Chapter 11

Catchment and Channel Hydrology

John B. Thornes†

Introduction

Solar radiation, wind, and water are the driving agents of desert landscapes. Water has four major roles to play. First, in sustaining any life forms that exist; secondly, as a chemical substance which interacts with other chemical substances, notably salts; third, as a medium of transport of mass; and, fourth, as a direct source of energy. The last role, though small by comparison with the roles of solar and wind energy, may none the less be critical in determining the threshold of operation of runoff, through its impact on infiltration. Since infiltration is one of the major thresholds in dryland morphological development, the factors which control it have a role out of all proportion to the energy involved.

We have come to take for granted the well-known hydrological cycle of arid lands (as illustrated in Fig. 11.1). The hydrological cycle of desert environments has the same inputs and outputs, and there is the same requirement for the conservation of mass and energy.

It is the relative importance of the different components which is critical. In temperate environments percolation, throughflow, saturated flow, and groundwater play a most significant role. In dry environments infiltration normally only occurs to shallow depths, soil moisture is consistently low or very low, and overland flow is important, with deep percolation and groundwater being relatively unimportant except in suballuvial or externally recharged aquifers. This relative importance also varies spatially and through time more than in temperate environments. For example, under even sparse vegetation infiltration becomes relatively much more important than between vegetation. Under seasonal control, infiltration may be much more important in the wetter than in the drier season, and so on. As this balance changes so does the relative role of different geomorphic processes. The differences between desert and temperate hydrology from a geomorphological point of view are therefore more complex than simply a shift in the magnitude and frequency of events.

The major differences, which are illustrated and developed in the remainder of this chapter, can be summarized as follows.

(a) Rainfall occurs at high intensities, with low overall amounts, at irregular intervals, often with a strong seasonal bias and usually with a very large interannual variability.
(b) The rain falls on ground with a sparse or nonexistent vegetation cover, which is irregular in its distribution and especially adapted to collect rainfall. Interception rates are low and highly variable and rapid direct evaporation of excess surface water is characteristic.
(c) Infiltration is largely controlled by the bare surface characteristics which range from sands to organic crusts and from stones to chemical precipitates.
(d) Losses due to evapotranspiration are dominated by soil-water availability and controlled by profile characteristics as well as by atmospheric stress; subsurface water movement may be significantly affected by regolith chemistry.
(e) Overland flow is relatively more likely when storms occur and the terrain over which it occurs may be exceptionally rough.

J.B. Thornes (✉)
Department of Geography, King's College London, Strand, London WC2R 2LS, UK
†deceased

A.J. Parsons, A.D. Abrahams (eds.), *Geomorphology of Desert Environments*, 2nd ed.,
DOI 10.1007/978-1-4020-5719-9_11,

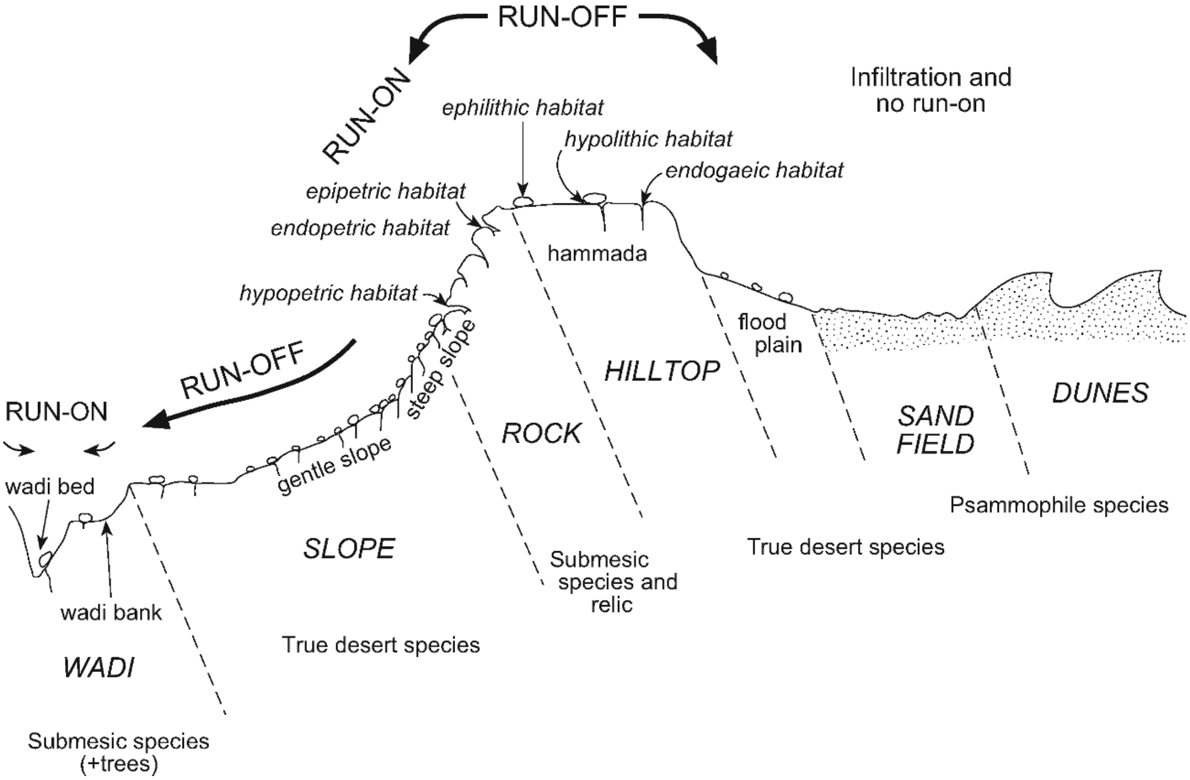


Fig. 11.1 Elements of the hydrological cycle in arid lands (after Shmida et al. 1986)

(f) Channel flow is ephemeral and, hence, significantly influenced by boundary conditions, especially transmission losses to alluvium.

(g) Groundwater obeys the same rules as in temperate environments, but suballuvial aquifers assume a more significant hydrological role with ephemeral channel flow.

Precipitation

Magnitudes at Different Timescales

From a geomorphological point of view, annual rainfalls tend to determine the overall character of the environment through the vegetation condition because, to a first approximation, the gross vegetation biomass and productivity are determined by rainfall for areas with less than about 600 mm (Walter 1971, Leith and Whittaker 1975). Above this figure nutrients become more significant in limiting plant growth. Below it the physiological response of plants in biomass terms to gross precipitation may actually be quite variable because ultimately it is the available moisture for plant growth that is the controlling factor. This availability reflects evapo-transpiration amounts and the water use efficiency of particular species (Woodward 1987). Shrubs and dwarf shrubs are the most conspicuous life forms in desert regions. At the wetter margin shrubs such as *Artemisia, Atriplex, Cassia, Ephedra, Larrea*, and *Retama* prevail, whereas in extreme deserts dwarf shrub communities are present, such as the Chenopodiacae, Zygophillaceae, and the Asteraceae (such as *Artemisia*). Typically the semi-deserts have 150–350 mm of rain, whereas the extreme deserts have less than 70 mm, according to Shmida (1985).

Generally here we follow Le Houerou (1979) in restricting deserts to environments with less than 400 mm y^{-1}, which in North Africa and in southern Europe corresponds to the northern limit of steppe vegetation, such as *Stipa tenacissima* and *Artemisia herba-alba*. Le Houerou differentiates extreme desert lands as those with less than 100 mm, corresponding to the northern border of typical desert plants, such as *Calligonium comosum* and *Cornulaca monacantha*.

Annual rainfall magnitudes are also reflected in altitudinal variations in plant cover. Whittaker and Niering (1964), for example, show the clear gradation from desert to forest developed in the Santa Catalina mountains of Arizona.

There have been several models which relate geomorphic processes or responses to annual rainfall amounts and temperature (Peltier 1950, Langbein and Schumm 1958, Carson and Kirkby 1972), and these consistently show low total magnitudes of rainfall having low rates of geomorphic activity but with sharply increasing rates up to a maximum at about 300 mm.

There is an enormous interannual variability in total amounts in dry areas. Figure 11.2 shows interannual variations of rainfall totals for Murcia, south-east Spain, for 124 years, expressed as rainfall anomalies (mean annual rainfall/standard deviation). Here the mean annual rainfall is 300 mm, and the interannual coefficient of variation 35%, a figure typical of semi-arid environments. Bell (1979) has shown that, in general, annual precipitation variability is highest in zones of extreme aridity and that the lowest variations are at the poleward margins of the deserts, where frontal rainfall forms a greater proportion of the total annual rainfall. As a general rule the standardized rainfall anomaly decreases as the mean annual rainfall increases.

The geomorphic significance of seasonal rainfall distribution varies with evapotranspiration because these together determine soil moisture conditions and infiltration rates, decomposition rates for organic matter, soil erodibility, evaporation rates, solute movement, and chemical precipitates.

Figure 11.3 shows the annual pattern of soil moisture for the Murcia site shown in Fig. 11.2. The relative importance of seasonal rainfall totals depends on how much the plant cover varies. Typically in arid and semi-arid conditions the seasonal impact is principally on annual herbs. In semi-arid environments these may account for as little as 5% of the total cover, the amount increasing as the conditions become drier. The seasonality of rainfall is generally expressed in some ratio of monthly to annual rainfall. For example the Fournier index (Fournier 1960) expressed seasonality by the index p^2/P, where p is the rainfall of the wettest month and P the annual rainfall. By regression analysis he showed that this index is related to total sediment yield. The yield increases with seasonality and relief, and the rate of increase is greater in desert areas. More recently Kirkby and Neale (1987) indicated that it is seasonality, expressed as the degree of rainfall concentration, which leads to the characteristic Langbein and Schumm (1958) curve.

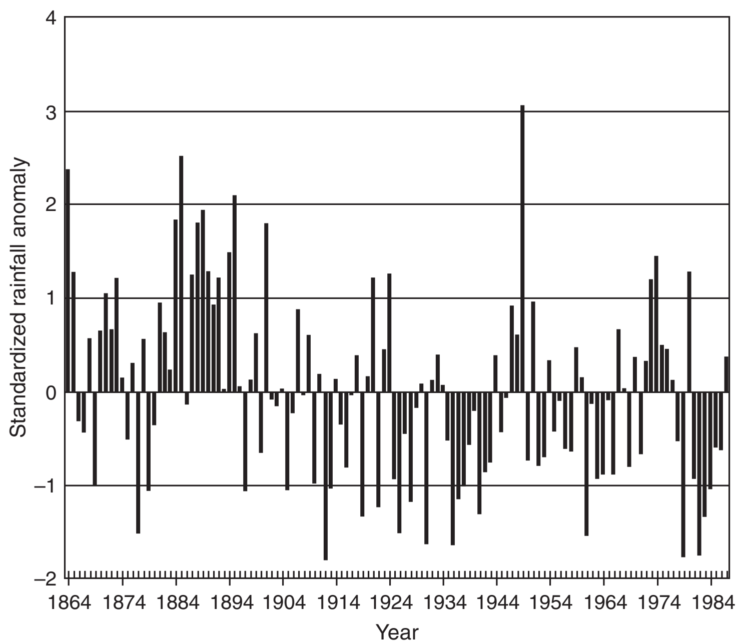
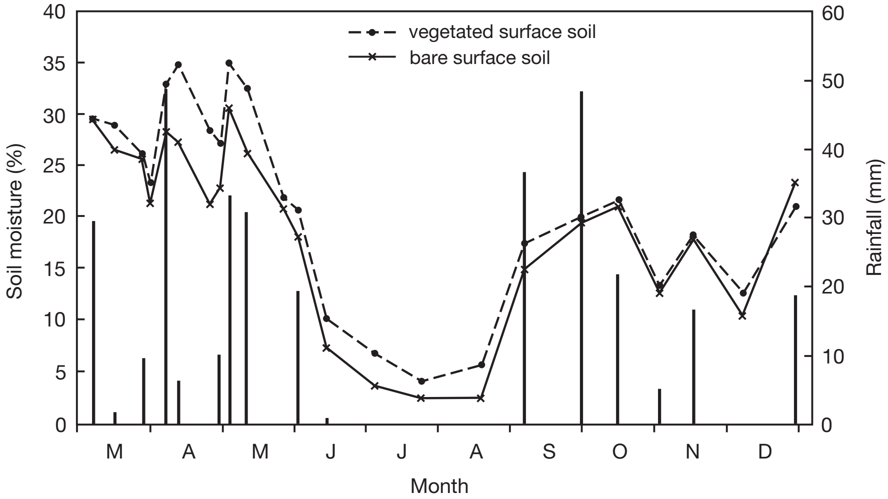


Fig. 11.2 Annual rainfall anomalies for Murcia, south-east Spain (after Thornes 1991)

Fig. 11.3 Seasonal fluctuations in soil moisture for the El Ardal site, Murcia, 1989-1990 under different ground conditions based on data for the MEDALUS project



In areas of extreme aridity the rainfall is not strongly seasonal in amount except in the Gobi–Tarim basin, where cyclonic storms are blocked by the persistent Siberian High in winter. Towards the desert margins rainfall seasonality increases and most significant is the distribution of rainfall in relation to temperature. Around the central Australian desert, for example, long periods without rain are usual in the northern winter and the southern summer. In the Mojave Desert rain occurs in winter, in the Chihuahuan Desert in the summer, and the Sonoran Desert has both winter and summer rainfalls.

Event rainfall magnitudes are important in determining soil moisture characteristics, but generally in desert areas it is the high intensity that is geomorphologically important, as discussed below.

The idea of coupling magnitude and frequency was first developed comprehensively in geomorphology by Wolman and Miller (1960). Put simply it draws attention to the fact that it is not only the magnitude of the force applied by a process but also how frequently the force is applied which determines the work done by the process. In temperate environments their results suggest that it is the medium scale events of medium frequency that perform most geomorphic work. Ahnert (1987) has used the same concept to express the magnitude and frequency of climatic events. Plotting the event magnitude on the vertical axis and the log (base 10) of the recurrence interval of the event on the horizontal axis provides an alternative representation of the relationship. By fitting a regression model to the data he obtained the relationship

$$P_{24} = Y + A \log_{10}(\mathrm{RI}) \tag{11.1}$$

where P_{24} is the daily precipitation of an event with recurrence interval of RI (in years), and the recurrence interval is obtained from RI $= (N + 1)$/rank by magnitude in a list of n daily rainfall totals. Ahnert showed that Y and A are characteristics of the rainfall, and he used them as indices of magnitude and frequency. The map of California (Fig. 11.4a) indicates that both A and Y are generally low in the Mojave Desert. The indices can be obtained for other time intervals, for partial series (such as P-threshold), and for other variables such as frost frequency. Moreover, it is possible to derive the magnitude and frequency product and plot this against return period. For example, Ahnert calculated the curve for $(P - T_r)F$, where P is the daily rainfall, T_r a threshold value, and F the frequency of the event of magnitude $(P - T_r)$, and plotted this against recurrence interval. The family of curves (Fig. 11.4b) shows that in desert areas the overall product is low and that the maximum of the product occurs in tens of years rather than several times a year. This implies that most work is done by rather less frequent events, but not by very rare events. De Ploey et al. (1991) developed this idea further to provide an index of cumulative erosion potential.

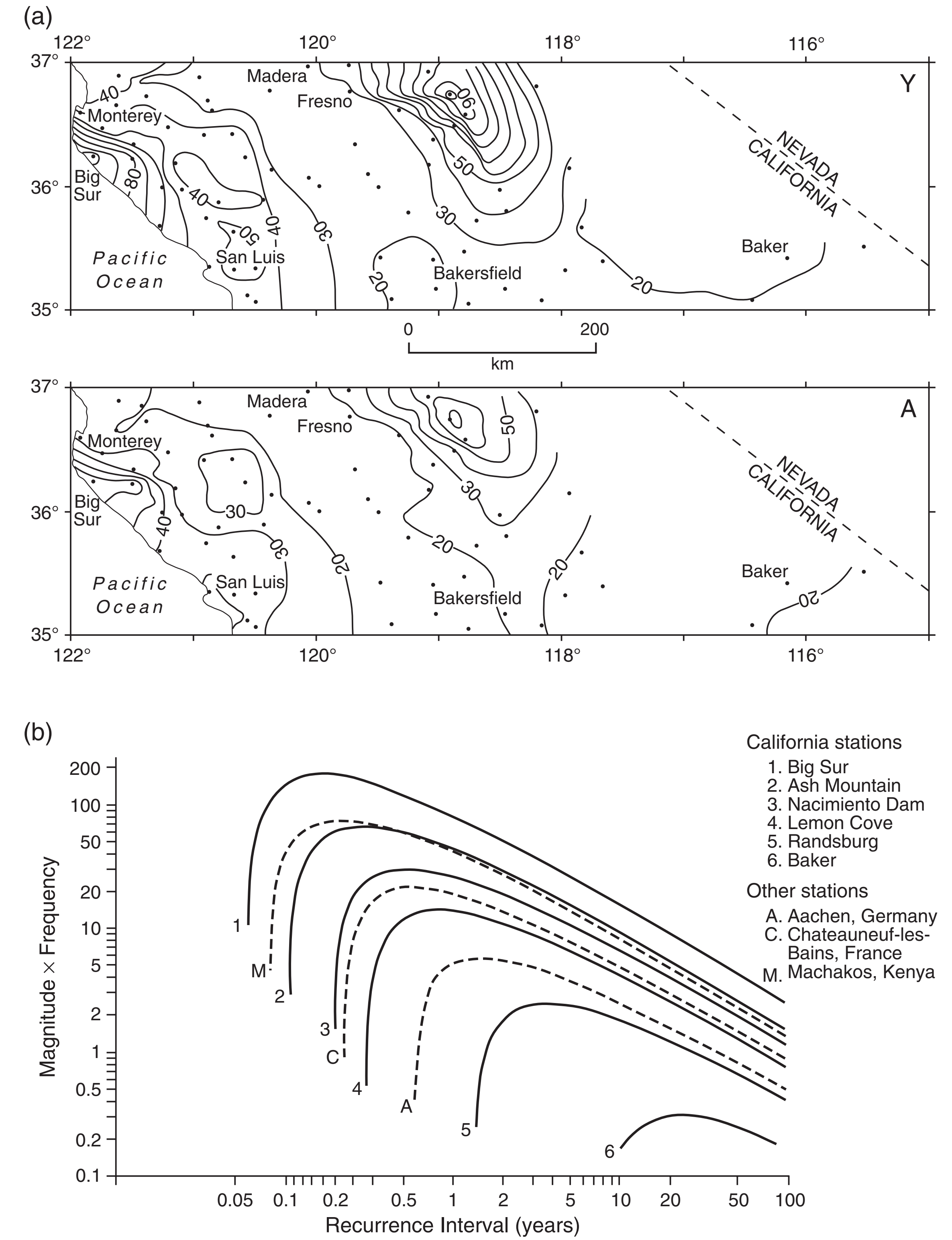


Fig. 11.4 Frequency and magnitude distribution of rainfall (after Ahnert 1987). (**a**) Regional distribution of values of *A* and *Y* in California. (**b**) The product of magnitude and frequency for a variety of stations

In temperate environments there are a large number of small events and a much smaller number of large events, so the distribution tends to negative exponential and gives a straight line on the semi-log plot. Ahnert (1987) used this relationship to explore the characteristics of storm events, and this provides a useful comparison of the models for desert and temperate environments (Fig. 11.4). This pattern of magnitude and frequency can also be expressed for individual seasons.

Rainfall Intensity

The intensity of rainfalls is especially important in determining the canopy storage and gross interception losses as well as the production of overland flow. Intensity also determines the compaction of soils and thus indirectly affects the infiltration capacity (Roo and Riezebos 1992). One of the simplest measures of intensity is the mean rainfall per rain day. It provides the key parameter for providing the exponential distribution of rainfall magnitudes and deriving the excess above a critical threshold value (Thornes 1976). It is sensitive to the definition of rain day, but generally a figure of 0.2 mm is taken as the threshold. Average figures for desert areas are between 5 and 10 mm, and these tend to increase to the margins. These figures do not differ greatly from temperate areas and are generally less in the extreme deserts.

While mean rainfall per rain day is a useful index when only daily rainfall amounts are available (which is the case for large parts of the Earth and especially desert regions), the absolute intensity (mm h^{-1}) generally increases as shorter periods are considered. Although record intensities occur in humid tropical areas, intensities for arid areas are generally higher than those of temperate regions. For example, Dhar and Rakhecha (1979), working in the Thar Desert of India where average annual rainfall is 310 mm, found that one-day point rainfalls of the order of 50–120 mm have a recurrence interval of 2 years and that rainfalls associated with monsoon depressions can be from 250 to 500 mm in a single day. The peak daily rainfall can be several times the annual average. Berndtsson (1987), for example, observed that on 25 September 1969, 400 mm of rain fell at Gabes, central Tunisia, which is about five times the mean annual rainfall. In semi-arid south-eastern Spain the storms of 18–19 October 1973 produced rainfalls equal to the mean annual amount (300 mm) in a period of about 10 h. In this semi-arid environment rainfall intensities of 70 mm h^{-1} typically recur about every 5 years.

Using the 1-h duration and 2-year return period, Bell (1979) obtained figures typically of the order of 10–20 mm for central desert areas and 20–50 mm along desert margins. Again, this suggests that it is the desert margins where runoff effects are most likely to be geomorphologically important. He estimated the 'maximum probable' 1-h rainfalls for the Australian arid region to be of the order of 180 mm at the poleward edge to 280 at the equatorwards margin. Similar figures have been estimated for the desert areas of the United States by Herschfield (1962), and Bell (1979) believed that they are probably typical of the arid zone in other parts of the world. Typically the 2-year 1-h value has an intensity about ten times that of the 2-year 24-h rainfall.

Temporal Variability

For geomorphic purposes most data are of relatively short duration. Typically 40–50 years of daily rainfall records are available for the arid areas of the world. Autographic data, from which storm profiles can be constructed, are even rarer. For developing models of the impact of rainfall series it is therefore sometimes necessary to simulate series which have the same characteristics as the actual data. The problem is that small sample sizes preclude normal forcasting techniques and the variability of total rainfall in an event is quite high.

Within individual storm events there are often significant variations in intensity. Figure 11.5 illustrates a storm profile for the semi-arid Cuenca area of central Spain. The total rainfall for the storm was 24 mm, with an average intensity of 6 mm h^{-1} while the high burst in minutes 35–39 produced an intensity of 36 mm h^{-1}. Within-storm rainfall has been modelled by Jacobs et al. (1988) and a description of their approach is found below.

The distribution of events is reflected in the seasonal distribution of cumulative rainfall totals and these have a strong effect on plant growth in dry areas. For

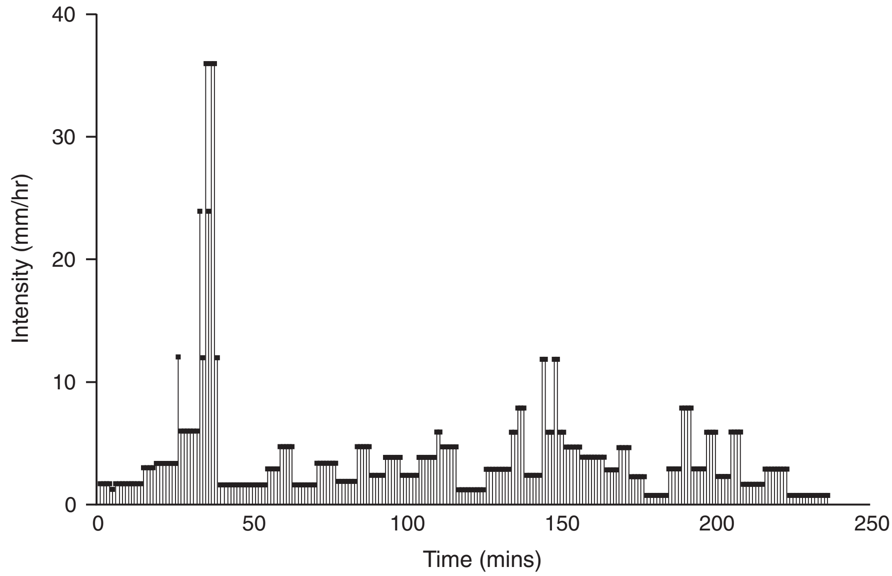


Fig. 11.5 Rainfall storm profile for an autumn storm in Cuenca, Spain

example Fig. 11.6 shows the contrasts in rainfall patterns within the year for normal (1981), exceptionally wet (1961) and exceptionally dry years (1979, 1982) in south-east Spain. Lack of autumn rains inhibits the germination of seedlings, whereas lack of soil moisture in spring limits plant growth.

The seasonal distribution of event magnitudes and durations is also important from a runoff and infiltration point of view. Generally event magnitudes and durations reflect prevailing weather types, with cyclonic storms often giving long-duration, low-intensity rainfalls and convective (summer) storms giving short-duration, high-intensity rainfalls.

A simple seasonal model is provided by the Markov chain, in which sequences of rain and no-rain days and the rainfall magnitudes for these days are generated. Thornes and Brandt (1994) have provided a synthetic generator for the semiarid regions of south-east Spain using this principle. The sequence of rain or non-rain days is obtained from a Markov probability matrix of transitions from rain-rain, rain–dry, dry–rain, and dry–dry days derived from the historical records on a sea-

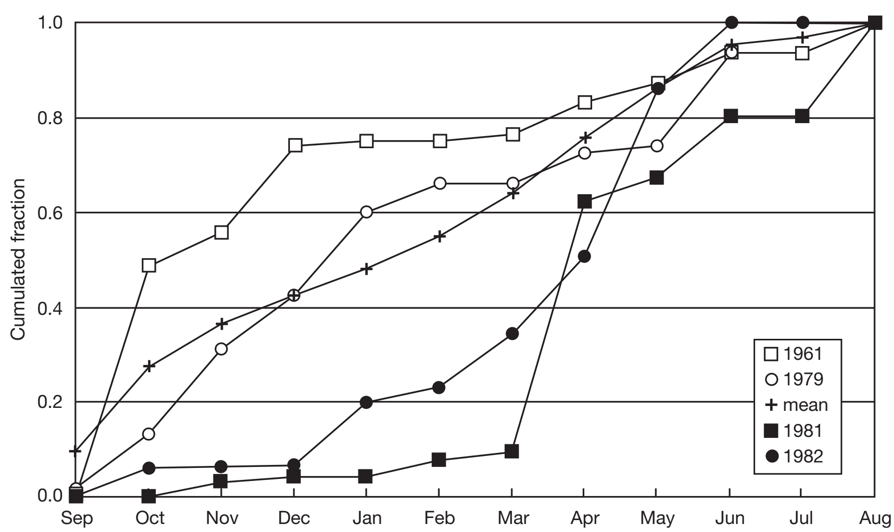


Fig. 11.6 Seasonal variations in the annual distribution of rainfall for the province of Murcia, Spain. Key: □ = 1961; ○ = 1979; + = mean; ■ = 1981; ● = 1982

sonal basis. The magnitudes are then obtained from a two-parameter gamma distribution, again with seasonal parameters. Both the transition matrices and the gamma parameters were found to be stable over time in this environment.

A more sophisticated model was used by Duckstein et al. (1979) to investigate and simulate the structure of daily series. They define an event as a day having a precipitation greater than a constant amount, say 0.2 mm. The statistics of interest are then the number of events per unit time, the time between events, the depth of precipitation, the duration of precipitation, and the maximum intensity of rainfall in an event. A sequence is a number of rainfall events separated by three or fewer days. Then the dry spell duration is a run of days without rain, and the average seasonal maximum of the run of dry days defines extreme droughts. Finally, interarrival time is the time between the beginning of one sequence and the beginning of the next; in statistical terminology it is the renewal time.

For convective storms occurring in summer the events are independent and short (cf. Smith and Schreiber 1973). The probability of n events of j days duration is then given by the Poisson mass function

$$f_n(j) = \exp(-m)m_{\mathrm{j}}/j! \tag{11.2}$$

where m is the mean number of events per season.

The probability that the interarrival time T is t days is given by

$$f_T(t) = u\exp(-ut) \tag{11.3}$$

where u is estimated from $1/\overline{T}$, with $\overline{T}$ the mean inerarrival time between events. For Tucson, Arizona, u is estimated to be 0.27. Under some conditions, especially where the storms are more frequent (e.g. where a cyclonic element prevails), interarrival times may be better described by a gamma rather than an exponential distribution. Finally, the event magnitudes are also independent. For example in Granada province, southern Spain, Scoging (1989) found that 90% of summer storm rainfall lasted less than an hour and that the magnitude of the rainfall in an event is independent of its duration. Under these circumstances, which are most common for convective thunderstorms, the gamma distribution again provides a suitable model for rainfall magnitudes.

The amounts, intensities, and durations of storms are also found to vary significantly in the long term in dry as they do in temperate areas, and interest in these fluctuations has been heightened in recent years as the result of the potential implications of global warming for already dry areas.

Berndtsson (1987), for example, found that in Tunis from 1890 to 1930 there was a massive fall in mean annual rainfall and that after 1925 there was an oscillation of durations of 10–15 years. Similar phenomena have been recorded in other statistical analyses. Conte et al. (1989) demonstrated that oscillations in rainfall in the Western Mediterranean in the period after 1950 could be related to the Mediterranean Oscillation in barometric pressure. Thornes (1990) also observed both the steep decline in rainfall in the period 1890–1934 and oscillations in rainfall anomalies in the post-1950 period in the southern Iberian Peninsula. The rainfall decline, at 3 mm y^{-1}, is of the order of magnitude currently predicted for climatic changes over the next 60 years by climatologists from general circulation models. The rainfall anomaly oscillations are in phase with those recorded by Conte et al. and almost exactly out of phase with those described for the Sahel.

Spatial Variability

If arid zone rainfall is difficult to predict in time, it is almost as difficult to predict in space. The spatial distribution of storms tends to be random in subtropical deserts. There is usually a poor correlation of rainfall amounts at stations only 5 km apart. It is rare for a single station to experience more than one storm in a day, and the spacing between concurrent storms is typically 50–60 km.

The development of rainfall in convective storms is quite complex and accounts for much of the locally highly concentrated rainfall amounts and intensities that accompany desert storms. Well-defined storm cells are born and then decay, and the cellular structure can give rise to well-defined rates of attenuation of rainfall with distance from the cell centres. Jacobs et al. (1988) examined summer data from a network of 102 gauges distributed over a 154 km^2 area in Walnut Gulch Experimental Watershed, Arizona for the period 1970–1977, and they attempted to model the intensity and cumulative depth over temporal realizations of the process or for a particular event over space. The procedure assumes that a storm is born at maximum intensity and then decays exponentially through time, with the initial intensity being an independent exponentially distributed random variable. The spatial distribution of storm centres is modelled as a Poisson process, and the time of birth of each cell relative to storm outset is taken to be exponential. From these assumptions the variance and covariance structures of rainfall intensity and depth are derived analytically and appear to provide a reasonable representation of the structure of the storm patterns observed.

In comparison with the Walnut Gulch data, one of the best sets of rainfall data in a desert area in the world, the storm depths modelled by Jacobs et al. were found to be more homogeneous and isotropic over time. Zawadeski (1973) had already shown that storms are typically isotropic under about 10 km but increasingly anisotropic over larger distances.

Interception

Our understanding of surface hydrological processes is very largely in terms of specific surfaces, such as bare soils, plants of various species and architecture, and more recently in terms of stone cover. In fact, desert surfaces usually comprise complex mixtures of these different types, just as satellite images represent complex mosaics of land use (Fig. 11.7). This means that both the understanding and modelling of such surfaces tend to be difficult. Moreover the very nature of desert soils with their chemical crusts, concentrated aggregates, complex surface roughness, and particular plant types means that hillslope hydrology (as opposed to hydraulics) of such regions is still in its relative infancy. Earlier treatises on desert hydrology, for example, have tended to simply reiterate the received wisdom from temperate soils, which is, in reality, completely inappropriate.

Desert plants are structurally rather different from temperate and humid tropical plants in their life forms. They often have low leaf areas, especially adapted leaves and stems, and wide spacing (see below). The morphologies of several rangeland species of shrubs and grasses (e.g. *Cassia* sp., *Stipa tenacissima*) are designed to intercept rainfall and channel the water down as stemflow. This water infiltrates adjacent to the tree bole and produces enhanced water storage at depth. Slatyer (1965) showed that with mulga scrub once the interception store of 2–3 mm has been filled, stemflow accounts for up to 40% of the rain falling on the canopy, and falls of 15–20 mm of rain can cause about 100 l of stemflow from adult trees. Stemflow equal to

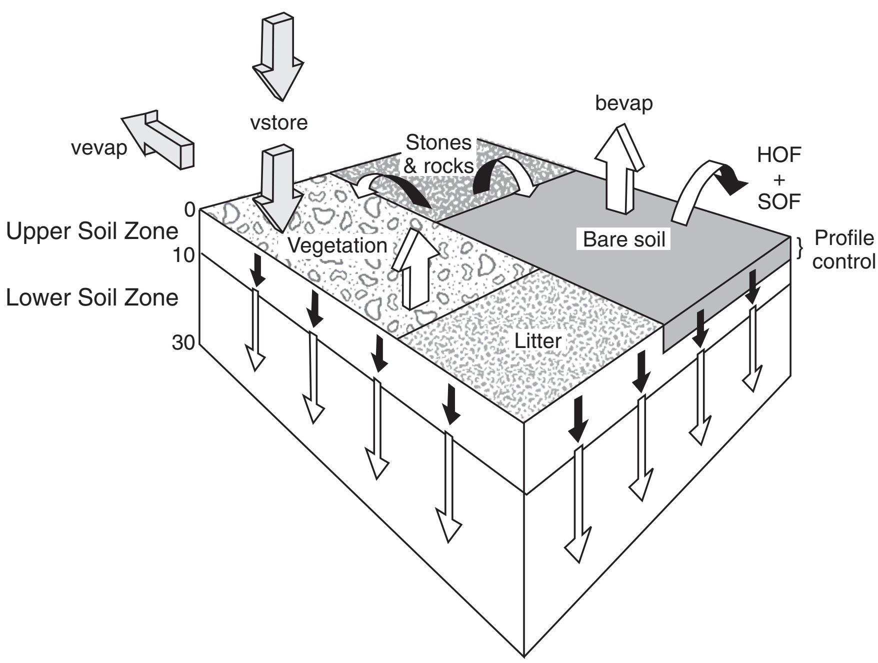


Fig. 11.7 Compound character of surfaces in semi-arid and arid environments, showing major fluxes of evaporation (vevap, bevap), transpiration (trans), interception (vstore) and Hortonian and saturated overland flow (HOF, SOF)

30–40% of intercepted rain has also been found in semi-arid environments in the Zaragoza area of Spain by Gonzalez-Hidalgo (1991).

Evans et al. (1976), on the other hand, reported that about 90% of rainfall was converted into throughfall for creosote bush and about 65% for bursage. However, the percentage for different storms ranged from 67% to greater than 100% and from 26% to greater than 100% for creosote bush and bursage, respectively. Generally, in rangeland, losses from the canopy through the direct evaporation of intercepted water are thought to be unimportant (Johns et al. 1984). However, though the total loss to gross interception may be small, its role in redirecting water to the concentrated root zone below the stem may be very significant. The concentration of nutrients in the root zone is partly a testimony to this concentration of wash from the canopy itself. De Ploey (1982) has attempted to provide a simple model to estimate the interception properties of plant cover on the basis of stem density and stem angle and has simulated the behaviour of stems in the laboratory with a reasonable degree of success.

Stone cover is also important in determining the interception of water, but the picture is somewhat confused for two reasons. First, losses to infiltration are estimated as the residual between water rate application by sprinkler and runoff from experimental micro-plots. Second, the loss rate is influenced by whether the stones are lying on the surface or embedded within it (Poesen et al. 1990). Our field experiments indicate that stones may intercept and store considerable volumes of water depending on their surface characteristics. Therefore, they probably account for much of the losses attributed to infiltration in sprinkler experiments. This question is further discussed below. The same is true of crusting. As yet we do not know the relative proportions attributable to interception, infiltration, and evaporation in determining the residual sum available for runoff.

Unsaturated Zone Processes

Soils in deserts are as variable as, if not more variable than, their temperate counterparts. The effects of soil texture are emphasized, as profile control is dominant in the overall water balance. For example, Hillel and Tadmoor (1962), in a comparison of four desert soils in the Negev, showed that sandy soils wetted deeply, have the largest storage, the highest infiltration rates, and the lowest evaporation when compared with rocky slope soils and loessic and clay soils. The latter had very low storage as a result of surface crusting, poor absorption, and high vaporation rates. As a result, the plant growth capabilities were markedly different. Others too have shown that in sandy soils the upward movement of water vapour, facilitated by the open pore structure, may lead to condensation of water near the surface in winter-cold desert soils in amounts of the order of 15–25 mm.

A major driving force in desert soils is the high rate of near-surface evaporation. Water moves up through the soil as a result of the thermal gradient mainly through vapour transfer (Jury et al. 1981). With available moisture the near-surface potential rates may reach 1500–2000 mm y^{-1}, at 2 m the rates are about 100 mm y^{-1}, and at 3–4 m they are practically negligible. As a result of this steep gradient, upward movement of water and, if available, salts is the dominant process in summer months, as in the *takyr* soils of Turkmenistan.

The stone content of desert soils generally increases with depth, though such soils may also have surface armouring induced by a variety of processes Chapter 5). Field bulk density may be quite high for stoney soils, typically of the order of 1.8 compared with values of 1.0–1.4 for non-stoney soils. Mehuys et al. (1975) found little difference between stoney and non-stoney soils in the relationship between conductivity and soil tension, but strong differences in the relationship between conductivity and soil moisture. In two out of three cases the conductivity is higher in stoney than in non-stoney soils with milar soil moistures.

Infiltration

Infiltration rates on bare weathered soils in dry environments are often relatively high when surfaces are not compacted, especially where vegetation or a heavy surface litter occurs. Runoff is often critically determined by the controls of infiltration, especially vegetation at plot scales (Francis and Thornes 1990). In experiments using a simple sprinkler system at rates of 55 mm h^{-1}, with an expected recurrence interval of about 10 years, infiltration rates may be as high as 50 mm h^{-1} with runoff accounting for as little as 2–3% of the rainfall. However, under all circumstances the infiltration rates are highly variable.

In general, infiltration rates increase with litter and organic matter content of soils. One of the earliest investigations, by Lyford and Qashu (1969), showed that infiltration rates decreased with radial distance from shrub stems due to lower bulk density and higher organic matter content under the shrubs. Later investigations by Blackburn (1975) on semi-arid rangeland in Nevada examined the effects of coppice dunes (i.e. vegetation bumps) on infiltration, runoff, and sediment production. Using a sprinkler simulator at 75 mm h^{-1} on small plots, he analysed the average infiltration rates at the end of 30 min and found that the coppice dune rate was three to four times that for the dune interspace areas. The soil characteristics of the interspaces were most important in determining runoff rates, and infiltration rates were strongly negatively correlated with percentage bare ground. Similarly, Swanson and Buckhouse (1986), in a comparison of three subspecies of big sagebrush (*Artemisia tridentata, A. wyomingensis*, and *A. vaseyana*) occupying three different biotopes in Oregon, found no significant differences in final infiltrability between habitat type and climax understorey species. However, infiltration in shrub canopy zones had generally higher rates than shrub interspaces, confirming the results obtained by Lyford and Qashu (1969) and Blackburn (1975).

There is often an interaction between infiltration capacity, vegetation cover and aspect in dry environments. For example, Faulkner (1990), examining infiltration rates on sodic soils in Colorado, has found that aspect has a strong influence on infiltration rates through vegetation cover, though some caution needs to be exercised here because of the role of piping in infiltration rate and runoff control in the bare badland areas.

Gifford et al. (1986) examined the infiltrability of soils under grazed and ungrazed (20 years) crested wheatgrass (*Agropyron desertorum*) in Utah using both sprinkling and ring infiltrometers. They found that infiltration rates measured by the ring infiltrometer were 2.3–3.2 times higher than those measured by the sprinkling infiltrometer (cf. Scoging and Thornes 1979) and that season and grazing/nongrazing were the main sources of variation. The spring infiltrability values were double the summer values, and the ungrazed were three times the grazed values.

Infiltration rates are generally reduced by crusting, which may result from algal growth, rainfall compaction, and precipitate development. Thin algal crusts have an important role in stabilizing the soil surface, though there seems to be some disagreement as to its hydrological significance. Some workers claim that the crust, by limiting the impact of raindrops on a bare soil, actually prevents the development of a true rainbeat crust and thus the algae increase infiltration. A contrary view, however, is that the biological crusts have a strong water repellency and so increase runoff. Alexander and Calvo (1990) and Yair (1990) have investigated the organic crusting and attest to its significance in controlling surface runoff. Alexander and Calvo found that lichens provide crusting on north-facing but not on south-facing slopes and that lichens produce more rapid ponding and runoff but not more erosion. Yair reported that removal of the crust from sandy dune soils produced a dramatic increase in the time to runoff and that the runoff coefficient was an order of magnitude lower, largely due to reinfiltration.

On the other hand, compaction by raindrop impact leading to sealing significantly reduces infiltration (Morin and Benyami 1977). Roo and Riezebos (1992) have attempted to provide a model for the impact of crust development on infiltration amounts.

Finally, stones control the infiltration rates into soils to some extent. The general conclusion of the many field and laboratory studies on infiltration rates under a rock cover indicates a generally positive correlation between infiltration and stone cover, but Poesen et al. (1990) argued that this depends on whether the rocks are partially buried or resting on the surface. In the former case there tends to be a negative correlation because the stones intercept the rainfall and lead to higher evaporation and runoff rates, whereas in the latter case they tend to inhibit crusting and so encourage infiltration. Working on shrub lands, both Tromble et al. (1974) and Wilcox et al. (1988) obtained negative correlations between infiltration and stone cover, but the latter accounted for this in terms of the reduced opportunity for plant cover development, which enhances infiltration as noted above. In a field experiment Abrahams and Parsons (1991) also obtained a negative correlation between stone cover and infiltration rate and again explained this in terms of the higher infiltration rates under bushes where stones were generally absent (see Fig. 9.11). They concluded that, at least for shrub covered pediments, inverse relations between infiltration and stone cover could be used in modelling.

Evaporation from Surfaces and Transpiration from Plants

Most literature for dry environments suggests that there is a high correlation between evapotranspiration and rainfall (Scholl 1974) and that in general the total annual evapotranspiration for desert environments is equal to the cumulative annual infiltration into the soil for the year. This is more or less equal to the annual precipitation minus the annual runoff. Moreover, at least once per year the soil moisture becomes more or less negligible and deep drainage is more or less negligible relative to total annual evapotranspiration. Therefore, the principal interest is in the temporal distribution of evapotranspiration during the year.

Actual estimates of the relative balance of the various components give a clear idea of the importance of evapotranspiration in the overall water budget. Renard (1969) suggested that evaporation and evapotranspiration from Walnut Gulch is about 85–90% of the incoming rainfall, and in the study by Floret et al. (1978) in Tunisia evaporation from bare soil alone was calculated to be between 40 and 70% of the total depending on the year. Evans et al.'s (1981) comparative analysis of bare soil and different plant species (creosote, mesquite, and sagebrush) revealed no significant differences between the bare soil and the vegetated sites for any extended period of time. Nor were there any differences between plant species, and most of the evapotranspiration took place in periods immediately following rainfall. Then the rates varied between 2.3 and 10 mm d^{-1}, with rates of 1 mm d^{-1} for extended periods of time between rainfall events and rates of less than 0.1 mm d^{-1} for the driest parts of the year.

These results emphasize the ability of desert plants to exist at high water stresses. Final water contents correspond to suctions of −10 to −50 bars. Typically the plant stress is uniformly distributed throughout the plant at dawn and in equilibrium with the soil suction. As the day develops the plant stresses increase so that for *Larrea tridentata stresses* of 40–65 bars are reached, and in *Ceratoides lanata*, for example, 120 bars potential is not uncommon.

The evapotranspiration process is clearly crucially important for dry environments, even when vegetation cover is small, and exercises a significant control on the total water budget. Since this budget is controlled by the plant cover, which also plays a major role in the hydrology and hydraulics of overland flow, we return to the vegetation cover next.

A general classification of xerophytic plants is given in Fig. 11.8 (Levitt 1972). Xerophytic plants can escape drought seasons by an ephemeral habit or resist drought either by avoiding it by reducing water consumption or by using the maximum possible when it is available. The latter strategy is a characteristic especially of phreatophytes, such as mesquite and tamarisk. Other plants are able to tolerate drought by mechanical means. Mosses and lichens, for example, can expand again after dehydration. Such poikilohydrous plants have the capacity to take up water

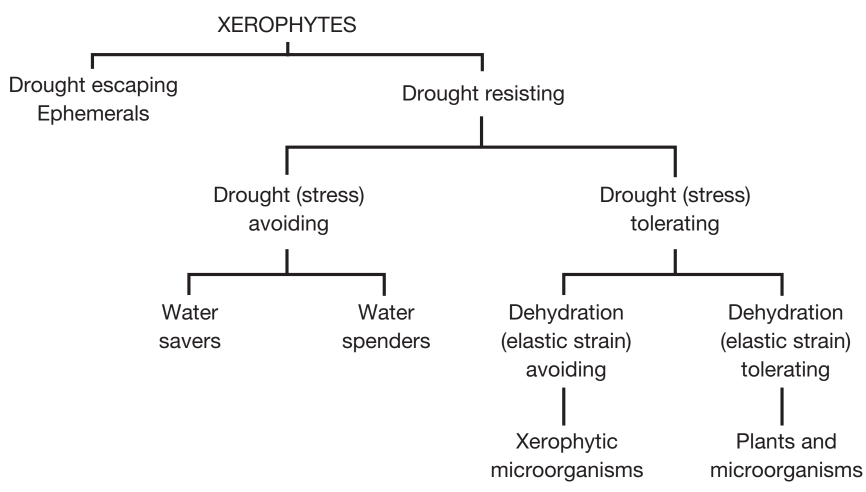


Fig. 11.8 Structure of xerophytic plant types (after Levitt 1972)

instantaneously and survive extreme and prolonged desertification.

Specific observations on desert vegetation have been made by Nobel (1981) in the Sonoran Desert. He investigated the desert grass *Hilaria rigida*, which is a clump grass of the C4 type and has amongst the highest photosynthetic rates so far reported. In this species, clumps of increasing size tend to be further from a random point and further from their nearest neighbour of the same species. This larger spacing suggests pre-emption of groundwater. This conclusion is partly borne out by the fact that leaves on small clumps (less than 10 culms) have a threefold higher rate of transpiration per unit leaf area than do leaves on larger clumps (over 200 culms), indicating that the water transpired does not vary much with clump size. Given that the amount of CO_2 fixed per unit ground area does not differ much with clump size, the pattern of large and small clumps can be quite stable. This may also be indicated by the low interdigitation of root systems for different clumps. This high level of adjustment suggests that clump life could be quite long. Somewhat similar conclusions were reached by Phillips and McMahon (1981).

Some species are evergreen, while others are facultatively drought deciduous. Several studies have indicated that the leaf area index (LAI) declines as water potential declines, and Woodward (1987) has developed a model which predicts LAI in terms of a cumulated water deficit, arguing that when the water deficit falls below −80 mm abscission begins (Fig. 11.9).

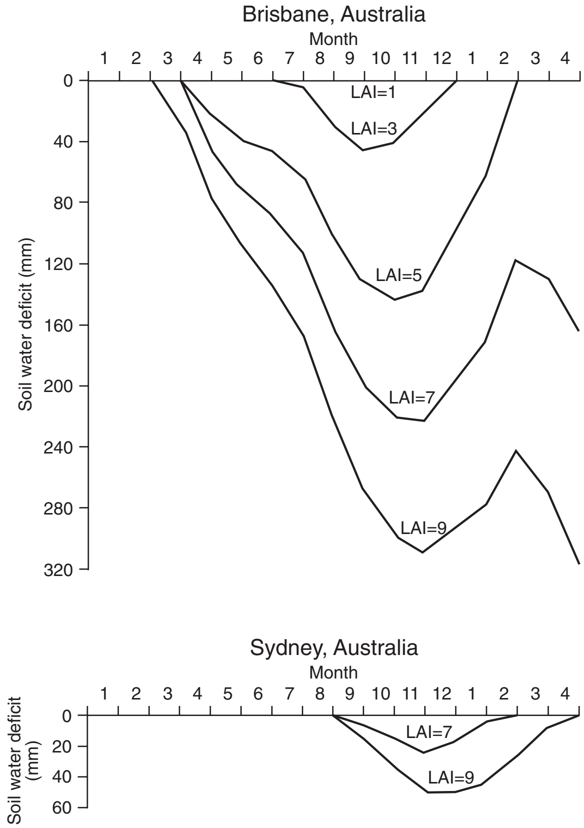


Fig. 11.9 Predicted monthly soil-water deficit for soils at two sites in Australia in accordance with Woodward's (1987) model. LAI = leaf area index (after Woodward 1987)

In addition to changes of leaf shape and plant structure, other special plant physiological adaptations occur which give desert plants more efficient water use. For example, succulent plants possess pathways for the fixation of CO_2 that enable assimilation to proceed with the stomata open only at night, when potential for evaporation from cladodes and leaves is minimal (Fuchs 1979).

The plant cover and biomass is a function of the production and distribution of the photosynthetic production, and this is usually controlled by the evapotranspiration rate. In modelling the impact of climatic change on erosion in which the plant cover is a most critical variable, the key issue in arid and semi-arid areas is the plant water balance and its impact on net production (i.e. the difference between the plant material produced by photosynthesis and that consumed for maintenance). It is generally assumed that the plant biomass comes into equilibrium with the available water (cf. Eagleson 1979, Thornes 1990, and see below).

Lane et al. (1984) demonstrated for the Mojave Desert that the best predictors of annual plant productivity were linear relations with mean annual precipitation ($r^2 = 0.51$), seasonal precipitation (January–May, $r^2 = 0.74$), annual transpiration ($r^2 = 0.84$), and seasonal transpiration ($r^2 = 0.90$).

Net plant productivity in the south Tunisian steppe is approximately 1000 kg ha^{-1} y^{-1} of dry matter or approximately 5.5 kg ha^{-1} mm^{-1} of rainfall. Wet years produced 1550 kg of dry matter per hectare and dry years about 1060. The difference is due to the distribution of rainfall. This distribution is especially important when the annuals and perennials occur at different times of the year. Floret et al. (1978) found that annuals peak in May, while perennials peak in June and July. For maximum production of annuals, rainfall should be regularly distributed between autumn and the end of spring, while bushes do better if the deep soil layers are still wet in mid-May.

Modelling vegetation growth and water consumption for sparse vegetation can be illustrated by the ARFEQ model of Floret et al. (1978). These authors modelled primary production and water use in a south Tunisian steppe, comprising dominantly *Rhatherium suaveolus*. This is a zone of typically 170–180 mm y^{-1} rainfall, falling mainly in September and March. The soil is a gypsous sandy clay with an organic matter content of the order of 0.4–0.7%. The cover of perennial species is about 35% *Stipa* sp., *Aristida* sp., and *Plantago albicans*. *Rantherium* accounts for a further 30% with about 26 500 tussocks ha^{-1}. The surfaces typically have 1600 kg ha^{-1} of above-ground phytomass and 50–80 kg ha^{-1} of litter on the soil.

In the ARFEQ model bare soil drainage is an inverse function of the storage above field capacity; potential evapotranspiration is calculated as a function of the measured Piche evaporation; and actual evapotranspiration is related to potential through the dimensionless water content of the water content of the upper horizon. For the plant cover, total transpiration is obtained from a comparison of the atmospheric demand and the soil-water stress. The atmospheric demand is expressed in terms of LAI, stomatal conductance, and Piche evaporation. The soil-water resistance is obtained from the difference between soil-water stress and leaf stress mediated by a root resistance, which is inversely proportional to the amount of root mass in a given soil layer. Leaf-water potential is then assumed to meet the requirement that there is an equilibrium between the atmospheric demand and the root extraction. This is a common assumption for water-limited plant covers (i.e. that the plant water use is in equilibrium with the applied stress). In practice this assumption may often be doubtful given the high interannual and longer-term variability. Below we examine how changes in water availability and redistribution lead to vegetation cover extinction and, hence, to changes in the hydrological regime over the longer term.

From these hydrological considerations, gross photosynthesis is assumed to be a function of water status, LAI, and incoming solar radiation (Feddes 1971), and the photosynthetic rate less respiration and senescence is used to determine dry matter production. The results of the model suggest that approximately 0.26–0.54 g of dry matter are produced per kilogram of H_2O, rising in spring to 1.26 for annuals and 1.39 for perennials.

While still in the domain of unsaturated control, it is worth noting that there have been significant new developments in estimating the impact of large areas of bare surface on the now widely accepted Penman–Monteith equation (Monteith 1981), though the developments are still largely restricted to agricultural crops, such as millet. The Penman–Monteith model is based on the estimation of potential evapotranspiration controlled by atmospheric demand through the resistances of plant stomata (stomatal resistance), the canopy overall, and the atmosphere above the canopy to water vapour transport. However, it has been observed recently that a substantial amount of water is lost from the surface by evaporation from dry soils. For example, Gregory (1991) reported that as much as 75%, and commonly of the order of 35%, is lost by evaporation from bare soil surfaces between row crops. There seem to be two possible approaches to dealing with this problem: the first is to have separate estimators for crops and bare areas; and the second it to have a combined model which takes both into account. With regard to the first approach, Ritchie (1972) developed a model which assumed that after rain, soil evaporation is determined by net radiation at the soil surface. However, following this initial phase, evaporation from the bare soil becomes increasingly limited by the ability of water to diffuse through the drying soil surface. In this second phase, Ritchie set cumulative soil evaporation to be inversely proportional to the square root of time. Many observations of desert soils have empirically supported the use of this proportionality even though this model allows no interaction between the soil and the canopy. Two particular models are important in the second approach: ENWATBAL developed by Lascano et al. (1987) and the sparse crop model of Shuttleworth and Wallace (1985). In the latter the near-surface layer (Fig. 11.10) comprises an upper layer, which loses moisture directly from the canopy, and an interactive soil substrate, which has its own surface resistance and an aerodynamic resistance between the soil and the canopy. The combined evapotranspiration is then set up as:

$$LE_c + LE_s = C_c\mathrm{PM}_c + C_s\mathrm{PM}_s \qquad (11.4)$$

in which L is the latent heat of vaporization, E_c the canopy transpiration, E_s the bare soil evaporation, and

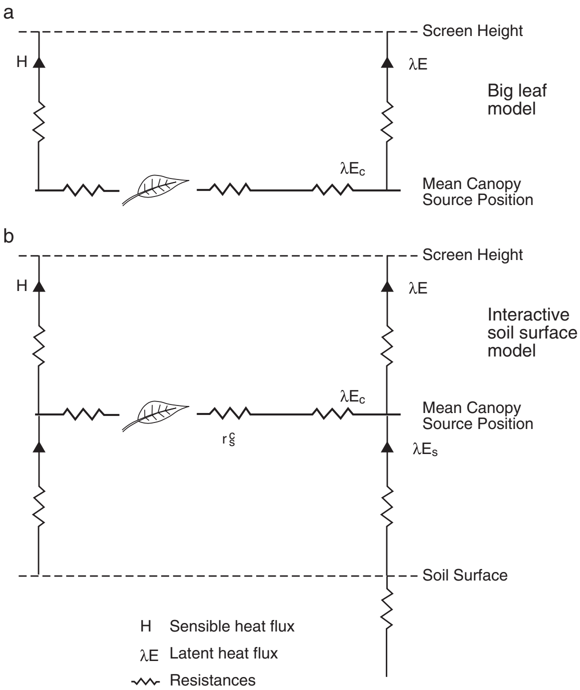


Fig. 11.10 The Shuttleworth and Wallace (1985) model of canopy and bare soil: (**a**) classical 'big leaf' model, (**b**) interactive soil substrate model. *E* is total latent heat flux, with subscripts c and s representing canopy and bare ground, respectively, and *r* is the canopy resistance

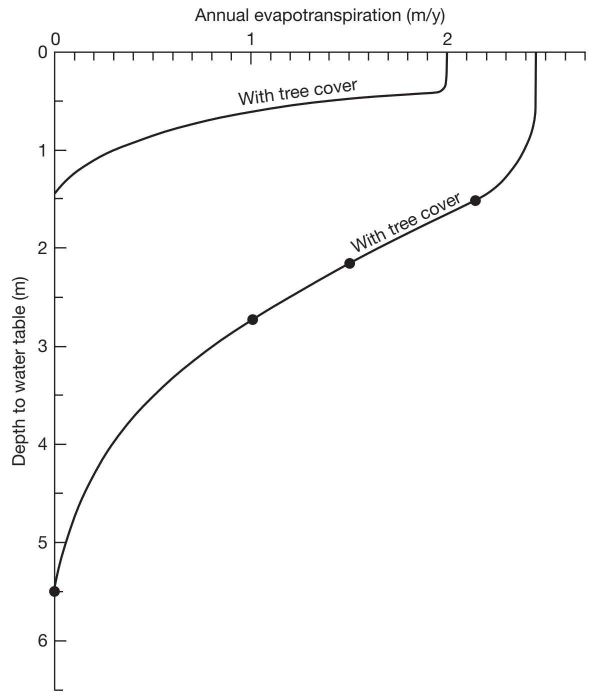


Fig. 11.11 Gila valley phreatophytes water loss by evapotranspiration over depth, with (*solid lines* with *circles*) and (hypothetically) without salt cedar cover

PM_c and PM_s the respective Penman–Monteith estimates. The coefficients C_c and C_s are then empirical coefficients which are functions of the aerodynamic resistance and bulk stomatal resistance of the crop and the soil (Wallace 1991).

Phreatophytes

It was argued in the 1960s that phreatophytes in the American South-west were major consumers of water and that the elimination of phreatophyte vegetation might be considered as a suitable method of water harvesting. Bouwer (1975) considered that potentially 6.4 million hectares of phreatophytes could be consuming $30 \times 10^9\,m^3$ of water. The uptake of groundwater by plant roots varies diurnally and seasonally, but fundamentally the evapotranspiration rate appears to be related to the depth of water table. Bouwer (1975) has modelled it as such on the assumption that the system is in steady state. The deeper the roots the greater the water table can sink and still maintain essentially potential evapotranspiration rates. Because roughness and total leaf area tend to increase with increasing plant height, and trees and shrubs are generally deeper rooted than grasses, potential evapotranspiration for deep-rooting vegetation is higher than that for shallow-rooting plants, and higher still than that for bare soil.

Van Hylckama (1975) has estimated the water loss with and without salt cedar cover in the lower Gila valley, Arizona (Fig. 11.11). His analysis suggests that with salt cedar the surface groundwater rate would yield about $2.5\,m\,y^{-1}$ of water loss with trees and $2.0\,m\,y^{-1}$ without. This difference increases sharply as the water table falls. Bouwer, however, noted that the reduction in channel seepage due to phreatophyte control increases and the actual amount of water which can be saved decreases as distance from the channel decreases.

Hillslope Runoff

In drylands overland flow is patchy in occurrence, reticular in character, and often terminates before reaching the channel by reinfiltrating on the hillslope

(Yair and Lavee 1985). Its characteristics are largely determined by the disposition of stones, vegetation, and macroscale roughness. Conversely, it plays an important role in the location of plant activity at the drier margins, and there may be a complex interaction between overland flow, reinfiltration, plant cover, and erosion. This reticular flow has been the subject of much recent research, since the important observations of Emmett (1970). Abrahams et al. (1986), Thornes et al. (1990), Dunne et al. (1991), and Baird et al. (1992) have all drawn attention to the impact of surface topographic stones and vegetation in concentrating the flow into rills and eventually gullies. Reticular flow contrasts significantly with sheet flow. In the past overland flow has often been modelled as sheet flow by means of the kinematic cascade model, which assumes that an overland flow of uniform depth is propagated down a hillslope comprising a series of smooth planes. The significance of reticular flow lies in the restriction of flow-induced infiltration and wash erosion to the actual flow-wetted area at the surface, which in turn has important implications for a wide range of processes. In Chapter 9 the generation and propagation of reticular runoff in hydraulic terms is discussed in detail. Here we restrict ourselves to the main controls on the quantity and timing of slope runoff in terms of the water reaching the channel system.

On slopes mantled with coarse debris, the stone content exerts a substantial control on hillslope runoff process. Yair and Lavee (1974) examined runoff from coarse scree slopes using rainfall simulators at three intensities: 60 mm h^{-1} for 10 min, 30 mm h^{-1} for 10 min, and 15 mm h^{-1} for 20 min. In these experiments the runoff was generated mainly in the lower areas (i.e. within the gullies). The explanation given was that the larger blocks (15–20 cm) in these areas yield more runoff and concentrate the water on small underlying patches of fine-grained materials, thus exceeding the infiltration rates in these fine-grained receiving areas.

Stones, like vegetation (Fig. 11.12), concentrate the flow so that velocities are higher, infiltration localized, and scour restricted in space. Baird et al. (1992) have

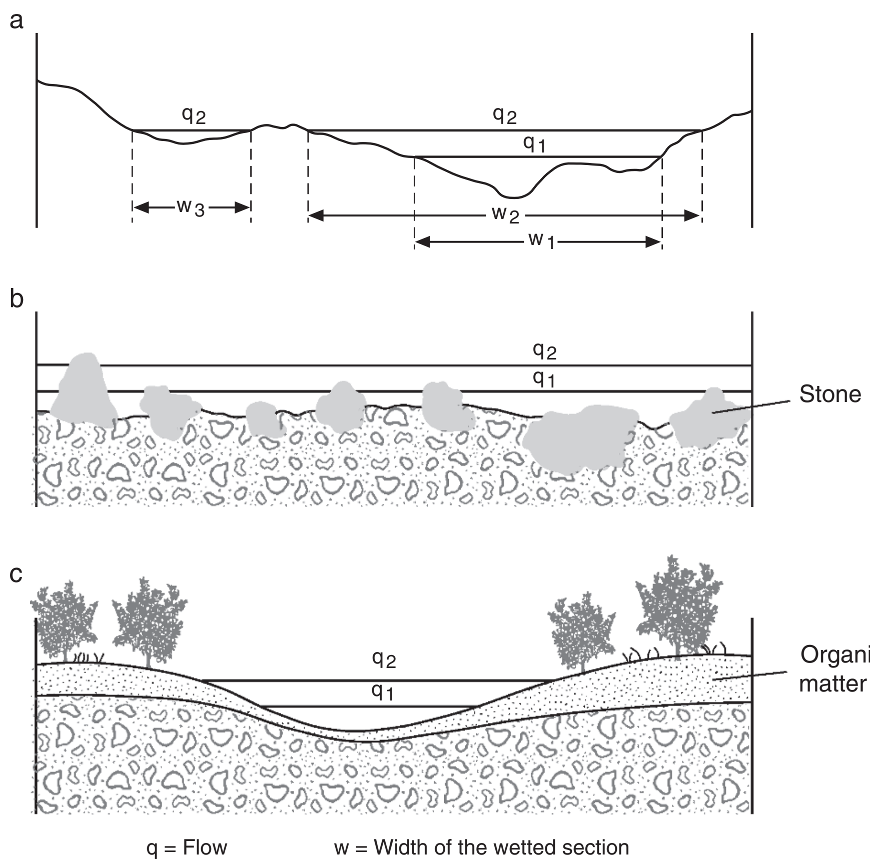


Fig. 11.12 Diagrams illustrating roughness elements. In (**a**) rills provide different flow depths and widths. In (**b**) the roughness elements are formed by surface rocks. In (**c**) the roughness elements are created by bushes and generate different infiltration amounts according to the thickness and disposition of organic matter

modelled this process following the free volume concept of Thornes et al. (1990). In their model the rainfall is assumed to be distributed across the entire surface but water entering a slope section is allowed to fill the hollows to a depth equivalent to its volume. Assuming a normally distributed ground roughness, they avoid describing the detailed topography in each cell and generate a distribution of flow velocities on the basis of depth and slope distributions for the wetted areas. As the flow increases in depth, it drowns surfaces of different infiltration properties and infiltration rates evolve dynamically for different flow conditions. Results of the operation of this model show significant changes in the hydrograph and in sediment yields when compared with unrilled slopes (Fig. 11.13).

At the hillslope scale, Yair and Lavee (1985) demonstrated a complex interaction between runoff generation, slope properties, and vegetation cover on slopes in the northern Negev at Sede Boqer. They observed that flow from bare upper-slope rock surfaces passes on to lower slopes with increasing amounts of colluvium. There is a progressive diminution of water availability to plants further downslope, resulting in a change in species composition as well as plant cover density. In the upper part of the colluvium, perennial species are dominated by Mediterranean species, whereas lower down Saharo-Arabian species with a lower water demand dominate.

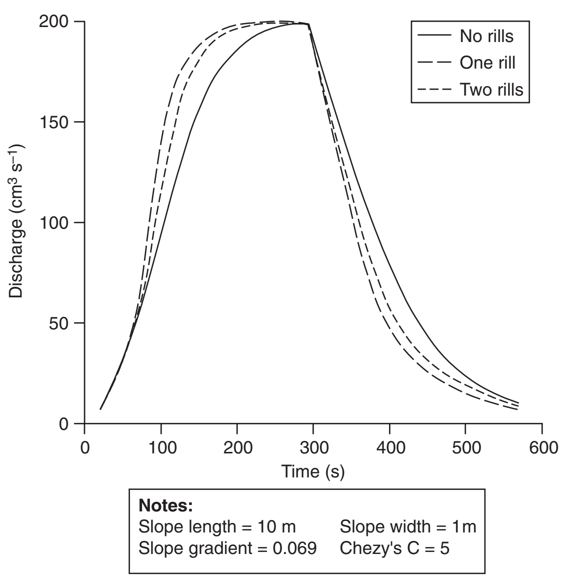


Fig. 11.13 Results from the RETIC model showing that the existence of rilling tends to steepen-up both the rising and receding limbs of the hydrograph

The distribution of species according to water availability and erosion competition was modelled by Thornes (1990) on the assumption of a wedgeshaped colluvium of increasing thickness downslope. This model yields an accumulation of water with increasing distance downslope, common in more humid Mediterranean environments, so that plants with a lower water use efficiency are located at the base of the slope (Fig. 11.14). Kirkby (1990), on the other hand, modelled the actual runon and reinfiltration conditions at Sede Boqer using a digital simulation model. Here the microtopography is generated by an exponential distribution, infiltration and evaporation follow storage-controlled laws, and water is routed kinematically. Some typical results are shown in Fig. 11.15a, giving the simulated overland flow for the bedrock surface and soil-covered sections and illustrating the pronounced effects of reinfiltration in the latter. The effects of redeposition are also felt so that the peak of erosion is in mid-slope, the net effect being to sweep the sediment in a downslope direction through time, leading to a redistribution of the whole complex of evaporation, recharge, and plant growth in that direction, which is offset only by the deposition of loess on the slope. The effects of the progressive movement of the system downslope are illustrated by the changes in the estimated relative plant densities through time shown in Fig. 11.15b. After 100 years the peak densities are predicted at about 28 m and by 500 years at about 42 m. At a given time the rise in plant cover continues downslope to a peak until no more water is available for evapotranspiration, after which it falls again dramatically in a downslope direction (Fig. 11.15b). The effect of this sweeping of sediment is to concentrate the vegetation in a narrow band in the zone of maximum infiltration losses. This peak vegetation cover actually moves slightly upslope through time. This complex interaction between vegetation, infiltration, and erosion at the hillslope scale is thought to replicate the actual conditions described for Sede Boqer quite well.

Much of the observation and modelling of dryland processes has been at the point, plot, or hillslope scale, and the problem of scaling up these observations and models to a regional scale has hardly been mentioned. An exception to this is the work of Pickup (1988) in Australia. Although this work is essentially directed at the problem of grazing-induced erosion, through the idea of scour–transportfill (STF) sequences, the

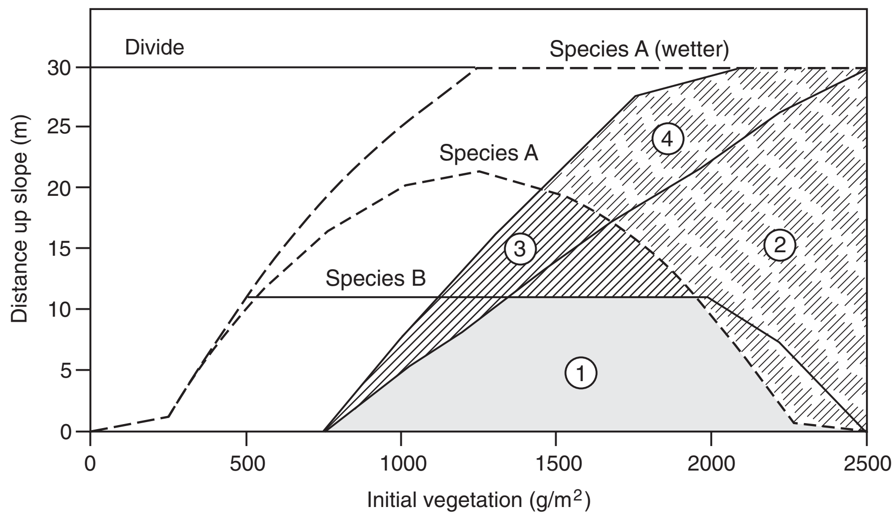


Fig. 11.14 The distribution of two species competing for water on a slope as a function of the initial vegetation cover. Species B is restricted to the lower part of the slope (below 12 m) almost regardless of the initial conditions. Species A can extend further upslope and under wet conditions can reach the divide, provided that the initial vegetation exceeds about 1200 g m^{-2}. In zone 1 both species are unchanging in biomass. In zone 3 (with diagonal hatching) species A is stable under dry conditions but species B will decrease. In zones 2 and 4 species A is stable and of unchanging biomass under wet conditions, but unstable and decreasing under dry conditions (after Thornes 1990)

approach is essentially hydrological in character. The spatial organization of runoff controls was also examined in a semi-arid mountain catchment by Faulkner (1987). This work also serves to remind us of the importance of snowmelt in semi-arid mountains as well as cold deserts. The hypothetical model of domains is shown in Fig. 11.16a, where the relative dominance of different processes is dependent on aspect and vegetation response. Faulkner proceeded to simulate the runoff production from the different surfaces in terms of snowmelt and rainfall runoff generation using established techniques. Figure 11.16b shows not only the importance of snowmelt on the hydrograph but also the different spatial impact its contribution has on the pattern of channel runoff.

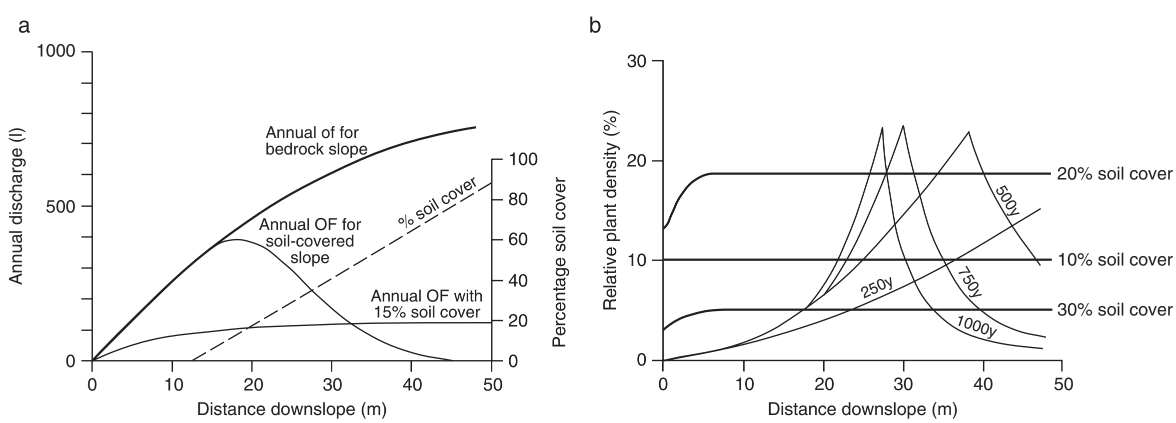


Fig. 11.15 Kirkby prediction of overland flow (OF) and vegetation density: (**a**) annual runoff on bedrock and on a colluvial slope with increasing colluvium in a downslope direction; (**b**) estimated plant cover and its evolution as erosion and deposition take place through runon from bedrock to a colluvial surface (after Kirkby 1990)

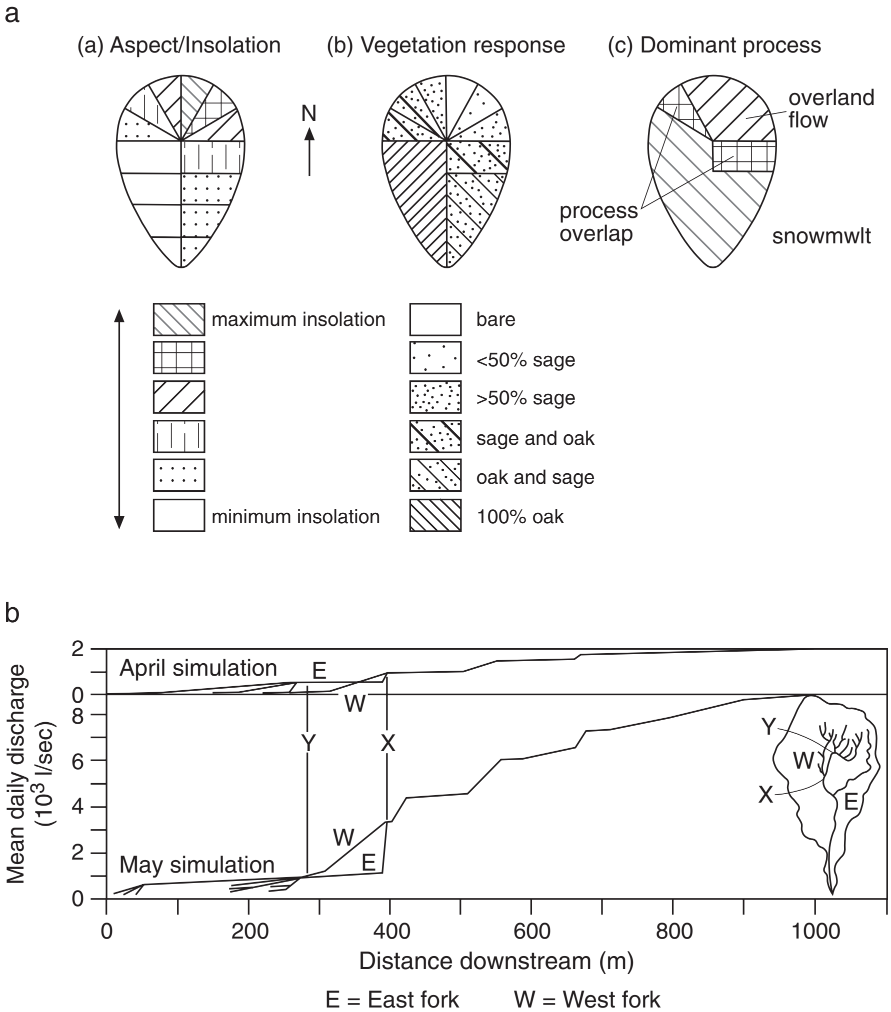


Fig. 11.16 (**a**) Hypothetical process domains in a mountain semi-arid snowmelt-dominated mountain catchment. (**b**) Simulated downstream changes in mean daily discharges in the pre-melt and melt seasons in Alkali Creek, Colorado (after Faulkner 1987)

At the regional scale the measured runoff redistribution of rainfall represents a complex range of factors. The various sources of losses to the surface affect the nature and timing of runoff in semi-arid environments on the hillslopes as well as in the channels. Pilgrim et al. (1979) estimated for the Fowler's Gap Arid Zone Research Station site in New South Wales initial losses prior to runoff of 5.35 mm with a continuing loss of typically 2 mm h^{-1}. However, as Wheater and Bell (1983) point out, catchment losses vary according to the scale of the runoff processes. In this regard, Shanen and Schick (1980) showed that in the Middle East initial loss estimates are 2.5 mm for small plots, 5.5 mm for 1–7 ha catchments, and 7–8 mm for a 3.45 km^2 catchment. This is a reflection of the general law that percentage runoff decreases with size of catchment, but in dry lands the losses are generally more acute. There is a mixture here too of direct losses from the catchment surface and channel transmission losses. The latter become particularly important with the development of an infiltrating channel fill (see below).

In recent years it has become generally more recognised following the work of Le Bissonais on runoff in relation to erosion in N.W. France that the surface characteristics of soils notably infiltration and crusting are not only important in runoff at several different scales but that they can be characterised and mapped using remotely sensed imagery. Coupled with GIS this is an important innovation in hillslope hydrology for large catchments. A simple example was Bull's (1991) distinction between loose sediment and rocky slopes or plateaux. In a direct application to arid zone hydrology on a very large scale Lange (1999) used six terrain types to represent hydrologically relevant surface characteristics in modelling runoff for the 1400 km^2 Nahel (Wadi) Zin in southern Israel. Each terrain type was studied in the field by obtaining infiltration parameters on small plots (Yair 1990) and detailed analyses of natural rainstorms on instrumented slopes. For the remaining terrain types, infiltration parameters were assessed by investigations of the stability of top-soil crusts and stony pavements (Yair 1992).

It was recognised long-ago (Thornes and Gilman 1983) that different Tertiary lithologies played a key role in runoff generation in Bronze-Age times in South-East Spain. It is the formal coupling of G.I. S. field studies and remote sensing in desert catchments that underlies the importance of recent work, especially recognising the intrinsic difficulties of working in and obtaining parameters for large desert areas. I raises the issues of what parameters are relevant and at what scales for defining hydrological similar surfaces.

Dalen et al. (2005) addressed precisely this question in their study of simulated and actual runoff in a dryland catchment in S.E. Spain. This proceeded through the development of a new runoff model base on the Green-Ampt algorithm coupled to the SCS Curve Number approach. This was used with a simulation model to examine the effects of rainfall intensity and geomorphological catchment characteristics on runoff at different scales, ranging from hill-slopes to moderate size catchments (up to 100 km^2). Vegetation was not taken into account. Nor was the aerial distribution of infiltration or crusting, but it is recognised as 'of great importance'.

It was found that spatial variations in rainfall intensity during the storm were most important in controlling run-off because of re-infiltration of run-off on long slopes. The 'effective intensity' decreases as (slope length)$^{-0.5}$ for major storms.

The second most important effect is produced by catchment geomorphic characteristics. Shape effects the distribution of travel times and channel cut-and-fill sequences affect transmission losses (Thornes 1977). Fresh incisions, such as badlands quickly convey water to the outlet.

Although crusting and infiltration properties were not specifically modelled, evidence from nearby areas demonstrate unequivocally that they are key controls on runoff (Li et al. 2005).

The HYSS approach seems destined to revolutionise hill-slope hydrology and thereby attempts to estimate runoff for large desert catchments.

Overall Water Balance

Accepting that these complexities arise in runoff regimes and that they have strong distributed patterns at the local scale, there is nevertheless an important role for a general model of the overall water balance, especially in the light of attempts to connect general circulation models to models of surface change. One of the most significant developments in this respect has been Eagleson's (1979) attempt to provide a hydrological model for the principal hydrological balance components in the overall annual hydrological cycle. This complex model rests on a statistical dynamic formulation of the vertical (point) water budget through equations expressing the infiltration, exfiltration, transpiration, percolation, and capillary rise from the water table both during and between storms. By asserting that the vegetation growth is in equilibrium and never in a stressed condition, Eagleson derives the expectation of the annual evapotranspiration and the optimal vegetation cover density for any location for which the required parameters are available. Typically the overall partition of the water budget and its division into climatic regimes are shown in Fig. 11.17a as a function of rainfall m_{PA}. The E outside the brackets indicates that the values are the statistically expected values. The solid lines indicate the major controls. Actual evapotranspiration $E^*_{T_\mathrm{A}}$ increases with rainfall to a maximum when it equals potential evapotranspiration $E^*_{P_\mathrm{A}}$. Infiltration I_A increases with rainfall until it is limited by saturated percolation to groundwater, which is a function of the mean duration of the rainy season m_τ and the saturated hydraulic conductivity $K(1)$. The uppermost, straight, solid line is the line through which mean annual rainfall m_{P_A} is equal to mean annual potential evapotranspiration $E^*_{P_\mathrm{A}}$, so to the right of this line there is always excess water.

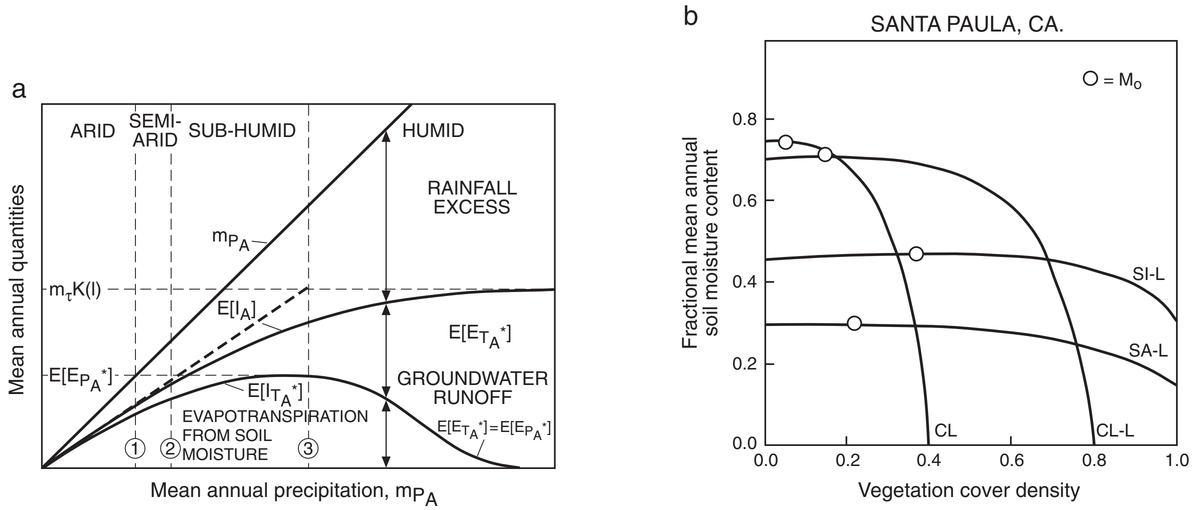


Fig. 11.17 (**a**) The hydrological components and classification of hydrological regimes according to Eagleson's model. (**b**) The relationship between plant cover density, soil type, and soil moisture in the same model

The vertical dashed lines associated with numbers just above the horizontal axis define climatic zones. The lower bound of the semi-arid zone (1) occurs when the mean annual rainfall is greater than the mean actual annual evapotranspiration, recalling that the latter includes the effects of seasonality, vegetation cover, soil type, and so on. The upper bound of this zone (2) separates the semi-arid from the subhumid regimes. In the former, infiltration and evapotranspiration are profile- (soil-) controlled, whereas in the latter evapotranspiration is climate-controlled. Above this (above rainfall at 3) infiltration is greater than actual evapotranspiration and the soil will be fully humid. Figure 11.17(b) shows the response of vegetation cover to soil moisture for different soil types (silt-loam, sandy loam, clay-loam, and clay). The point marked M_o is the optimum vegetation cover for the prevailing soil characteristics and soil moisture in an unstressed condition in Santa Paula, California (Eagleson 1979).

Groundwater

Although deep percolation from rainfalls is generally small and, therefore, groundwater is generally relatively unimportant from a geomorphic point of view, this is not true of suballuvial aquifers. Here the aquifer is recharged from transmission losses from channel flow and, therefore, it has important implications for runoff and transporting capacity. Moreover, the suballuvial rockhead configuration may lead to groundwater resurgence and availability of water to phreatophytes. Figure 11.18a shows the configuration of the alluvial fill of the Mojave River, California. It can be seen from this diagram that the suballuvial aquifer thins and shallows in the vicinity of Victorville, to the south of Bell Mountain, and again near Barstow. These subsurface ‘narrows’ force the water near to the surface, causing a significant growth of phreatophytes, which have a significant effect on channel behaviour and morphology. Figure 11.18b shows the pattern of precipitation at Lake Arrowhead and the resulting flow at Victorville and Barstow for the spring of 1969. By the time the flow had reached Barstow, the transmission losses to the bed had eliminated the lower flows. Finally, Fig. 11.18c shows the evolution of ground water in the suballuvial aquifer. This graph indicates the remarkable recovery of levels by recharge in the 1969 and 1978 floods. Buono and Lang (1980) reported that by the time the flow had reached Victorville 43% of it had been lost, and that 50% had been lost by the time it reached Barstow.

In another study of a much smaller channel system, Butcher and Thornes (1978) attempted to model the impact of transmission losses on channel flow and the survival of flow to the main channel. They showed that

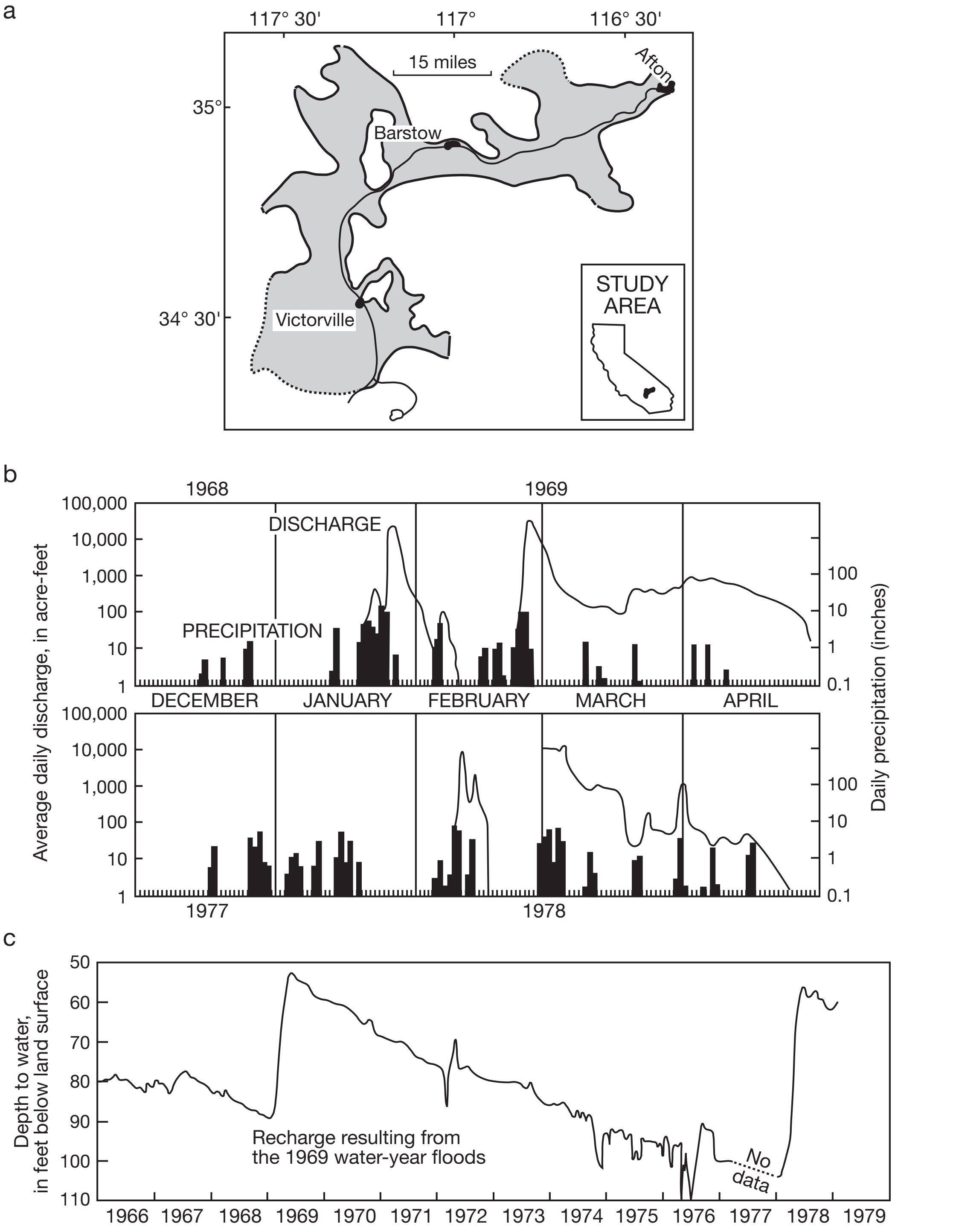


Fig. 11.18 Transmission losses in the Mojave River, California (after Buono and Lang 1980). (**a**) General location map showing alluvial fill. (**b**) Precipitation and flow for the floods of 1969 and 1978. (**c**) Changes in the water level below the surface at Baker

coupled with the transmission losses, the propensity for flows to survive in smaller channels and reach the main channel flow system also depended on the contributions of yet smaller channels to the system in question. The pattern of contributions of smaller subcatchments, called the area increment function, determined the likelihood that a flow would survive as far as the next tributary input. Butcher and Thornes modelled flows using a kinematic cascade routing model (KINGEN) developed by Woolhiser (1976) coupled with a transmission loss model for reaches between major tributaries based on a simple Kostiakov-type equation and the opportunity time for flow permitted by the passage of the flood wave. Smaller infiltration rates were found to produce disproportionately greater survival lengths. This application to a specific channel was generalized by Thornes (1977), who combined three principles – a differential equation for channel losses developed from Burkham's (1970) model, the hydraulic geometry of ephemeral channels (Leopold and Miller, 1956), and an area increment function based on Shreve (1974) – to solve the equations for the probability of survival downstream given various parameters derived from field studies. The general spatial scheme is shown in Fig. 11.19, together with typical theoretically derived survival curves. The implications of this model for geomorphic phenomena are illustrated in the next chapter. An excellent summary of the modelling of transmission losses was given by Lane (1980), and Lane (1982) developed a semi-empirical approach to transmission losses in the context of distributed modelling for small semi-arid watersheds.

Channel Flow

McMahon (1979) summarized the characteristics of runoff in desert environments on the basis of an examination of 70 annual flow records and peak discharge series from six desert zones. He found that runoff in desert channels was more varied than in humid channels, a fact confirmed by McMahon et al. (1987). Examining global runoff variations, these authors found that as runoff decreases, variability (defined as the standard deviation/mean annual runoff) increases. Consequently, extrapolation from humid zones is not acceptable. In testing the carryover effect, it was found that in desert channels the first-order serial correlation between annual flows (for 50 records) is 0.01. McMahon summarized the situation as follows:

(a) flood events are irregular and of short duration;
(b) data from gauging records are extremely poor;
(c) high stream velocities ($4\,\mathrm{m\,s^{-1}}$) are common;
(d) high debris loads occur during the passage of flood waves;
(e) gauging station controls are sandy and unstable;
(f) drainage boundaries are indefinite;
(g) overbank flows occur frequently;
(h) underflow is often a large part of flood events.

These conclusions are borne out by a few examples. Costa (1987) itemized the 12 largest floods ever measured in the United States. All occurred in semi-arid to arid areas, with mean annual rainfalls ranging from 114 to 676 mm; 10 had mean annual rainfalls less than 400 mm. This finding conflicts with the results of Wolman and Gerson (1978) who concluded that differences in maximum possible runoff from a single severe storm in different physiographic regions and climatic regimes appear to be insignificant. However, the United States experience does not suggest that these floods were due simply to high rainfall intensity. Rather it indicates that for a given intensity, rains in semi-arid and arid environments produce more runoff per unit area than they do in temperate environments. This finding reflects the whole panoply of surface infiltration and runoff controls dealt with earlier in the chapter. Mean velocity in these floods ranged from 3.47 to $9.97\,\mathrm{m\,s^{-1}}$, and shear stresses and unit stream powers were several hundred times greater than those produced in larger rivers.

The analysis of more individual extreme events is even more perplexing. Wheater and Bell (1983) found that problems with data are severe and that flood hydrology in arid zones was (and still is) largely unquantified. They exemplify the difficulties involved with reference to Wadi Adai in Oman where their reconstruction of a specific major flood depended heavily on a mixture of modelling and parameter estimation from sparse data. The wadi has a catchment area of $370\,\mathrm{km^2}$, channel slopes of 0.13–0.50, and a channel width of 50–150 m. On the basis of intermittent flood records stretching back to 1873, the flood of 3 May 1981 had a recurrence interval of 100–300 years. The main flood

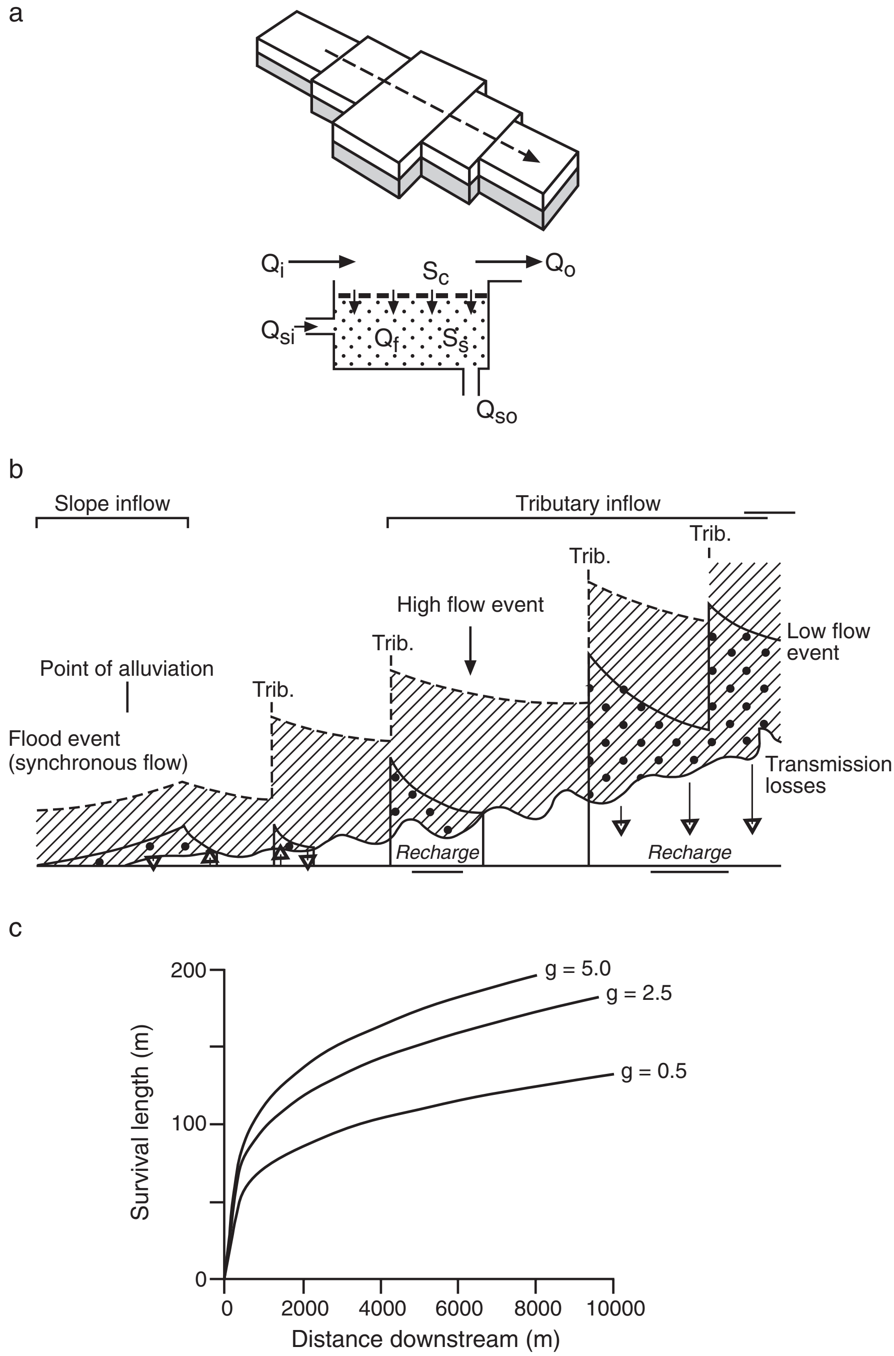


Fig. 11.19 (**a**) The effects of transmission losses on flow (after Thornes 1977). Q is surface discharge, Q_s is subsurface discharge, Q_f is transmission losses, S_c is channel sediment discharge and S_s is subsurface sediment discharge. The subscripts i and o denote inflow and outflow, respectively. (**b**) Hypothetical pattern of flows for two flood events. (**c**) Distribution of survival lengths with variation in the parameter g, which is the rate of change of discharge with area

peak arrived at 0300 h, the recession started at 0530 h, and there was hardly any flow by 1700 h. Flow velocities were estimated to be 3.6–4.5 m s^{-1}.

Descriptions abound of the characteristic features of flow in ephemeral channels, though the number of well-documented sites at which they have been properly investigated is still quite small. Among these the best are the Walnut Gulch Experimental Watershed, Arizona (Renard 1970) and the Nahal Yael Watershed, Israel (Schick 1986). The events in ephemeral channels, separated by zero flows, are often asynchronous across the channel network and are sometimes accompanied by well-developed bores and periodic surges in flow. Moreover, the recession limbs are usually very steep. All these features have been modelled, sometimes in combination, and Lane (1982) has considered their application to Walnut Gulch. We now consider the modelling of some of the components. The geomorphic impact of these phenomena is considered in the following chapter.

The development of a steep bore at the front of ephemeral channel flow is caused by a combination of intense storms, rapid translation of the water under conditions of high drainage density, and the interaction of wave movement and transmission losses. The second of these is addressed in Chapter 12. The development of a bore is analogous to the release of water in an irrigation ditch or in a dam burst. At the front edge of the wave the water is initially shallow and the bed friction high. This is compounded by the exhaust of air from the bed as infiltration forces it out, sometimes trapping it between the advancing subsurface wetting front and the groundwater.

Smith (1972) modelled these effects using the kinematic wave approach, which assumes that consideration of the momentum of the flow can be neglected. He predicted the advance rate, surface profiles, and modifications with time to kinematic wave flow over an initially dry infiltrating plane. Although the kinematic wave approach has been the subject of much recent criticism, it appears to be appropriate for small catchments, where it is possible to resolve the physical detail without compromising the deterministic nature of the model (Ponce 1991). Smith's approach was to set up the continuity equation to take account of the transmission losses – that is

$$\mathrm{d}h/\mathrm{d}t + \mathrm{d}(uh)/\mathrm{d}x = q(x, t) \qquad (11.5)$$

where h is flow depth, u the local velocity, x the distance along the channel, t is time, and $q(x, t)$ the local inflow (e.g. from tributaries or valley-side walls) or outflow (i.e. transmission losses). Figure 11.20a shows the basic setup of his model. A sharp fronted flood wave is advancing from left to right across an inclined channel bed. Beneath the bed the graphs show that infiltration rate is greatest near the front of the advancing wave and that the cumulative infiltration increases away from the advancing front. Laboratory experiments in a glass-sided flume by the author (Thornes 1979) clearly reveal a sharp wetting front that is quite even in sand-sized sediments, but has a fractal appearance in more variable sediments. The kinematic approximation is given by the stage-discharge equation

$$Q = \alpha \cdot h^{m+1} \qquad (11.6)$$

in which α and m are constants. The wave front itself is assumed to move as a kinematic shock. For infiltration, Smith set

$$-q(x) = K(T)^{-a} + f_0 \qquad (11.7)$$

in which $q(x)$ is the infiltration rate at x, T is the time since surface ponding, K and a are coefficients related to the characteristics of the channel bed materials, and f_0 is the long-term transmission loss rate. This is essentially the Kostiakov equation. Typical advance rates of the front are shown for the model (which replicated field data of Criddle (1956) very well) in Figure 11.20b. The curve suggests that the velocity of the wave front (the gradient of the line) increases with time, the average rate over this period being about 1 m min^{-1} in this particular case. Typical figures in semi-arid channels seem to be of the order of 1–3 km h^{-1} (as illustrated for Walnut Gulch in Fig. 11.20c), and the model replicates well the steepening of the wave front downstream. Behind the peak flow of the hydrograph, which comes almost immediately after the wave front, the deeper water provides a relatively lower roughness and a higher velocity, and so the wave front grows.

Infiltration also takes its toll on the recession limbs of ephemeral channel flood waves, which are also usually steeper than in temperate channels. The recession limb can be imagined as a wedge-shaped reservoir moving down channel, with the tail of the flow catching up with the front as transmission losses draw in

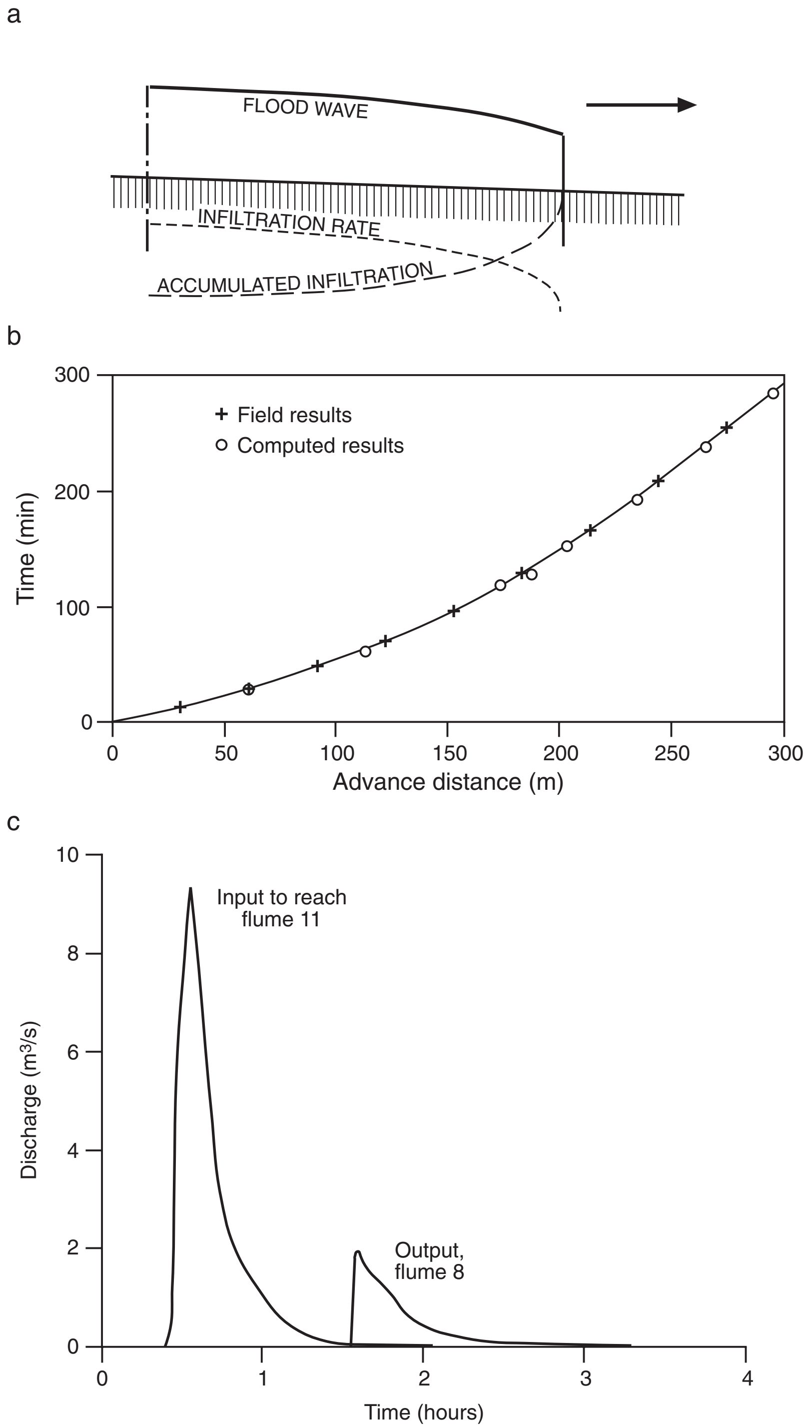


Fig. 11.20 Smith's (1972) model of wave front propagation in an infiltrating channel: (**a**) general set up of model, (**b**) downstream progress of flood wave compared with field data, and (**c**) actual propagation of a flood wave in Walnut Gulch between two stations

the shallower water of the recession limb, as has been modelled by Peebles (1975).

Not all rising limbs are associated with a single wave of water, and Foley (1978) described the onset of flow to be a series of translatory waves of small amplitude building up to flood depth. Periodic surges are also a characteristic of steady ephemeral channel flow, especially in larger channels. These appear as sudden increases in stage and may move as waves faster than the actual flow of the water itself, usually, but not always in a downstream direction. They can be responsible for extensive damage and also for some special sediment transport and depositional effects exclusive to this kind of flow. Leopold and Miller (1956) provided an excellent description of such a flow in which they observed a series of bores each 0.15–0.30 m high and between 0.5 and 1 min apart. One possible explanation is that the waves represent asynchronous contributions from different parts of the channel network. However, Leopold and Miller (1956) ascribed them to a type of momentum wave associated with the hydraulics of the channel itself. Such surges have also been simulated in the laboratory by Brock (1969), and R.J. Heggen (unpublished data 1986) described changes in width and velocity as well as height in these phenomena. Generally, under Green's Law, the elevation of a sinusoidal wave decreases as channel width increases, so the surges tend generally to attenuate downstream.

Conclusion

From this chapter it is evident that while the essential controls of arid and semi-arid hydrology are much the same as those of temperate environments, it is the time–space intensity of the hydrological processes which leads to significant differences in the corresponding geomorphic processes. Above all it is the time and space distribution of precipitation which leads to a much less ordered patterning of processes and the resulting forms. Even in arid environments, and certainly in semi-arid environments, the spatio-temporal diversity of rainfall leads to a distinctive pattern of plant water use, plant growth, and the resulting biological activity, both on hillslopes and, to a lesser extent, in channels. These two interrelated elements, water and biological growth, are the key to understanding past, present, and future deserts.

References

Abrahams, A.D. and A.J. Parsons 1991. Relation between infiltration and stone cover on a semiarid hillslope, southern Arizona. *Journal of Hydrology* **122**, 49–59.

Abrahams, A.D., A.J. Parsons and S.H. Luk 1986. Resistance to overland flow on desert slopes. *Journal of Hydrology* **88**, 343–63.

Ahnert, F. 1987. An approach to the identification of morphoclimates. In *International Geomorphology 1986*, Volume II, V. Gardiner (ed.), 159–88. Chichester: Wiley.

Alexander, R.W. and A. Calvo 1990. The influence of lichens on slope processes in some Spanish badlands. In *Vegetation and Erosion*, J.B. Thornes (ed.), 385–98. Chichester: Wiley.

Baird, A., J.B. Thornes and G. Watts 1992. Extending overland flow models to problems of slope evolution and the representation of complex slope surface topographies. In *Overland Flow: Hydraulics and Erosion Mechanics*, A.J. Parsons and A.D. Abrahams (eds), 199–211. London: University College Press.

Bell, F.C. 1979. Precipitation. In *Arid Land Ecosystems*, Volume I, D.W. Goodall and R.A. Perry (eds), 373–93. Cambridge: Cambridge University Press.

Berndtsson, R. 1987. Spatial and temporal variability of rainfall and potential evapotranspiration in Tunisia. *International Association of Scientific Hydrology Publication* 168, 91–100.

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

Bouwer, H. 1975. Predicting reduction in water losses from open channels by phreatophyte control. *Water Resources Research* **11**, 96–101.

Brock, R.R. 1969. Development of roll-wave trains in open channels. *Journal Hydraulics Division, Proceedings of the American Society of Civil Engineers* **95**, 1401–27.

Bull, W.B. 1991. *Geomorphic Response to Climate Change*, New York: Oxford University Press.

Buono, A. and D.J. Lang 1980. Aquifer recharge from the 1969 and 1978 floods in the Mojave River basin, California. In *Water Resources Investigations Open File Report 80–207*. Washington, DC: U.S. Government Printing Office, U.S. Geological Survey.

Burkham, D.E. 1970. A method for relating infiltration rates to stream flow rates in perched streams. *U.S. Geological Survey Professional Paper* 700-D, 226–71.

Butcher, G.C. and J.B. Thornes 1978. Spatial variability in runoff processes in an ephemeral channel. *Zeitschrift für Geomorphologie Supplement Band* **29**, 83–92.

Carson, M.A. and M.J. Kirkby 1971. *Hillslope form and process*. Cambridge: Cambridge University Press.

Conte, M., A. Giuffrida and S. Tedesco 1989. The Mediterranean oscillation. In *Conference on Climate and Water*, L. Huttunen (ed.), 121–38. Helsinki: Government of Finland Printing House.

Costa, 1987. Hydraulics and basin morphology of the largest flash floods in the conterminous United States. *Journal of Hydrology*, **93**, 313–38.

Criddle, C.L. 1956. Methods for evaluating irrigation systems. In *Agricultural Handbook*, 82, Washington, DC: U.S. Department of Agriculture.

Dalen, E.N., M.J. Kirkby, P.J. Chapman and L.J. Bracken 2005. Runoff generation in S.E. Spain. Poster presentation to the European Geophysical Union Meeting, Vienna.

De Ploey, J. 1982. A stemflow equation for grasses and similar vegetation. *Catena* **9**, 139–52.

De Ploey, J., M.J. Kirkby and F. Ahnert 1991. Hillslope erosion by rainstorms – a magnitude and frequency analysis. *Earth Surface Processes and Landforms* **16**, 399–410.

Dhar, O.N. and P.R. Rakhecha 1979. Incidence of heavy rainfall in the Indian desert region. *International Association for Scientific Hydrology Publication* **128**, 13–8.

Duckstein, L., M. Fogel and I. Bogardi 1979. Event based models of precipitation. *International Association of Scientific Hydrology Publication* **128**, 51–64.

Dunne, T., W. Zhang and B.F. Aubry 1991. Effects of rainfall, vegetation and microtopography on infiltration and runoff. *Water Resources Research* **27**, 2271–87.

Eagleson, P.S. 1979. Climate, soil and vegetation. *Water Resources Research* **14**, 705–77.

Emmett, W.W. 1970. The hydraulics of overland flow on hillslopes. *U.S. Geological Survey Professional* Paper 662–A.

Evans, D.D., T.W. Sammnis and B. Asher 1976. Plant growth and water transfer interactive processes under desert conditions. *US/IBP Desert Biome Research Memoir* 75–44. Logan, UT: Utah State University.

Evans, D.D., T.W. Sammnis and D.R. Cable 1981. Actual evapotranspiration under desert conditions. In *Water in Desert Ecosystems*, D.D. Evans and J.L. Thames (eds), 195–219. Stroudsburg, PA: Dowden, Hutchinson and Ross.

Faulkner, H. 1987. Gully evolution in response to flash flood erosion, western Colorado. In *International Geomorphology 1986*, Volume I, V. Gardiner (ed.), 947–73. Chichester: Wiley.

Faulkner, H. 1990. Vegetation cover density variations and infiltration patterns on piped alkali sodic soils. Implications for modelling overland flow in semi-arid areas. In *Vegetation and Erosion*, J.B. Thornes (ed.), 317–46. Chichester: Wiley.

Feddes, R.A. 1971. Water, heat and crop growth. Thesis, Agricultural University of Wageningen.

Floret, C., R. Pontanier and S. Rambal 1978. Measurement and modelling of primary production and water use in a south Tunisian steppe. *Journal of Arid Environments*, **82**, 77–90.

Foley, M.G. 1978. Scour and fill in steep sand-bed ephemeral streams. *Bulletin of the Geological Society of America* **89**, 559–70.

Fournier, F. 1960 *Climat et erosion*. Paris: Presses Universitaire.

Francis, C.F. and J.B. Thornes 1990. Runoff hydrographs from three Mediterranean vegetation cover types. In *Vegetation and erosion*, J.B. Thornes (ed.) 363–84. Chichester: Wiley.

Fuchs, M. 1979. Atmospheric transport processes above arid land vegetation. In *Arid Land Ecosystems*, Volume 1, D.W. Goodall and R.A. Perry (eds), 393–433. Cambridge: Cambridge University Press.

Gifford, G.F., M. Merzougi and M. Achouri 1986. Spatial variability characteristics of infiltration rates on a seeded rangeland site in Utah, USA. In *Rangeland: A Resource Under Siege*, P.J. Joss, P.W. Lynch and O.B. Williams (eds), 46–7. Canberra: Australian Academy of Sciences.

Gonzalez-Hidalgo, C. 1991. Aspect, vegetation and erosion on slopes in the Violada area, Zaragosa. Unpublished Ph.D. thesis, University of Zaragosa.

Gregory, P.J. 1991. Soil and plant factors affecting the estimation of water extraction by crops. *International Association for Scientific Hydrology Publication* **199**, 261–73.

Herschfield, D.M. 1962. Extreme rainfall relationships. *Journal of Hydraulics Division, Proceedings of the American Society of Civil Engineers* **88**, 73–92.

Hillel, D. and N. Tadmoor 1962. Water regime and vegetation in the central Negev Highlands of Israel. *Ecology* **43**, 33–41.

Jacobs, B.L., I. Rodriguez-Iturbe and P. Eagleson 1988. Evaluation of homogenous point process description of Arizona thunderstorm rainfall. *Water Resources Research* **24**, 1174–86.

Johns, G.G., D.J. Tongway and G. Pickup 1984. Land and water processes. In *Management of Australia's Rangelands*, G.N. Harrington, A.D. Wilson and M.D. Young (eds), 25–40. Canberra: Commonwealth Scientific and Industrial Research Organization.

Jury, W.A., J. Letey and L.H. Stolzy 1981. Flow of water and energy under desert conditions. In *Water in Desert Ecosystems*, D.D. Evans and J.L. Thames (eds), 92–113. Stroudsburg, PA: Dowden, Hutchinson and Ross.

Kirkby, M.J. and R.H. Neale 1987. A soil erosion model incorporating seasonal factors. In *International Geomorphology 1986*, Volume II, V. Gardiner (ed.), 189–210. Chichester: Wiley.

Kirkby, M.J. 1990. A simulation model for desert runoff and erosion. *International Association for Scientific Hydrology Publication* 129, 87–104.

Lange, J., C. Leibundgut, N. Greenbaum and A.P. Schick 1999. A non-calibrated rainfall-runoff model for large, arid catchments. *Water Resources Research* **35(7)**, 2161–2172

Lane, L.J. 1980. Transmission losses. *National Engineering Handbook*, Section 4, Chapter 19. Washington, DC: U.S. Printing Office, U.S. Department of Agriculture, Soil Conservation Service.

Lane, L.J. 1982. Distributed model for small semi-arid watersheds. *Journal Hydraulics Division. Proceedings of the American Society of Civil Engineers* **108**, 1114–31.

Lane, L.J., E.M. Romney and T.E. Hakenson 1984. Water balance calculations and net production of perennial vegetation in the northern Mojave desert. *Journal Range Management* **37**, 12–18.

Langbein, W.B. and S.A. Schumm 1958. Yield of sediment in relation to mean annual precipitation. *Transactions of the American Geophysical Union* **39**, 1076–84.

Lascano, R.J., C.H.M. van Bavel, J.L. Hatfield and D.R. Upchurch 1987. Energy and water balance of a sparse crop: simulated and measured soil evaporation. *Soil Science Society of America Journal* **51**, 1113–21.

Le Houerou, N. 1979. North Africa. In *Arid Land Ecosystems*, Volume 1, D.W. Goodall and R.A. Perry (eds), 83–107. Cambridge: Cambridge University Press.

Leith, H. and R.H. Whittaker 1975. *The Primary Productivity of the Biosphere*, Ecological Studies 14. New York: Springer.

Leopold, L.B. and J.P. Miller, 1956. Ephemeral streams – hydraulic factors and their relationship to the drainage net. *U.S. Geological Survey Professional Paper* 282-A.

Levitt, J. 1972. *Responses of Plants to Environmental Stress*. New York: Academic Press.

Li, X.-Y., A. González and A. Solé Benet 2005. Laboratory methods for the estimation of infiltration rate of soil crusts in the Tabernas Desert badlands. *Catena* **60**, 255–266.

Lyford, F.P. and H.K. Qashu 1969. Infiltration rates as affected by desert vegetation. *Water Resources Research* **5**, 1373–6.

McMahon, T.A. 1979. Hydrological characteristics of arid zones. *International Association of Scientific Hydrology Publication* **128**, 105–23.

McMahon, T.A., B.L. Finlayson and R. Srikanthan 1987. Runoff variability: a global perspective. *International Association of Scientific Hydrology Publication* **168**, 3–12.

Mehuys, G.R., L.H. Stolzy, J. Letey and L.V. Weeks 1975. Effect of stones on the hydraulic conductivity of relatively dry desert soils. *Soil Science Society of America Proceedings* **39**, 37–42.

Monteith, J. 1981. Evaporation and surface temperature. *Quarterly Journal of the Royal Meteorological Society* **107**, 1–27.

Morin, J. and Y. Benyami 1977. Rainfall infiltration into bare soils. *Water Resources Research* **13**, 813–7.

Nobel, P.S. 1981. Spacing and transpiration of various sized clumps of a desert grass *Hilaria rigida. Journal of Ecology* **69**, 735–42.

Peebles, R.W. 1975. Flow recession in the ephemeral stream. Unpublished Ph.D. thesis, University of Arizona, Tucson.

Peltier, L.C. 1950. The geographical cycle in periglacial regions. *Annals of the Association of American Geographers* **50**, 214–36.

Phillips, D.L. and J.A. McMahon, 1981. Competition and spacing patterns in desert shrubs. *Journal of Ecology* **69**, 97–115.

Pickup, G. 1988. Modelling arid zone soil erosion at the regional scale. In *Essays in Australian Fluvial Geomorphology*, R.F. Warner (ed.), 1–18. Canberra: Academic Press.

Pilgrim, D.H., I. Cordery and D.G. Doran 1979. Assessment of runoff characteristics in arid western New South Wales. *International Association for Scientific Hydrology Publication* **128**, 141–50.

Poesen, J., F. Ingelmo-Sanchez and H. Mucher 1990. The hydrological response of soil surfaces to rainfall as affected by cover and position of rock fragments in the top layer. *Earth Surface Processes and Landforms*, **15**, 653–72.

Ponce, V.M. 1991. The kinematic wave controversy. *Journal of Hydraulic Engineering* **117**, 511–25.

Renard, K.J. 1969. Evaporation from an ephemeral stream bed: discussion. *Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers* **95**, 2200–4.

Renard, K.J. 1970. The hydrology of semi-arid rangeland watersheds. *U.S. Department of Agriculture, Agricultural Research Service*, 41–162. Washington, DC: U.S. Government Printing Office.

Ritchie, J.T. 1972. Model for predicting evaporation from a row crop with incomplete cover. *Water Resources Research* **8**, 1204–13.

Roo, A.P.J. and H. Th. Riezebos 1992. Infiltration experiments on loess soils and their implications for modelling surface runoff and soil erosion. *Catena* **19**, 221–41.

Schick, A.P. 1986. Hydrologic aspects of floods in extreme arid climates. In *Flood geomorphology*, V.R. Baker, R.C. Kochel, R.C. and P.C. Patton (eds), 189–203. Wiley: New York.

Scholl, D.G. 1974. Soil moisture flux and evapotranspiration determined from soil hydraulic properties in a chaparral stand. *Soil Science Society America Journal* **40**, 14–18.

Scoging, H. 1989. Run-off generation and sediment mobilisation by water. In *Arid Zone Geomorphology*, D.S.G. Thomas (ed.) 87–116. London: Belhaven.

Scoging, H. and J.B. Thornes 1979. Infiltration characteristics in a semi-arid environment. *International Association Scientific Hydrology Publication* **128**, 159–68.

Shanen, L. and A.P. Schick 1980. A hydrological model for the Negev Desert Highlands – effects of infiltration, runoff and ancient agriculture. *Hydrological Sciences Bulletin* **25**, 269–82.

Shmida, A. 1985. Biogeography of desert flora. In *Ecosystems of the World*, Volume 12A, *Hot Desert and Arid Shrublands*, Evenari, M., I. Noy-Meir and D.W. Goodall (eds) 23–77. Amsterdam: Elsevier.

Shmida, A., M. Evenari and I. Noy-Meir 1986. Hot desert ecosystems an integrated view. In *Ecosystems of the World*, Volume 12B, *Hot Deserts and Arid Shrublands*, M. Evenari, I. Noy-Meir and D.W. Goodall (eds), 379–88. Amsterdam: Elsevier.

Shreve, R.L. 1974 Variations of main stream length with basin area in river networks. *Water Resources Research* **10**, 1167–77.

Shuttleworth, W.J. and J.S. Wallace 1985. Evaporation from sparse crops – an energy combination theory. *Quarterly Journal Royal Meteorological Society* **111**, 839–55.

Slatyer, R.O. 1965. Measurements of precipitation interception by an arid zone plant community (*Acacia aneura*). *UNESCO Arid Zone Research* **25**, 181–92.

Smith, R.E. 1972. Border irrigation advance and ephemeral flood waves. *Proceedings American Society of Civil Engineers* **98(IR2)**, 289–307.

Smith R.E. and H.A. Schreiber 1973. Point processes of seasonal thunderstorm rainfall 1. Distribution of rainfall events. *Water Resources Research* **9**, 871–84.

Swanson, S.R. and J.C. Buckhouse 1986. Infiltration on Oregon lands occupied by three subspecies of big sagebrush *Artemisia. U.S. Department of Agriculture, Agricultural Research Service* INT-200, 286–91. Washington, DC: U.S. Government Printing Office.

Thornes, J.B. 1976. *Semi-Arid Erosional Systems*, Geography Research Papers No. 7. London: London School of Economics.

Thornes, J.B. 1977. Channel changes in ephemeral streams, observations, problems and models. In *River Channel Changes*, K.J. Gregory (ed.), 317–55. Chichester: Wiley.

Thornes, J.B. 1979. Fluvial Processes. In *Process in Geomorphology*, C.E. Embleton and J.B. Thornes (eds), 213–72. London: Arnold.

Thornes, J.B. 1990. The interaction of erosional and vegetational dynamics in land degradation: spatial outcomes. In *Vegetation and Erosion*, Thornes, J.B. (ed.) 41–53. Chichester: Wiley.

Thornes, J.B. 1991. Environmental change and hydrology. In *El Agua en Andalucia III*, V. Giraldez (ed.), 555–70. Cordoba: University of Cordoba.

Thornes, J.B. and J.C. Brandt. 1994. Erosion-vegetation competition in a stochastic environment undergoing climatic change. In *Environmental Change in Drylands*, A.C. Millington and K. Pye (eds). 305–20. Chichester: Wiley.

Thornes, J.B. and Gilman A. (1983) Potential and actual erosion around archaeological sites in south-east Spain' In J. de Ploey (ed.), *Rainfall Simulation, Runoff and Soil Erosion*, Catena Supplement No.4, Catena, Cremelingen.

Thornes, J.B., C.F. Francis, F. Lopez-Bermudez and A. Romero-Diaz 1990. Reticular overland flow with coarse particles and vegetation roughness under Mediterranean conditions. In *Strategies to Control Desertification in Mediterranean Europe*, Rubio, J.L. and J. Rickson (eds), 228–43. Brussels: European Community.

Tromble, J.M., K.G. Renard and A.P. Thatcher 1974. Infiltration for three rangeland soil-vegetation complexes. *Journal Range Management* **41**, 197–206.

Van Hylckama, T.E.A. 1975. Water use by salt cedar in the lower Gila River valley (Arizona) *U.S. Geological Survey Professional Paper* 491-E.

Wallace, J.S. 1991. Measurement and modelling of evaporation from a semi-arid system. *International Association Scientific Hydrology Publication* **199**, 131–48.

Walter, H. 1971. *Ecology of Tropical and Subtropical Vegetation*. Edinburgh: Oliver and Boyd.

Wheater, H.S. and N.C. Bell 1983. Northern Oman flood study. *Proceedings of the Institute of Civil Engineers* **75**, 453–73.

Whittaker, R.H. and W.A. Niering 1964. The vegetation of the Santa Catalina Mountains, Arizona. 1. Ecological classification and distribution of species. *Journal of the Arizonian Academy of Science* **3**, 9–34.

Wilcox, B.P., M.K. Wood and J.M. Tromble 1988. Factors influencing infiltrability of semiarid mountain slopes. *Journal of Range Management* **41**, 197–206.

Wolman, M.G. and R. Gerson 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. *Earth Surface Processes* **3**, 189–208.

Wolman, M.G. and J.P. Miller 1960. Magnitude and frequency of forces in geomorphic processes. *Journal Geology* **68**, 54–74.

Woodward, F.I. 1987. *Climate and Plant Distribution*. Cambridge: Cambridge University Press.

Woolhiser, D.A. 1976. Overland flow. In *Unsteady Open Channel Flow*. K. Mahmood and V. Yevjevich (eds), 485–508. Fort Collins, CO: Water Resources Publications.

Yair, A. 1990. Runoff generation in a sandy area – the Nizzana Sands, Western Negev, Israel. *Earth Surface Processes and Landforms* **15**, 597–609.

Yair, A. 1992. The control of headwater area on channel runoff in a small arid watershed in Parsons, A.J. and Abrahams, A. (Ed) *Overland Flow*, London, University College Press, 53–68.

Yair, A. and H. Lavee 1974. Areal contribution to runoff on scree slopes in an extreme arid environment – a simulated rainfall experiment. *Zeitschrift für Geomorphologie Supplement Band* **21**, 106–21.

Yair, A. and H. Lavee 1985. Runoff generation in arid and semi-arid environments. In *Hydrological Forecasting*, M.G. Anderson and T. Burt (eds), 183–220. Chichester: Wiley.

Zawadeski, I.I. 1973. Statistical properties of precipitation patterns. *Journal of Applied Meteorology* **12**, 459–71.