (a) stones resting on the soil surface, and (b) stones partially embedded in the soil. The graphs are generalizations of the experimental results of Poesen and Lavee (1991).

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 8.13, 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 8.2a 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 fall 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. 8.2b), suggesting that the change in sediment yield is also a function of differences in summer vegetation growth.

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 to 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 Boquer 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. 8.14). Erosion rates were greater on the Shivta than on the Drorim and

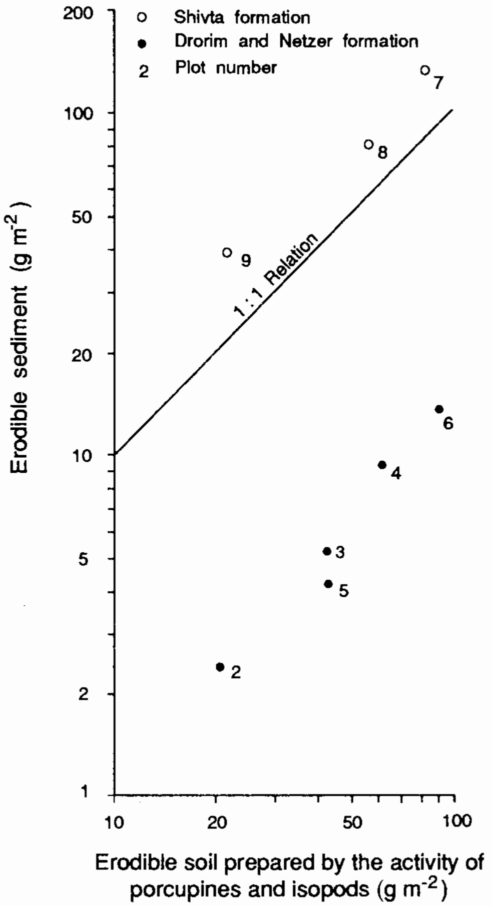

Figure 8.14 Graph of eroded sediment against erodible soil prepared by porcupines and isopods for the Sede Boquer experimental site, northern Negev, Israel. Data from Yair and Shachak (1987, table 10.4).

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 Rutin (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 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 8.1).

RAINDROP EROSION

On desert hillslopes where the vegetation cover is generally sparse (Thomas 1988), the impact of rain-

Table 8.1 Biological activity and sediment production in northern Negev, Israel (data from Yair and Rutin 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 Boquer	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

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

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):

$$Q_{rs} = \frac{E \cos\theta}{R\gamma_b}[0.301 \sin\theta + 0.019 D_{50}^{-0.220} (1 - \exp^{-2.42 \sin\theta})] \tag{8.7}$$

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. 8.15). 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 (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. 8.16). By multiplying the travel distances in Figure 8.16 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 Figure 8.17, which shows that erosion by rainsplash and unconcentrated overland flow reaches a maximum at annual precipitations of 300 to 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.

Kotarba (1980) monitored splash transport on two plots with gradients of 12° and 15° on the Mongolian

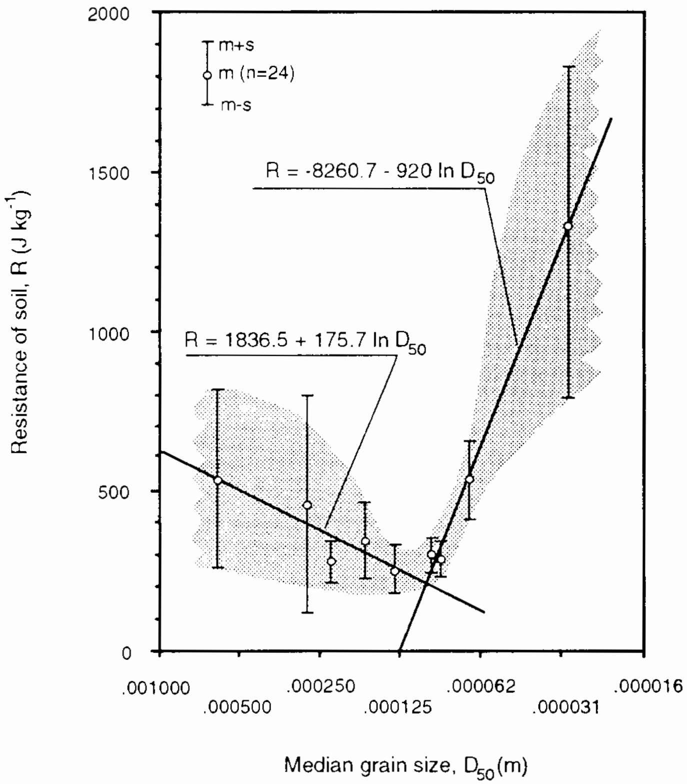

Figure 8.15 Relation between the mean resistance of loose sediments to rainsplash and their median grain size. The standard deviation (s) of experimental values is shown with each mean (m).

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 to 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 to 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 damp 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 to 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.

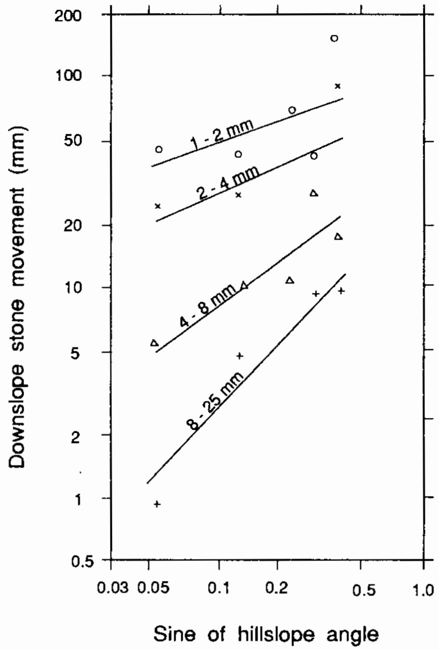

Figure 8.16 Graph of downslope stone movement by rainsplash and unconcentrated overland flow against hillslope gradient for different stone sizes. Data were collected by Kirkby and Kirkby (1974) from painted stone lines on hillslopes near Tucson, Arizona.

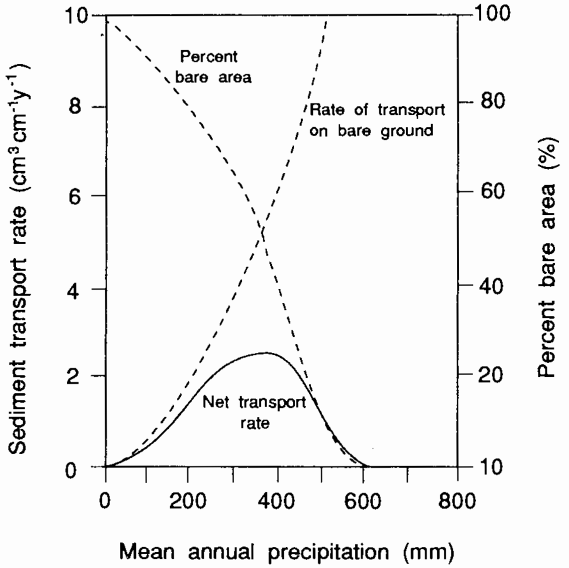

Figure 8.17 Generalized relations of sediment transport rate, percentage bare area, and net transport rate (calculated as the product of the former two variables) to mean annual precipitation for the southern United States (after Kirkby 1969).

Rain Dislodgement

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. 8.18).

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 to 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.

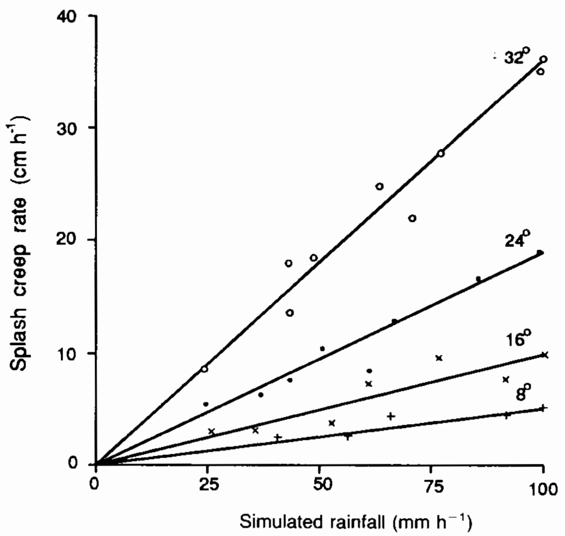

Figure 8.18 Graph of splash creep rate measured in the laboratory against rainfall intensity for different slope gradients (after Moeyersons 1975).

INTERRILL EROSION

Soil Detachment

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 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 8.2, 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.

Although soil detachment in interrill areas 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 $REF = Q \times Q_p^{0.33}/A$, 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.

Most process-based soil erosion models include an equation for the interrill erosion rate D_i that attempts to incorporate the factors that have the greatest influence on this process. For example, the equation in the WEPP (Water Erosion Prediction Project) model has the form (Nearing *et al.* 1989)

$$D_i = K_i I_e^2 C_e G_e (R_s/w) \qquad (8.8)$$

In this equation K_i denotes the base line erodibility which is a function of soil properties; I_e is the effective rainfall intensity during the interval when rainfall intensity exceeds the infiltration rate (i.e. during runoff); C_e represents the effect of the plant canopy in reducing raindrop detachment by intercepting and absorbing its kinetic energy; G_e captures the effect of ground cover on both raindrop and flow detachment – ground cover impedes raindrop detachment by intercepting rainfall and hinders flow

Table 8.2 Sediment yields for covered and uncovered runoff plots, Walnut Gulch, Arizona

Plot number	*Status*	*Gradient (degrees)*	*Percentage vegetation*	*Percentage stones*	*Sediment yield, G* ($g\ m^{-2}\ min^{-1}$)	*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	

detachment by reducing flow velocities; R_s denotes rill spacing; and w rill width. That D_i increases with R_s is presumably due to flow detachment increasing with R_s. The efficacy of equation 8.8 in predicting interrill erosion on desert hillslopes has yet to be evaluated. Work by Simanton *et al.* (1984) suggests that the definition of ground cover may need to be extended to include stones. Furthermore, the fact that desert hillslopes generally have wider interrill zones than the croplands for which the equation was designed may present problems (see below).

It is interesting to note that equation 8.8 does not include a gradient term, though revised versions of the WEPP model are expected to correct this oversight (Liebenow *et al.* 1990). The omission of a gradient term from equation 8.8 is due to the fact that in the WEPP model interrill detachment is assumed to occur wholly by raindrop impact, and detachment by this mechanism increases slowly with gradient (Meyer *et al.* 1975, Quansah 1985). However, desert hillslopes may be steeper and their interrill flow paths longer than those for which the WEPP model was designed. Consequently, gradient is likely to have a significant influence on interrill detachment on such hillslopes and should be included in models for estimating this quantity. Inasmuch as the interrill erosion rate is largely governed by the detachment rate, interrill erosion varies with gradient, but the relation is less steep than that for rill erosion, where detachment is caused almost entirely by flowing water (Meyer *et al.* 1975).

Sediment Transport

Sediment transport in interrill areas may be accomplished by rainsplash and overland flow. Field and laboratory experiments have indicated that sediment transport by rainsplash on bare surfaces is only about one-twentieth of that by overland flow (Young and Wiersma 1973, Lattanzi *et al.* 1974, Morgan 1978). The proportion for natural desert hillslopes is probably fairly similar.

Sediment transport by interrill overland flow is generally represented by either bedload or total load formulae originally developed for rivers (e.g. Komura 1976, Moore and Burch 1986) or simple empirical formulae in which transport capacity is related to a measure of flow intensity (e.g. Foster 1982, Gilley *et al.* 1985a, b, Hartley 1987, Everaert 1991). Although a variety of hydraulic variables is used in these formulae to predict sediment transport capacity, for illustrative purposes, the following discussion will focus on formulae employing total shear stress $\tau = \rho_f g h S$, where ρ_f is the density of the fluid. The use of such formulae overlooks two major considerations.

The first is that the measure of flow intensity controlling sediment transport capacity is not τ but grain shear stress τ_g (Govers and Rauws 1986, Nearing *et al.* 1989), and that grain shear stress expressed as a percentage of total shear stress $\%\tau_g$ is equal to $\%f_g$. Abrahams and Parsons (1991c) found that on gravel-covered hillslopes at Walnut Gulch, $\%f_g$ is typically about 5% (Fig. 8.9), which signifies that $\%\tau_g$ too is about 5%. Therefore, τ must be replaced by τ_g when the above formulae are applied to overland flow otherwise sediment transport capacity will be grossly overestimated. Figure 8.19 illustrates the rapid decrease in transport capacity as form resistance increases in shallow overland flow in a laboratory flume. Inasmuch as form resistance is likely to be large and, hence, $\%\tau_g$ small on all desert hillslopes, it is vital that sediment transport modelling be based on τ_g rather than τ.

The second consideration that the use of formulae that relate sediment transport capacity to flow properties disregards is the influence of raindrop impact on transport capacity. Guy *et al.* (1987) found that raindrop impact was responsible for 85% of the sediment transport capacity in flow depths 0.15 to 0.39 times the mean raindrop diameter. These are very shallow flows (maximum depth 1.5 mm), and the influence of raindrop impact might be expected

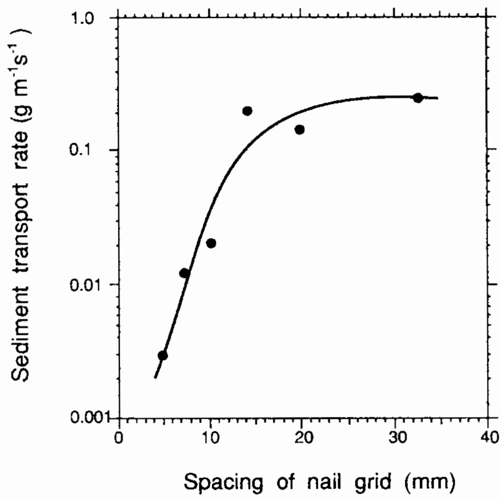

Figure 8.19 Graph of sediment transport rate against nail spacing for a series of flume experiments on beds of well-sorted sands. The nails were arranged in a grid and imparted form resistance to the flow. Thus form resistance increases as nail spacing decreases (after Moss 1980).

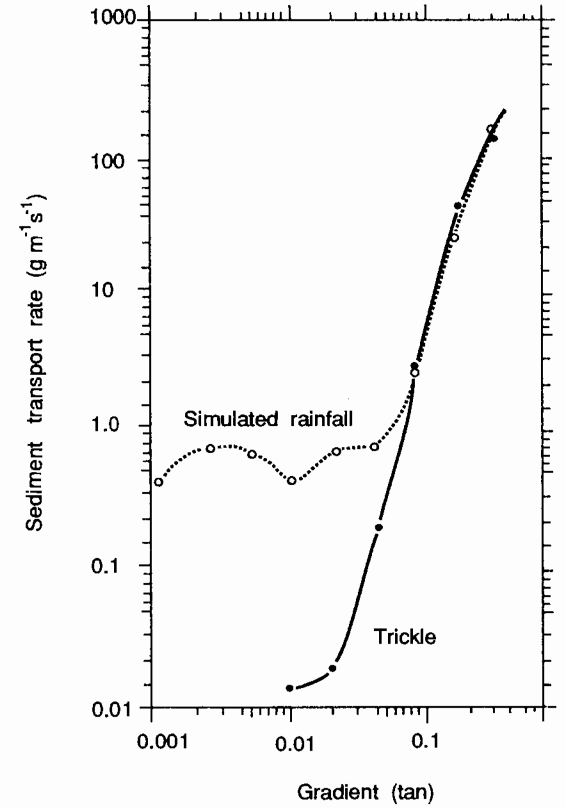

Figure 8.20 Graph of sediment transport rate against gradient for overland flow generated by simulated rainfall and trickle on a bed of poorly sorted sands (after Moss 1980).

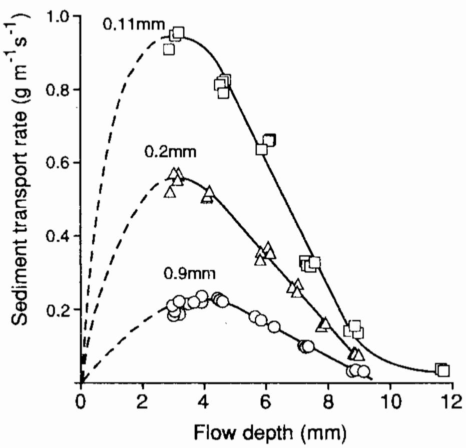

Figure 8.21 Graph of sediment transport rate against flow depth for three sizes of particle (after Kinnell 1991).

to decrease as flow depth increases. Moreover, the laboratory experiments by Moss (1980) suggest that the effect of raindrop impact diminishes as gradient increases and becomes insignificant on hillslopes steeper than 5° (Fig. 8.20). Nevertheless, these findings demonstrate that raindrop impact enhances sediment transport capacity independently of its effect on soil detachment and that in shallow flows on gentle slopes rainfall properties are far more important than flow properties in controlling transport capacity.

In a recent series of laboratory experiments, Kinnell (1991) examined the controls of sediment transport rate in shallow flows where the transport rate was not at capacity and soil particles were detached by raindrop impact. Figure 8.21 shows that the sediment transport rate peaks when the flow depth is about 1.5 times the drop diameter. It decreases as flow depth increases and becomes negligible at depths equal to about 3 times the drop diameter. This negative relation reflects the fact that soil detachment decreases as greater depths of flow increasingly cushion the bed from raindrop impact. For flow depths smaller than 1.5 drop diameters, the sediment transport rate in the flow decreases as flow depth decreases because increasing amounts of sediment are transported by splash. Kinnell found that sediment transport rate also varied linearly with rainfall intensity and flow velocity: greater rainfall intensities eject more particles from the bed into the flow, and greater flow velocities transport these particles farther downstream before they settle back to the bed. However, because flow velocity increases with rainfall intensity in natural rainstorms, these two factors augment one another to cause soil detachment capacity to appear to vary with rainfall intensity squared (equation 8.8).

RILL EROSION

Two conditions must be met in order for rills to form: (a) overland flow must exert sufficient shear stress on the ground surface to detach soil particles, and (b) the transport capacity of the flow must not be satisfied by sediment supplies by raindrop detachment or flow from upslope (Meyer 1986). Many studies have emphasized the first condition (e.g. Moeyersons 1975, Savat and de Ploey 1982, Rauws 1987). However, as Dunne and Aubrey (1986) have shown, rills will not develop, even where the flow is competent to scour its bed, if the flow is fully loaded with raindrop-detached sediment.

Rills develop as a result of significant flow detachment and, once formed, continue to erode chiefly by this mechanism. Field experiments suggest that flow detachment in rills is linearly related to effective shear stress (Van Liew and Saxton 1983). In the

WEPP model the relation has the form (Nearing *et al.* 1989)

$$D_c = K_r(\tau_g - \tau_c) \tag{8.9}$$

where D_c is the detachment capacity of the flow (kg s^{-1} m^{-2}), K_r is a rill soil erodibility parameter (s m^{-1}), τ_g is grain shear stress (Pa), and τ_c is the critical shear stress of the soil (Pa). Note that τ_g is used in place of τ in equation 8.9. As in interrill flow, form resistance dominates grain resistance, and τ_g is a relatively small proportion of τ. Foster *et al.*'s (1984b) analysis of flows through a fibreglass replica of a rill yielded values for $\%\tau_g$ ranging from 8.6 to 38.5%.

In interrill areas detachment is accomplished principally by raindrops and is independent of the sediment load. In contrast, in rills detachment is mainly the result of flow shear stresses and is dependent on the sediment load in the flow. This concept was expressed mathematically by Foster and Meyer (1972) in the equation

$$D_f = D_c[1 - (G/T_c)] \tag{8.10}$$

where D_f is the rill erosion rate (kg s^{-1} m^{-2}), G is the sediment load (kg s^{-1} m^{-1}), and T_c is the sediment transport capacity in the rill (kg s^{-1} m^{-1}). The basic idea here is that as the sediment load approaches the transport capacity, the rill erosion rate becomes a progressively smaller proportion of the detachment capacity of the flow. Equation 8.10 applies where $T_c \geq G$ and D_f is positive. Where $T_c < G$ and D_f is negative, deposition occurs at a rate estimated in the WEPP model by (Nearing *et al.* 1989)

$$D_f = [v_f/(h\ V)](T_c - G) \tag{8.11}$$

where v_f is the fall velocity of the sediment (m s^{-1}).

Perhaps the best equations available for estimating T_c for rill flows are those developed by Govers (1990) from a series of flume experiments using sediment ranging in size from silt to coarse sand. Different empirical equations were developed using shear stress, Bagnold's (1980) effective stream power, and Yang's (1972) unit stream power as the predictive hydraulic variable, and nomograms were produced indicating how the coefficients in these equations vary with sediment size. As in the case of interrill flow, sediment transport capacity of rill flow is affected by raindrop impact and the expenditure of energy in overcoming form resistance. Because rill flows are usually deeper than a few millimetres, the effect of raindrops can for practical purposes be neglected. Form resistance, however, dominates in rill flow as it does in interrill flow (Foster *et al.* 1984b), and this must be taken into account when using Govers' equations to estimate T_c.

MODELLING OVERLAND FLOW EROSION

The concepts employed in process-based erosion models have evolved gradually since Meyer and Wischmeier (1969) suggested mathematical expressions for detachment and transport by rainfall and runoff as conceptualized by Ellison (1944). Important subsequent developments were the proposal of equation 8.10 by Foster and Meyer (1972) and the distinction between rill and interrill sources of sediment by Meyer *et al.* (1975). A variety of process-based erosion models has been formulated during the past 15 years (e.g. Foster *et al.* 1977, 1981, Beasley *et al.* 1980, Gilley *et al.* 1985a, b, Woolhiser *et al.* 1990). One such model has been the WEPP model (Nearing *et al.* 1989). It is extremely tempting to apply soil erosion models such as the WEPP model to desert hillslopes, as the technology is both advanced and accessible and there is no alternative model available in the geomorphic literature. However, existing process-based soil erosion models in general and the WEPP model in particular have a number of deficiencies with respect to desert hillslopes.

First, the WEPP model assumes that the interrill erosion rate can be estimated by equation 8.8. Although this equation allows for soil detachment and sediment transport by interrill overland flow, it was developed for cropland where most detachment and transport is accomplished by raindrops and interrill zones are narrow (typically 0.5 m: Gilley *et al.* 1990). On desert hillslopes interrill zones are much wider (in extreme cases up to 500 to 600 m: Dunne and Aubrey 1986), and soil detachment and especially sediment transport by overland flow become more important.

Not all process-based soil erosion models minimize the role of interrill overland flow. The interrill erosion model of Gilley *et al.* (1985a, b) assumes that detachment of soil particles occurs wholly by raindrop impact and that the transport of these particles downslope is accomplished wholly by overland flow. An example of the output of this model is given in Figure 8.22. Gilley *et al.*'s model suggests that over the greater part of most interrill areas the erosion rate is equal to the detachment capacity – that is, it is detachment-limited. Only on very gentle slopes (e.g. near divides) and in areas of deposition (e.g. concave footslopes) is it equal to the transport capacity – that is, transport-limited. Indeed, because the portion of the interrill zone where erosion is

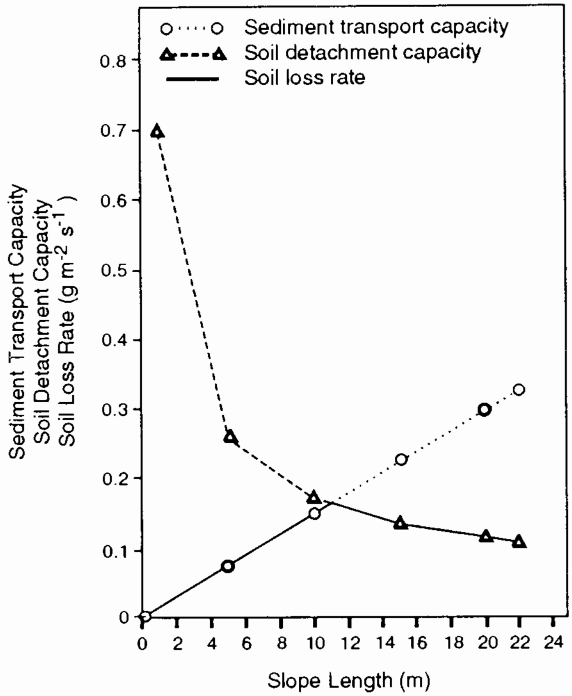

Figure 8.22 An example of the output of the interrill erosion model of Gilley *et al.* (1985a, b). The gradient of the interrill area is 6% (3.4°).

transport-limited is small, some models (e.g. Dillaha and Beasley 1983) have simply assumed that erosion over the entire interrill area is detachment-limited.

Recent work by Parsons and Abrahams (1992), however, has indicated that on a semi-arid hillslope at Walnut Gulch, the interrill erosion rate 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, signifying 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.

A second weakness of available process-based models is that they assume that soil detachment takes place only while runoff is occurring. Such models overlook the accumulation of loose sediment on the ground surface at the start of runoff. This accumulated sediment, which may be considerable on desert hillslopes owing to the typically long intervals between runoff events, has three origins: (a) it is detached by rainfall or flowing water during preceding storms but not removed by overland flow (David and Beer 1975); (b) it is detached by rainfall during the current storm prior to the start of runoff (Yair and Lavee 1977); and (c) it is detached by weathering processes between storms (Ellison 1945, Emmett 1970). The accumulated loose sediment exerts a strong influence on interrill erosion. This influence is commonly manifested in (a) a decline in sediment concentration after the start of runoff reflecting the progressive depletion of detachment storage (Ellison 1945, Yair and Lavee 1977, Abrahams *et al.* 1988) (Fig. 8.23) and (b) a positive correlation between storm soil loss and either the amount of biological activity or intensity of weathering prior to the storm (Yair and Lavee 1981, Roels 1984a) (Fig. 8.14). As a result, erosion rates do not always correlate well with rainfall or surface variables such as appear in equation 8.8.

A third weakness of current process-based erosion models that limits their utility on desert hillslopes is that they were not designed for use on steep hillslopes (>10°) covered with coarse debris, as many desert hillslopes are. Some models may perform quite well on gentle hillslopes where the surface debris is relatively fine. However, on steep hillslopes with nearly complete covers of coarse debris, the hydrology, hydraulics, and erosion mechanics are very different, and these models are unlikely to give good results.

GRAVITATIONAL PROCESSES

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,

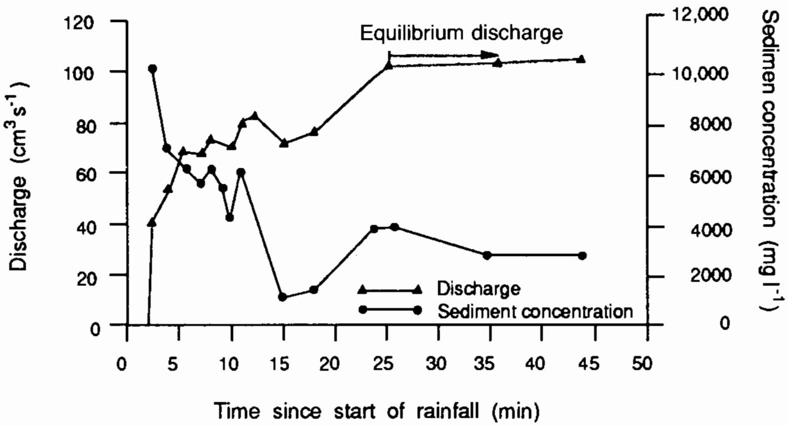

Figure 8.23 Graphs of overland flow discharge and sediment concentration against time since the start of rainfall for a runoff plot at Walnut Gulch, Arizona (after Abrahams *et al.* 1988).

ice, and air. A thorough review is not possible here; the emphasis will be on processes most common in arid environments. Weathering and mass wasting in arid climates generally conform to the following patterns: (a) classic creep processes are relatively unimportant due to the dry regolith and the paucity of fines; (b) deep-seated failures and landsliding are uncommon; and (c) mass movement is dominated by mobilization of surficial layers and their rapid transport ranging from motion of individual particles (talus falls) to debris flows.

CREEP PROCESSES

Slow, relatively continuous movement of thick regolith by matrix deformation is uncommon in arid regions because of the general dryness of the regolith and the paucity of plastic fines. Exceptions may include cold deserts with winter freeze–thaw and clay-rich badland regolith (see Chapter 9). Therefore most coarse-grained desert regolith beneath slopes less than about 25° are generally eroded primarily by wash processes unless mobilized into debris flows due to shallow regolith failure. Where creep is important, a rate of movement proportional to the gravitational stress ($\rho_b gh \sin\theta$) is probably sufficient, where ρ_b is the bulk soil density and h is the effective depth of creep (see Carson and Kirkby (1972) for further discussion of creep mechanics and Chapter 9 for use of creep processes in simulation modelling of slope evolution).

MOVEMENT OF DRY DEBRIS ON SCREE SLOPES

Steep, unstable slopes (talus slopes) of accumulated loose scree arriving from superjacent steep, bedrock slopes are common in desert environments, occurring most noticeably below cliffs (such as the cuesta scarps discussed in Chapter 7) but also as more localized patches on the flanks of many bedrock hills and mountains. Debris may be shed from the bedrock source slopes in a spectrum of events ranging from fall of individual rocks to large debris avalanches. The motion and deposition of individual rocks has been investigated in experiments and theoretical studies (Kirkby and Statham 1976, Statham and Francis 1986). It is generally hypothesized that a rock particle of mass M_p falls to the head of a scree slope from a freefall through a distance z, so that its initial velocity v_0 is

$$v_0 = \sqrt{2gz} \tag{8.12}$$

After the initial impact at the head of the scree slope of inclination θ only the downslope momentum is preserved, so that its initial velocity u_0 parallel to the slope is $v_0 \sin\theta$. As the particle progresses down the slope it bounces, rolls, and slides over particles already on the scree. The net downslope force F acting on the particle of mass M_p is the difference between the downslope component of gravity $M_p g \sin\theta$, and a frictional force $M_p g \cos\theta \tan\phi_d$, where ϕ_d is the dynamic angle of friction. The model assumes that ϕ_d is a constant when averaged over a large number of collisions and a large number of particles, and it ignores non-elastic effects such as particle shattering and exchange of momentum between the falling rock and scree surface rocks through cratering effects. The model thus assumes that the average movement can be treated as a sliding friction over a rough surface. Assuming that the actual slope is less than the dynamic angle of friction, the particle will come to rest in a mean distance $\bar{x}$ given by

$$\bar{x} = \frac{M_p u_0^2}{2F} = \frac{z \sin^2 \theta}{\cos \theta \tan \phi_d - \sin \theta} \tag{8.13}$$

Experiments involving dropping stones on to inclined boards (Kirkby and Statham 1976) have shown that the actual travel distance down the slope varies considerably, but can be represented by an exponential distribution

$$P(x) = e^{-x/\bar{x}} \tag{8.14}$$

where $P(x)$ is the probability of the rock travelling at least the distance x downslope. The exponential distribution implies that the probability of stopping in a given distance interval remains constant downslope (so long as θ and ϕ_d remain constant).

Large rocks move more readily over smaller scree than over a scree equal or larger in size to the mobile particle. Kirkby and Statham's experiments have shown that the friction angle ϕ_d varies with the grain size D^* of the particle and of the particles forming the scree surface D

$$\tan \phi_d = \tan \phi_0 + kD/D^* \tag{8.15}$$

The parameter k depends upon the shapes of the impacting and scree particles and ϕ_0 is the friction value for mobile rocks very much coarser than the scree. Kirkby and Statham (1976) indicated that equation 8.15 is not valid for mobile rocks much smaller in size than the scree.

This model ignores a number of processes and conditions which may be important under some conditions, including air friction on both the freefalling and moving rock, rocks arriving at the scree surface after moving over a sloping bedrock surface (rather than freefall), momentum loss due to cratering (particularly at the first impact at the head of the scree), and interaction between moving particles. None the less, the model provides an excellent fit to experimental data and explains both the concave planform and downslope sorting found on many talus slopes (see below).

On some steep, tall scarps failure may involve a considerable volume of rock and the motion may be by avalanche rather than by motion of individual particles. Gerber and Scheidegger (1974) assume that motion of an avalanche can be described by equations analogous to 8.12 and 8.13. The most common approach to modelling avalanche utilizes the same equations of motion for debris flows that are discussed below.

DEBRIS FLOWS

The movement of mixtures of particulates, water, and air exhibits different flow behaviour depending upon relative amounts of these components and on the grain size distribution of the particulates (Pierson and Costa 1987, Costa 1988). Water flows with cohesionless sediment loads less than about 20% by volume transport sediment via turbulent suspension or by traction at the bed. In these 'water flood' flows (considered under Hydraulic Processes, above and in Chapter 12) the suspended sediment has little effect on flow viscosity and standard fluid flow and sediment transport equations for overwash and channel flows are appropriate. When sediment concentration ranges from 20% to about 47% by volume the flow is 'hyperconcentrated' in that sediment increases flow viscosity and decreases turbulent intensity. A slight shear strength may also be imparted to the flow. Flow density attains values (bulk specific gravity of 1.3 to 1.8) such that buoyant suspension becomes important, and grain-to-grain contacts introduce dispersive stresses (Costa 1984, 1988). The transition to hyperconcentrated flow may occur at sediment concentrations less than 10% by volume for clays (Hampton 1972, 1975). Sediment in hyperconcentrated flows still behaves largely as individual grains, such that larger grains are deposited first as flow velocity diminishes and the resulting sedimentary deposits resemble those of water floods (Costa 1988).

When sediment concentrations exceed about 47% by volume (20% or even less for clay–water mixtures) particle interactions become strong enough that the flow has a well-defined shear strength and a viscosity much greater than water; these are generally called 'debris flows' (Costa 1988). In these flows cohesion, buoyancy, grain-to-grain support, and dispersive stresses resulting from grain collisions (Bagnold 1954) become the dominant mechanisms of sediment suspension and transport, whereas turbulent suspension is greatly diminished relative to water floods. The rheology or resistance to flow in debris flows is a complicated function of sediment concentration, sediment size range, and rate of shearing.

At the highest concentrations of debris relative to matrix the grains are essentially completely supported by grain-to-grain contacts and collisions. Such flows are called granular flows by Pierson and Costa (1987) and include such phenomena as debris and rock avalanches, earthflows, and shallow slides on talus slopes and dune slip faces.

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 hyperconcentrated or water flood flows during the waning stages. 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.

A wide variety of constitutive laws relating resisting shearing stress τ_r to flow and sediment properties have been proposed (see, for example, the reviews in Iverson 1985, Chen 1987, Iverson and Denlinger 1987, Major and Pierson 1992). A fairly general one-dimensional relationship can be expressed as

$$\tau_r = c + \sigma \tan \phi + m' \left(\frac{dv}{dz} \right)^{n'}, \qquad (8.16)$$

where c is shear strength, σ is normal stress (acting on a surface parallel to the bed), ϕ is a dynamic friction angle (not necessarily the same as in equation 8.13), m' and n' are coefficients, and v is downstream velocity at a height z above the bed. A more general three-dimensional formulation of a similar relationship can be found in Iverson (1985, 1986), where it is also applied to slow earthflow landslides. Flows for which $0 < n' < 1$ are called pseudo-plastic (shear-thinning) as contrasted with dilatent (shear-thickening) for $n' > 1$ (Major and Pierson 1992). Two commonly assumed special cases are Bingham plastics for which $n' = 1$, $\phi = 0$, and $m' = \eta_p$ (the plastic viscosity), and Coulomb–Bingham plastics with $n' = 1$, $m' = \eta_p$, but $\phi > 0$.

Flows that are well characterized by the Coulomb–Bingham and Bingham models exhibit a characteristic flow pattern in which the boundaries of the flow are in a laminar shear with a parabolic velocity profile, whereas the centre and top of the flow, where shear stresses are low, is a rigid plug. A rigid central plug has been observed or inferred for many debris flows (Johnson 1965, 1970, Costa 1984, Johnson and Rodine 1984, Pierson 1986). For a two-dimensional flow that is steady and uniform, such that $\sigma = \rho_b g h \cos \theta$ and the gravitational shearing stress is $\tau = \rho_b g h \sin \theta$, the thickness h_c of the plug in Coulomb–Bingham flow occurs for $dv/dz = 0$ and $\tau_r = \tau$:

$$h_c = \frac{c}{\rho_b g (\sin \theta - \cos \theta \tan \phi)} \qquad (8.17)$$

Johnson (1970) and Johnson and Rodine (1984) presented solutions for the plug thickness and velocity distribution for both two-dimensional flows and flows in circular channels. Coulomb–Bingham flows will cease flowing if the flow thickness becomes less than or equal to h_c through thinning of the flow, downstream decrease in θ, or increase in c or ϕ through loss of water or incorporation of more debris.

In a series of experiments with various mixture ratios and total concentrations of sand, silt, and clay Major and Pierson (1992) found estimated values of n' ranging from 0.3 to 3.3. Greatest variability in rheological properties occurs for shear rates (dv/dz) less than 5 s^{-1}, particularly when the volume concentration of sand exceeds 20% and frictional interactions between sand grains dominate shearing resistance. Major and Pierson (1992) found that the Bingham model is a reasonable description of flow behaviour for shear rates greater than 5 s^{-1}. However, both c and η_p increase exponentially with volumetric sediment concentration C_v:

$$c = k_1 e^{b_1 C_v} \text{ and } \eta_p = k_2 e^{b_2 C_v} \qquad (8.18)$$

where k_i and b_i are experimentally determined coefficients (Major and Pierson 1992). Increase of sediment concentration of a few percent can produce order-of-magnitude increases in c and η_p. Very strong dependencies of shear strength of debris flows on sediment concentration have also been noted by Rodine and Johnson (1976), Johnson and Rodine (1984), Costa (1984), Pierson (1986), O'Brien and Julien (1988), and Phillips and Davies (1991).

Clasts may be supported by a variety of mechanisms, including buoyancy (up to 75% to 90% of their weight) (Johnson 1970, Hampton 1975, 1979, Middleton and Hampton 1976, Rodine and Johnson 1976, Costa 1984, 1988), dispersive stresses due to collisions between particles (Bagnold 1954), cohesion (Johnson 1970, Hampton 1975, Rodine and Johnson 1976), structural support from grain to grain contacts which can support up to one-third the weight of coarse particles in flows with solid volumetric concentrations greater than 35% (Pierson 1981), and turbulence in dilute flows. The mechanisms that concentrate coarse grains at the surface may include dispersive stresses introduced by the greater number of collisions with smaller grains on the bottom relative to the top of clasts in shear flow (Iverson and Denlinger 1987), kinetic sieving in which fines settle through voids between coarse clasts during particle vibrations (Middleton 1970, Bridgewater 1980), and angular momentum applied to large particles in shear flows (with clast tops rotating downstream) such that large clasts can roll over smaller ones (Iverson and Denlinger 1987).

A comprehensive theoretical foundation for constitutive relationships (e.g. equation 8.16) for debris flows based upon the mechanics of the matrix and solid interactions (including grain-to-grain collisions) is lacking. Iverson and Denlinger (1987) pointed out the complexity of the processes involved in debris flows, including collisions, rubbing, rotations, vibrations and possibly fracturing of the solid phase and flow, compression, vibration and cavitation of the matrix. The Bingham and Coulomb–Bingham models (Johnson 1965, 1970, Yano and Daido 1965) do not include dynamic particle interactions but can explain flow over diverse slopes, formation of a rigid plug, support of large clasts, and the blunt nose of debris flows (Iverson and Denlinger 1987).

Takahashi (1978, 1980, 1981) and McTigue (1979) have proposed theoretical models based upon Bagnold's (1954) observations that mutual particle collisions in granular flows produce dispersive stresses that can help support such flows. These theories assume a uniform grain size and uniform particle concentration, and result in constitutive equations with no cohesive term and shear stresses proportional to the square of the shear rate ($n' = 2$ in equation 8.16). These theories can explain some observed velocity profiles in debris flows, inverse grading, and segregation of coarse particles in the snout of debris flows. However, the limitation of uniform grain size and particle concentration, as well as neglect of particle–matrix interactions, is a major shortcoming (Iverson and Denlinger 1987). More general experiments and theoretical formulations of particle collisions in dry grain flows (e.g. Lowe 1976, Middleton and Hampton 1976, Savage 1979, 1984, 1989, Lun *et al.* 1984, Campbell and Brennen 1985, Drake and Shreve 1986, Haff 1986, Hui and Haff 1986, Melosh 1987, Drake 1990) may ultimately provide better debris flow models that incorporate inelastic collisions and variable particle sizes and concentrations. However, Iverson and Denlinger (1987) pointed out that solid–fluid interactions are important in debris flows, particularly the cushioning effect on collisions afforded by the matrix (Biot 1956, Davis *et al.* 1986, Iverson and LaHusen, 1989).

Source area characteristics and mechanisms for initiation of debris flows are not well understood. The role of steep relief and high rainfall intensity/duration is cited by nearly every study (e.g. Rice *et al.* 1969, Scott 1971, Campbell 1975, Nilsen *et al.* 1976, Cooley *et al.* 1977, Scott and Williams 1978, Weiczorek 1987, Florsheim *et al.* 1991). However, Hooke (1987, p. 512) pointed out that overland flow and water floods cannot be converted to mudflows simply by entrainment of sediment from flow boundaries due to the lack of turbulent mixing in debris flows. A variety of mechanisms have been identified (see general discussions in Johnson and Rahn 1970, pp. 310–33, Costa 1984, Johnson and Rodine 1984, Hooke 1987). The most common mechanism is probably the conversion of slumps and small landslides into debris flows either through dilation through incorporation of additional water in hollows and channels or through liquefaction if the original deposit has high intergranular porosity or extensive macropores (see reviews of the extensive literature in Costa 1984, Johnson and Rodine 1984; also see Ellen and Fleming 1987, McDonnell 1990, Guzzetti 1991). Johnson and Rodine (1984, pp. 313–7, 329–31) described the transformation of slumps into debris flows within a few seconds and after travel of only a metre or less. Slumps of slope or channel banks into rill and gully flows is another mechanism (Johnson and Rodine 1984, pp. 321–4). Such slumping may occur during flow events or between flows; in the latter case the debris forms small dams that create debris flows when they fail, incorporating the ponded water into the debris. The flow of water draining from bedrock slopes and chutes on to loose talus may be one of the most important mechanisms of debris flow generation on arid slopes (Johnson and Rodine 1984, pp. 329–31); such flows occur as high velocity jets that sluice the upper portions of talus slopes, and additional debris will be incorporated from the unstable talus slopes as the flow progresses downslope.

Debris flows from slopes in semi-arid and strongly seasonal climates commonly occur shortly after fire has destroyed the natural vegetation cover (Sidle *et al.* 1985, Wohl and Pearthree 1991). 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 role of water floods in creating rills and gullies on desert slopes is reasonably well understood (discussion above and in Chapters 9 and 12). However, the role of rockfalls, avalanches, and debris flows in slope sculpture is poorly characterized. Debris flows generally are confined to established channels, at least until they debouch on to pediments or fans, but there is disagreement as to the efficacy of such flows in creating or deepening channels. Lustig (1965), Pierson (1980), Janda *et al.* (1981), Johnson and Rodine (1984, p. 273) and Costa (1984, p. 274) cited evidence for erosion by debris flows, but Hooke (1987) disagreed, citing the lack of turbulence in such flows and the possibility that most observed channel erosion occurs from the waning-stage water floods occurring after the passage of the debris flow; similar views were advanced by Blackwelder (1928) and Morton and Campbell (1973). None the less, the momentum of debris flows is sufficient to shear trees and demolish structures, so that it is certainly capable of incorporating additional large boulders along the channel margins. Furthermore, debris flows exert shear stresses along their margins that may cause failure of partially weathered bedrock or loose regolith.

SLOPE STABILITY

Deep rotational landslides and slow-moving earthflows (Iverson 1986, Keefer and Johnson 1983) are rare in arid regions because of the shallowness and general aridity of the regolith. Slope failures generally involve stripping of thin surficial regolith layers, so that a stability analysis involving finite slopes with a potential failure surface parallel to the surface and flow likewise parallel to the surface (equations 9.1 to 9.5) is often sufficient. However, failures are often initiated at zones of groundwater seepage; such seepage will produce instability on gentler slopes than indicated by equations 9.1 to 9.5 (Iverson and Major 1986). Surficial regolith may be acted upon by a combination of gravity, seepage, and overland flow. A simplified torque-balance approach has been formulated by Kochel *et al.* (1985), Howard and McLane (1988), and Dunne (1990) for the critical stability of a surficial particle (Fig. 8.24).

$$F_g D \sin(\alpha - \theta) + F_c D - F_w D \cos\alpha - F_s D \cos(\theta + \psi - \alpha) = 0 \qquad (8.19)$$

where the forces due to gravity F_g, seepage F_s, cohesion F_c, and surface water flow F_w are assumed to be related to the particle size D by

$$F_g = C_1(\rho_s - \rho_f)gD^3 \qquad (8.20)$$

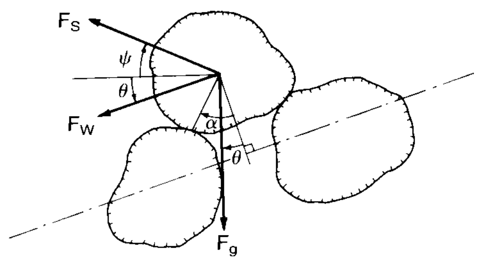

Figure 8.24 Definition of angles and forces on a surficial particle acted upon by seepage and runoff.

$$F_s = C_2\rho_f giD^3 \qquad (8.21)$$

$$F_c = C_3cD^2 \qquad (8.22)$$

$$F_w = C_4\tau_s D^2 \qquad (8.23)$$

where the C_i are coefficients determined by particle geometry, ρ_s is particle density, i is the hydraulic gradient, and τ_s is the shear stress exerted by the surface flow. Kochel *et al.* (1985) and Howard and McLane (1988) showed that the hydraulic gradient is uniquely related to the slope and seepage exit angles by the expression

$$i = \sin\theta/\cos(\psi + \theta) \qquad (8.24)$$

Howard and McLane (1988) investigated the mechanisms of combined seepage and runoff erosion for cohesionless sediments ($c = 0$).

Iverson and Reid (1992) and Reid and Iverson (1992) combined two-dimensional elastic solutions of regolith stress–strain behaviour, Darcian groundwater flow, and a Coulomb failure criterion to investigate slope stability with groundwater flow for saturated conditions. They concluded that stability of slopes with deep, saturated regolith is underestimated by infinite slope analysis such as conducted by Iverson and Major (1986). Slope conditions conducive to failure are high steepness and convexity of slope profiles and outcrop of low-conductivity materials that direct seepage vectors outward.

TRANSPORT OF COARSE DEBRIS

As the name implies, rock-mantled slopes are covered with a layer of coarse debris. Field observations have established that at least some of this debris moves downslope (as opposed to weathering *in situ*). However, the identity and relative importance of the processes causing this movement have been the subject of considerable debate. These processes will be discussed in this section.

The investigation of these processes has been dominated by studies in the American South-west. Lawson (1915) recognized that rock-mantled slopes (termed 'rock slopes' by him) have gradients less than 35° and are inclined at angles less than the angle of repose of the loose material on their surface. He maintained (p. 29) that 'sheets of water . . . wash down the detritus in times of cloud-burst', but 'rain wash co-operates with gravity and transportation is far more efficient' (p. 26). Bryan (1922, 1940), on the other hand, placed greater emphasis on the role of gravitational processes, such as falling, rolling, sliding, and creep, in transporting coarse surface debris. However, he conceded that smaller surface debris can also be undermined and carried away by rain wash, and that the development of gullies during 'cloud bursts' can strip surface debris from whole sections of hillslope. Melton (1965) appears to have held similar views to Bryan. He claimed that particles up to 64 mm in diameter could be transported downslope by hydraulic processes, whereas larger cobbles and boulders moved only by rolling and sliding on slopes steeper than about 28°, which he equated with the angle of sliding friction.

Two studies have recorded the actual movement of stones 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 ≥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'.

Although Melton (1965) emphasized gravitational processes while Kirkby and Kirkby (1974) and Abrahams *et al.* (1984) stressed hydraulic ones, their findings are not incompatible, as they examined different ranges of gradient. Indeed, a hypothesis consistent with all the findings is that coarse debris moves mainly by sliding and rolling on slopes steeper than 28° and by hydraulic activity on gentler slopes. Abrahams *et al.* (1990) investigated this hypothesis by analysing 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 them being drawn into the holes and tilted upslope.

That hydraulic processes, including runoff creep, appear to be the main mechanism whereby coarse debris is moved downslope on Bell Mountain does not necessarily mean that all the coarse debris is transported by this mechanism. Nor does it mean that hydraulic processes necessarily dominate on other desert hillslopes either in the American South-west or elsewhere. In the semi-arid mountains of southern California, coarse debris is moved down hillslopes close to their angle of repose (Melton's angle of static friction) by a combination of gravitational and hydraulic processes. Gravitational processes are locally referred to as dry erosion and hydraulic processes as wet erosion. Dry erosion includes surface creep, particle sliding, and particle flow and may be triggered by wetting and drying, earthquakes, wind, or animals. This process involves a wide range of debris sizes and is sensitive to vegetation cover, with rates of movement being greater under shrubs than under grass (Wohlgemuth 1986). Field measurements have established that dry erosion is more rapid than wet erosion, which includes rainsplash, unconcentrated overland flow, and rilling (Rice 1982, Wohlgemuth 1986).

A combination of gravitational and hydraulic processes also seems to be at work on hillslopes in semi-arid northern Kenya. Over a two-year period Frostick and Reid (1982, Reid and Frostick 1986) monitored the movements of three lines of painted stones ≥ 10 mm in diameter on a 30° hillslope. On the planar portion of the hillslope they found that the distance moved was positively correlated with size of particle (cf. Kirkby and Kirkby 1974, Abrahams *et al.* 1984), which led them to infer that these movements were due to unspecified gravitational processes. In contrast, in the chutes and rills, the particles moved much farther, but the distance moved was uncorrelated with particle size. The movement of particles in the chutes and rills was attributed to concentrated overland flow eroding interstitial fines, thereby undermining the coarse particles and facilitating their downslope movement (i.e. runoff creep).

Finally, there is ample evidence that on still other desert hillslopes, debris flows are largely responsible for the downslope transport of coarse debris. Such hillslopes are generally steep – that is, they are at or near the angle of repose of their surface material. Moreover, the relative importance of debris flows appears to vary with the character of the substrate. Substrates that produce an abundance of fine debris upon weathering are more likely to generate debris flows than those that do not. Similarly, substrates that favour the formation of a relatively impermeable caliche layer close to the surface, thereby aiding saturation of the surface materials, are more susceptible to debris flows than those that do not. Hillslopes where debris flows are the dominant process transporting coarse debris have been described by Mabbutt (1977, pp. 45–6), Gerson (1982), and Gerson and Grossman (1987).

To conclude, desert hillslopes appear to lie along a continuum with respect to the manner in which coarse debris is moved downslope (Carson and Kirkby 1972, p. 345). At one end of the continuum are slopes inclined at the angle of repose of their mantling debris. On these slopes gravitational processes, both dry and wet, dominate. On slightly less steep slopes, gravitational processes remain important, but quite coarse debris can also be moved directly by overland flow; in addition, much debris is probably moved indirectly by runoff creep. At the other end of the continuum are gentler slopes where hydraulic processes are largely responsible for the downslope movement of coarse debris by both direct and indirect means and gravitational processes such as creep are of little significance.

DEBRIS SLOPES

SLOPE FORM AND ADJUSTMENT

Debris slopes are usually thought of as relatively steep – that is, with gradients in excess of 10° – but in this chapter they include all rock-covered slopes, regardless of gradient, whose surface accumulation of coarse debris has weathered from the underlying rock. Typically the profiles of such 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 debris 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, 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 downslope, 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.* 1993) (Fig. 8.25).

A more detailed picture of the variation in mean particle (fines plus debris) size with gradient down a debris slope profile is presented in Figure 8.26. This

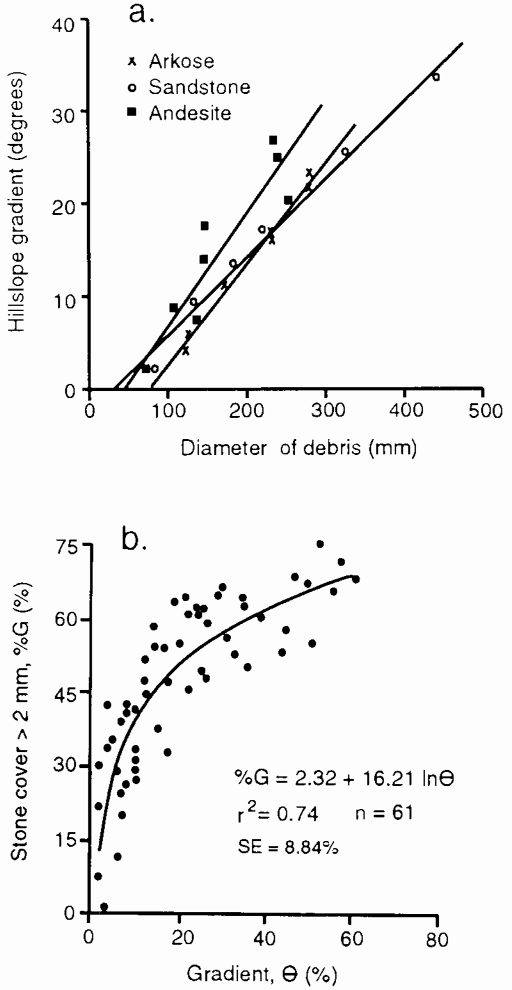

Figure 8.25 Graphs of hillslope gradient against various measures of particle size. (a) Graph of hillslope gradient against the mean size of the ten largest debris particles for three debris slopes on three rock types in southern Arizona (after Akagi 1980). (b) Graph of stone cover against hillslope gradient for 12 debris slopes in Walnut Gulch Experimental Watershed, Arizona.

profile is located in the Mojave Desert, California, and is underlain by closely jointed latitic porphyry (Fig. 8.27). Beginning 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, debris slopes without basal streams, including those that are both much steeper and much gentler than this example, and 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 debris slopes is suggested by a recent study of the relation between gradient S and mean particle size $\bar{D}$ for debris 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–$\bar{D}$ relations with similar slope coefficients but different intercepts (Fig. 8.28). However, on a given rock the slope coefficient varies with debris slope planform, being greater for plan-concave slopes than for plan-convex ones (Fig. 8.29).

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

$$G \propto X^m S^n / \bar{D}^p \qquad (8.26)$$

where G is sediment transport rate, X is horizontal distance from the divide, and m, n, and p are positive coefficients. Now if debris slopes are formed by and adjusted to hydraulic processes, equation 8.26 may be manipulated to ascertain how the S–$\bar{D}$ relation varies with debris slope planform. Because $\bar{D} \propto X^q$, where $q < 0$,

$$G \propto \bar{D}^{[(m/n)-p]} S^n \qquad (8.27)$$

Rearranging equation 8.27, one obtains

$$S \propto G^{1/n} / \bar{D}^{[(m/q)-p]/n} \qquad (8.28)$$

From equation 8.26 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 $\bar{D}$ in equation 8.28, and the steeper is the S–$\bar{D}$ relation. Thus the analysis predicts that plan-concave debris slopes have steeper S–$\bar{D}$ relations than plan-convex ones. The agreement between the analysis and observed variation in the S–$\bar{D}$ relation with planform implies that the debris slopes are formed by and adjusted to hydraulic processes.

In the above sediment transport equation (equation 8.26), the arithmetic mean particle size $\bar{D}$ was used as the measure of particle size because, of the several measures of particle size tested, it correlated

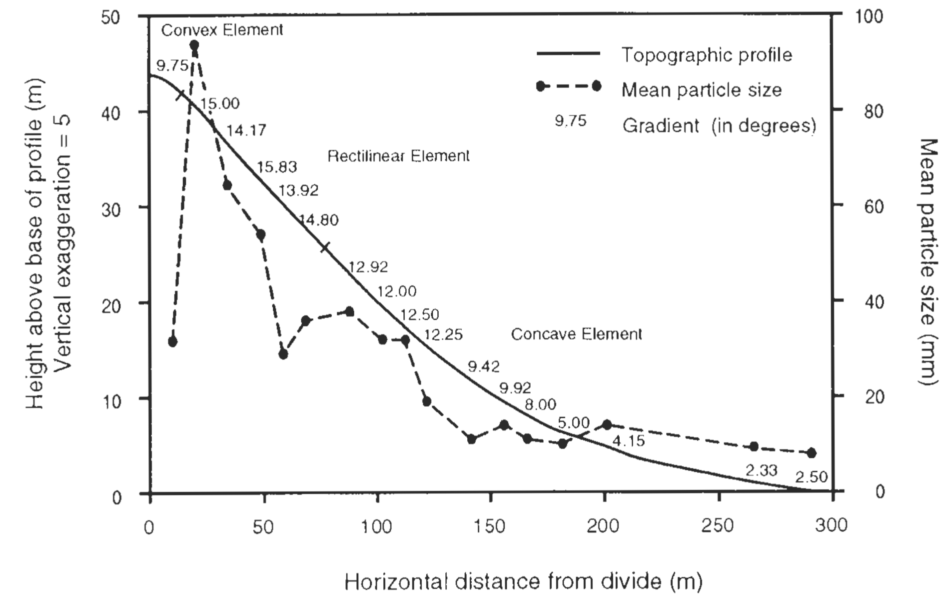

Figure 8.26 Debris slope profile showing the downslope variation in mean particle size (sample size 100) and gradient (measured length 5 m). The debris slope is depicted in Figure 8.27.

most highly with gradient. Variable $\bar{D}$ worked best in this instance possibly because it represented resistance to flow, and the contribution by each piece of debris 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 $\bar{D}$. For example, $\bar{D}$ 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.

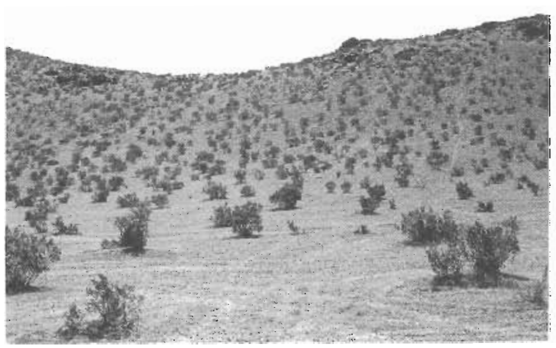

Figure 8.27 Photograph of well-adjusted debris slope underlain by latitic porphyry in Turtle Valley, Mojave Desert, California.

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. 8.13). 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 8.26 may be a useful start to the modelling of debris slope form, the situation on actual debris slopes is probably far more complex, and that a great deal more work is required to elucidate the effect of debris size and cover on sediment transport rate.

A discussion of debris 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 26) and have strongly influenced the form of many debris slopes (Chapter 21). 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 boul-

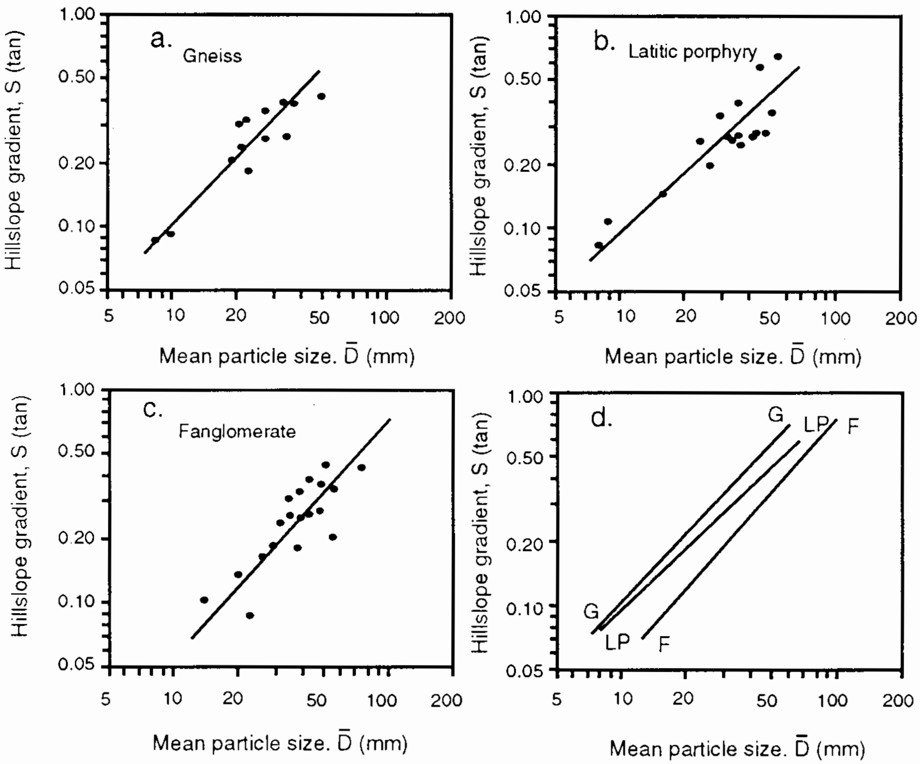

Figure 8.28 Graphs of hillslope gradient against mean particle size for plan-planar debris slopes underlain by (a) gneiss, (b) latitic porphyry, and (c) fanglomerate. The fitted lines in (a), (b), and (c) are reproduced in (d) for comparative purposes (after Abrahams *et al.* 1985).

der-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. 8.30). 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 debris slopes than is generally realized. Certainly, Oberlander's description of boulder-clad slopes in the Mojave Desert applies to debris slopes in most granitic terranes. Erosion on such slopes is often characterized as weathering-limited (e.g. Young 1972, p. 206, Mabbut 1977, p. 41). However, if this were truly 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 adjustment is inversely related to slope gradient and rock resistance (Fig. 8.31). 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

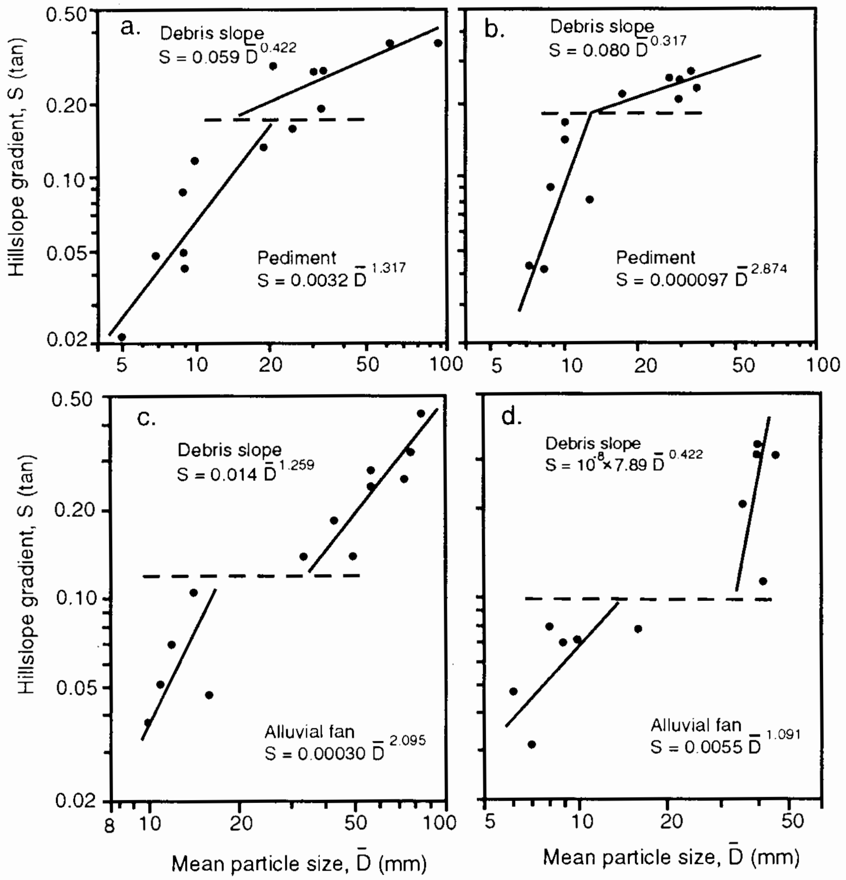

Figure 8.29 Graphs of hillslope gradient against mean particle size for two (a, b) plan-convex debris slopes and their basal pediments and two (c, d) plan-concave slopes and their basal alluvial fans (after Abrahams 1987). Note that the S–D relations are much steeper for the plan-concave slopes than for the plan-convex ones.

are far from being adjusted to contemporary hydraulic processes, especially in their steeper parts.

PIEDMONT JUNCTIONS

In desert landscapes, hills and mountains generally arise abruptly from their surrounding plains or piedmont. The transition zone between the piedmont and these upland areas 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 debris slope profiles. These piedmont junctions occur at the transition between a pediment and its backing hillslope and extend 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. 8.32a). At the other extreme are features whose concavity is so slight that they can reach lengths of 750 m (Fig. 8.32b) (Kirkby and Kirkby 1974). Numer-

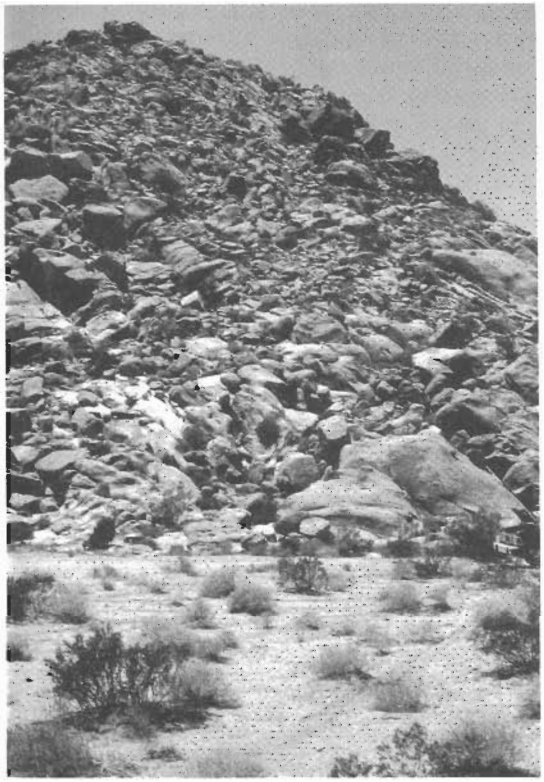

Figure 8.30 Photograph of a steep, poorly adjusted debris slope developed on widely jointed quartz monzonite in the Mojave Desert, California.

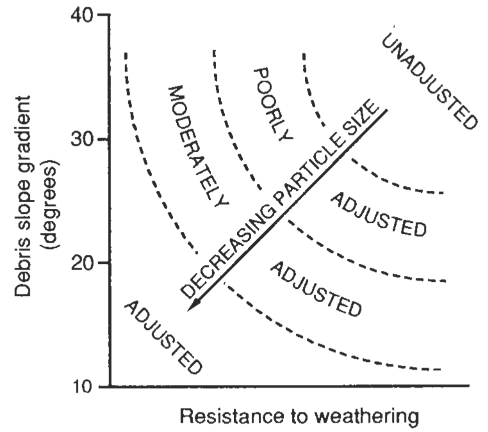

Figure 8.31 Graph of debris slope gradient against resistance to weathering, showing how the degree of debris slope adjustment varies with these variables (after Parsons and Abrahams 1987).

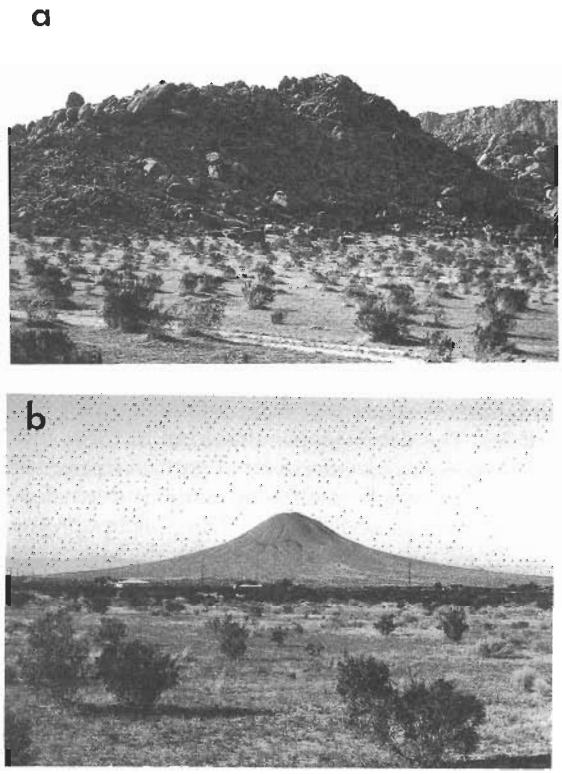

Figure 8.32 Photographs of piedmont junctions, showing (a) a narrow, highly concave one formed on widely jointed quartz monzonite, Mojave Desert, California, and (b) a broad, gently concave one developed on quartz monzonite with a latitic porphyry debris mantle derived from the caprock of the residual.

ous 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 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 debris 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 past two decades this relation has been 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 debris 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.

The foregoing discussion applies to debris slopes that are adjusted to present-day processes, at least in the vicinity of the piedmont junction. However, not all debris 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. 8.30). 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. Thus 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.

TALUS SLOPES

Talus or scree slopes are a ubiquitous accompaniment to steep bedrock slopes. Most occur as small fillets in breaks-of-slope in bedrock slopes, but more extensive talus occurs below resistant caprocks in layered rock (Chapter 7). Most studies of scree slope development and sedimentology have been undertaken in arctic and alpine environments (see references in Kirkby and Statham 1976, Statham and Francis 1986). Delivery of debris to talus slopes may occur as falls ranging from individual rocks to debris avalanches. A model of debris motion and deposition has been presented above (equations 8.12 to 8.15). Two characteristics of talus slopes are explained by this model (Kirkby and Statham 1976). Talus is commonly sorted downslope, with coarser blocks at the foot of the slope. The differential movement of rocks over a scree of a different size (equation 8.15) explains this sorting, since the largest rocks have enhanced mobility. Furthermore, talus slopes are generally concave due to the exponential travel distance decay function (equation 8.14); the concave tail is created by rocks composing the higher-mobility tail of the distribution. The concavity is best developed on short talus slopes and on talus extending out on to a horizontal rather than downward sloping basal surfaces (Kirkby and Statham 1976).

Because the dynamic friction angle of moving debris, which controls deposition, is 5° or more greater than the static friction angle, scree slopes are at least marginally stable when deposited (Statham and Francis 1986). Weathering of scree may increase mechanical stability of the slope due to generation and movement of fine debris leading to a wellgraded and dense packing. However, weathering also decreases infiltration capacity and can lead to enhanced runoff erosion and regolith saturation during precipitation events, which can decrease slope stability (e.g. equations 9.1 to 9.5) (Statham and Francis 1986). Concentrated runoff from bedrock slopes above the scree means that instability will occur most commonly at the head of the talus slope (Statham and Francis 1986), including mobilization of debris flows as discussed above.

Classic models of scree slope development assume an initially high cliff bordering a basal slope of low inclination, with the cliff backwasting, forming talus that gradually swamps the cliff face, leading ultimately to the entire slope becoming talus and a characteristic convex profile of the talus–bedrock contact (Fisher 1866, Lehmann 1933, Bakker and Le Heux 1946, 1947, 1950, 1952). This idealized pattern of evolution seldom occurs in nature (Statham and Francis 1986). Many small scree slopes in depressions on bedrock slopes are essentially surfaces of transportation, delivering as much material to the bedrock slope at the foot of the scree as arrives from upslope. Such scree may wax or wane in size depending upon the pattern of weathering and erosion of the super- and subjacent bedrock slopes. A similar pattern occurs on many basal screes debouching directly into fluvial washes or chutes, where input from above is balanced by erosion from below. In fact, the ideal model presented above is probably the exception, occurring primarily where a scarp is created fairly suddenly from faulting, landsliding, or fluvial undercutting. Even more complex relationships between talus emplacement and erosional removal occur on cuesta scarps due to the difference in erosional resistance between the caprock talus and the weak subcaprock unit. In these circumstances the talus may be eroded laterally and undergo several episodes of remobilization prior to removal (Chapter 7). Over the long run talus slopes exhibit a balance between addition of debris from above and its removal through weathering, runoff erosion, mass wasting, and basal erosion, although individual talus slopes may wax or wane in size due to changes in geo-topographic setting or climatic changes.

DISCUSSION

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 both in a quantitative and qualitative sense, but the long-term interaction of processes and materials to create specific types of desert slopes is poorly characterized.

ACKNOWLEDGEMENTS

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