CHAPTER 3

Factors influencing erosion

The factors controlling soil erosion are the erosivity of the eroding agent, the erodibility of the soil, the slope of the land and the nature of the plant cover.

3.1 Erosivity

3.1.1 Rainfall

Soil loss is closely related to rainfall partly through the detaching power of raindrops striking the soil surface and partly through the contribution of rain to runoff. This applies particularly to erosion by overland flow and rills, for which intensity is generally considered to be the most important rainfall characteristic. The effect of rainfall intensity is illustrated by the data for 183 rain events at Zanesville, Ohio, between 1934 and 1942, which show that average soil loss per rain event increases with the intensity of the storm (Table 3.1; Fournier 1972).

The role of intensity is not always so obvious, however, as indicated by studies of erosion in mid-Bedfordshire, England, taking data for the ten most erosive storms between May 1973 and October 1975. While intense storms, such as the one of 34.9 mm on 6 July 1973, in which 17.7 mm fell at intensities greater than 10 mm h^{-1}, produce erosion, so do storms of long duration and low intensity, like the one of 19 June 1973 when 39.6 mm of rain fell in over 23 hours (Morgan et al. 1986). It appears that erosion is related to two types of rain event, the short-lived intense storm where the infiltration capacity of the soil is exceeded, and the prolonged storm of low intensity that saturates the soil.

The response of the soil to rainfall may also be determined by previous meteorological conditions. This can again be demonstrated by data for Zanesville, Ohio (Fournier 1972). A storm of 19.3 mm on 9 June 1940 fell on dry ground and, despite the quantity, only 25 per cent went into runoff, most of the water soaking into the soil. On the following day, in a storm of 13.7 mm, 66 per cent of the rain ran off and soil loss almost trebled. The control in this case is the closeness of the soil to saturation, which is dependent on how much rain has fallen in the previous few days. The pattern of low soil loss in the first and high loss in the second of a series of storms is reversed, however, when, between erosive storms, weathering and light rainfall loosen the soil surface. Most of the loose material is removed during the first runoff event, leaving little for erosion in subsequent storms. This sequence is illustrated by studies in the Alkali Creek

Table 3.1 Relationship between rainfall intensity and soil loss

Maximum 5-min intensity (mm h^{-1})	Number of falls of rain	Average soil loss per rainfall (t ha^{-1})
0–25.4	40	3.7
25.5–50.8	61	6.0
50.9–76.2	40	11.8
76.3–101.6	19	11.4
101.7–127.0	13	34.2
127.1–152.4	4	36.3
152.5–177.8	5	38.7
177.9–254.0	1	47.9

Data for Zanesville, Ohio, 1934–42 (after Fournier 1972).

watershed, Colorado (Heede 1975b), where, following a year without runoff, a sediment discharge peak of 143 kg s^{-1} was observed on 15 April 1964 in a runoff from snowmelt of 2.21 m^3 s^{-1}. Next day, peak runoff increased to 3.0 m^3 s^{-1} but the sediment discharge fell to 107 kg s^{-1}. Although this type of evidence clearly points to the importance of antecedent events in conditioning erosion, no relationship was obtained between soil loss and antecedent precipitation in mid-Bedfordshire, England (Morgan et al. 1986).

The question arises of how much rain is required to induce significant erosion. Hudson (1981) gives a figure, based on his studies in Zimbabwe, of 25 mm h^{-1}, a value that has also been found appropriate in Tanzania (Rapp et al. 1972a) and Malaysia (Morgan 1974). It is too high for western Europe, however, where it is only rarely exceeded. Arbitrary thresholds of 10, 6 and 1.0 mm h^{-1} have been used in England (Morgan 1980a), Germany (Richter & Negendank 1977) and Belgium (Bollinne 1977) respectively.

Threshold values vary with the erosion process. The figures quoted above are typical for erosion by overland flow, rills and mass movements, which, as seen in Chapter 1, is characteristic of moderate events, whereas higher magnitude events are required for the initiation of fresh gullies. The distinction between these types of event can be blurred, however, in those areas that, by world standards, regularly experience what would be described elsewhere as extreme events. Starkel (1972) stresses the importance of regular gully erosion in the Assam Uplands where monthly rainfall may total 2000–5000 mm and in the Darjeeling Hills where over 50 mm of rain falls on an average of 12 days each year and rainfall intensities are often highest at the end of the rain event. In this very active landscape, overland flow and slope wash can start during rain storms of 50 mm with intensities greater than 30 mm h^{-1}; slides and slumps can occur after daily rains of 100 to 150 mm or a rainfall total of 200 mm in two or three days; and debris flows and mudflows are generated when 500 and 1000 mm of rain fall within two or three days (Froehlich & Starkel 1993). The effects of an extreme event may be long lasting and give rise to high soil losses for a number of years. The length of time required for an area to recover from a severe rainstorm, flooding and gullying has not been fully investigated but, in a review of somewhat sparse evidence, Thornes (1976) quotes figures up to 50 years.

3.1.2 Rainfall erosivity indices

The most suitable expression of the erosivity of rainfall is an index based on the kinetic energy of the rain. Thus the erosivity of a rainstorm is a function of its intensity and duration, and of

the mass, diameter and velocity of the raindrops. To compute erosivity requires an analysis of the drop-size distributions of rain. Laws and Parsons (1943), based on studies of rain in the eastern USA, show that the drop-size characteristics vary with the intensity of the rain, with the median drop diameter by volume (d_{50}) increasing with rainfall intensity. Studies of tropical rainfall (Hudson 1963) indicate that this relationship holds only for rainfall intensities up to 100 mm h^{-1}. At higher intensities, median drop size decreases with increasing intensity, presumably because greater turbulence makes larger drop sizes unstable. However, at intensities above 200 mm h^{-1}, coalescence of smaller drops takes place so that the median drop diameter begins to increase again (Carter et al. 1974). Considerable variability exists because the relationship between median drop size and intensity is not constant; both median drop size and drop-size distribution vary for rains of the same intensity but different origins (Mason & Andrews 1960; Carter et al. 1974; Kinnell 1981; McIsaac 1990). The drop-size characteristics of convectional and frontal rain differ, as do those of rain formed at the warm and cold fronts of a temperate depression system.

Despite the difficulties posed by these variations, it is possible to derive general relationships between kinetic energy and rainfall intensity. Based on the work of Laws and Parsons (1943), Wischmeier and Smith (1958) obtained the equation:

$$KE = 0.0119 + 0.0873 \log_{10} I \tag{3.1}$$

where I is the rainfall intensity (mm h^{-1}) and KE is the kinetic energy (MJ ha^{-1} mm^{-1}). Many researchers (Mason & Ramandham 1953; Carte 1971; Houze et al. 1979; Styczen & Høgh-Schmidt 1988) consider the drop-size distribution of rainfall described by Marshall and Palmer (1948) as representative of a wide range of environments. The equivalent formula for calculating kinetic energy is:

$$KE = 0.0895 + 0.0844 \log_{10} I \tag{3.2}$$

For tropical rainfall, Hudson (1965) gives the equation:

$$KE = 0.298\left(1 - \frac{4.29}{I}\right) \tag{3.3}$$

based on measurements of rainfall properties in Zimbabwe.

Given the variability of rainfall characteristics across the globe, it is not surprising that a large number of relationships have been established by different workers in different countries (Table 3.2). Many of these studies show that at intensities greater than 75 mm h^{-1}, the kinetic energy levels off at a value of about 0.29 MJ ha^{-1} mm^{-1} (Kinnell 1987). However, much higher values of 0.34–0.38 have been obtained in northern Nigeria (Kowal & Kassam 1976; Osuji 1989), Tuscany, Italy (Zanchi & Torri 1980), Okinawa, Japan (Onaga et al. 1988), Cévennes, France (Sempere-Torres et al. 1992), Portugal (Coutinho & Tomás 1995), Hong Kong (Jayawardena & Rezaur 2000) and Spain (Cerro et al. 1998). Carter et al. (1974) found that in the southern USA the kinetic energy increased to a maximum value at about 75 mm h^{-1}, decreased with further increases in intensity up to about 175 mm h^{-1} and then increased again at still higher intensities. In contrast, other studies in Japan (Mihara 1951) and in the Marshall Islands (McIsaac 1990) have recorded rainfall energies some 6–20 per cent lower than those calculated from eqn 3.1. Rainfall energy also varies with the density of the air raised to the 0.9 power; as a result, energy increases with altitude. Tracy et al. (1984) found that the kinetic energy of rainfall at 900–1800 m above sea level in Arizona was about 15 per cent higher than that predicted by eqn 3.1. Based on a review of previous research, van Dijk et al. (2002) proposed the following as a general equation:

Table 3.2 Relationship between kinetic energy of rain (KE, $MJ\,ha^{-1}\,mm^{-1}$) and rainfall intensity (I, $mm\,h^{-1}$)

Equation	Source
$E = 0.0119 + 0.0873\log_{10}I$	Used in Universal Soil Loss Equation (Wischmeier & Smith 1978); based on drop-size distributions of rainfall measured by Laws and Parsons (1943)
$E = 0.29(1 - 0.72e^{-I/20})$	Used in Revised Universal Soil Loss Equation; Brown and Foster (1987)
$E = 0.0895 + 0.0844\log_{10}I$	Based on drop-size distributions of rainfall measured by Marshall and Palmer (1948)
$E = 0.0981 + 0.1125\log_{10}I$	Zanchi and Torri (1980) for Toscana, Italy
$E = 0.359(1 - 0.56e^{-0.034I})$	Coutinho and Tomás (1995) for Portugal
$E = 0.0981 + 0.106\log_{10}I$	Onaga et al. (1988) for Okinawa, Japan
$E = 0.298(1 - 4.29/I)$	Hudson (1965) for Zimbabwe
$E = 0.29(1 - 0.6e^{-0.04I})$	Rosewell (1986) for New South Wales, Australia
$E = 0.26(1 - 0.7e^{-0.035I})$	Rosewell (1986) for southern Queensland, Australia
$E = 0.1132 + 0.0055I - 0.005 \times 10^{-2} I^2 + 0.00126 \times 10^{-4} I^3$	Carter et al. (1974) for south-central USA
$E = 0.384(1 - 0.54e^{-0.029I})$	Cerro et al. (1998) for Barcelona, Spain
$E = 0.369(1 - 0.69e^{-0.038I})$	Jayawardena and Rezaur (2000) for Hong Kong
$E = 0.283(1 - 0.52e^{-0.042I})$	Proposed by van Dijk et al. (2002) as a universal relationship

$$KE = 0.283(1 - 0.52^{-0.042I}) \tag{3.4}$$

This relationship generally provides estimates to within 10 per cent of measured values, although it overpredicts in climates with a strong coastal influence and underpredicts in semi-arid and sub-humid areas.

To compute the kinetic energy of a storm, a trace of the rainfall from an automatically recording rain gauge is analysed and the storm divided into small time increments of uniform intensity. For each time period, knowing the intensity of the rain, the kinetic energy of the rain at that intensity is estimated using an equation from Table 3.2 and this, multiplied by the amount of rain received, gives the kinetic energy for that time period. The sum of the kinetic energy values for all the time periods gives the total kinetic energy for the storm (Table 3.3).

To be valid as an index of potential erosion, an erosivity index must be significantly correlated with soil loss. Wischmeier and Smith (1958) found that soil loss by splash, overland flow and rill erosion is related to a compound index of kinetic energy (E) and the maximum 30-minute intensity (I_{30}). This index, known as EI_{30}, is open to criticism. First, being based on estimates of kinetic energy using eqn 3.1, it is of suspect validity for tropical rains of high intensity as well as for high altitudes and for oceanic areas like the Marshall Islands, where rainfall energies are rather low. Second, it assumes that erosion occurs even with light intensity rain, whereas Hudson (1965) showed that erosion is almost entirely caused by rain falling at intensities greater than $25\,mm\,h^{-1}$. The inclusion of I_{30} in the index is an attempt to correct for overestimating the importance of light intensity rain but it is not entirely successful because the ratio of intense erosive rain to non-erosive rain is not well correlated with I_{30} (Hudson, personal communication). In fact, there is no obvious reason why the maximum 30-minute intensity is the most appropriate parameter to choose. Stocking and Elwell (1973a) recommend its use only for bare soil conditions. With sparse

Table 3.3 Calculation of erosivity

Time from start of storm (min)	Rainfall (mm)	Intensity ($mm\,h^{-1}$)	Kinetic energy ($MJ\,ha^{-1}\,mm^{-1}$)	Total kinetic energy (col 2 × col 4) ($MJ\,ha^{-1}$)
0–14	1.52	6.08	0.0877	0.1333
15–29	14.22	56.88	0.2755	3.9180
30–44	26.16	104.64	0.2858	7.4761
45–59	31.50	126.00	0.2879	9.0674
60–74	8.38	33.52	0.2599	2.1776
75–89	0.25	1.00	0	0

Erosivity indices

Wischmeier index (EI_{30})

Maximum 30-minute rainfall	$= 26.16 + 31.50\,mm$
	$= 57.66\,mm$
Maximum 30-minute intensity	$= 57.66 \times 2$
	$= 115.32\,mm\,h^{-1}$
Total kinetic energy	= total of column 5
	$= 22.7724\,MJ\,ha^{-1}$
EI_{30}	$= 22.7724 \times 115.32$
	$= 2262.12\,MJ\,mm\,ha^{-1}\,h^{-1}$

Hudson index ($KE > 25$)

Total kinetic energy for rainfall intensity $\geq 25\,mm\,h^{-1}$	= total of lines 2, 3, 4 and 5 in column 5
	$= 22.64\,MJ\,ha^{-1}$

and dense plant covers they obtain better correlations with soil loss using the maximum 15- and 5-minute intensities respectively. The maximum 5-minute intensity was also found superior to I_{30} for short duration storms in Mediterranean countries (Usón & Ramos 2001). In order to overcome the likelihood of overestimating soil loss from high-intensity rainfall, the recommended practice with the EI_{30} index is to use a maximum value of $0.28\,MJ\,ha^{-1}\,mm^{-1}$ for the E component for all rains above $76.2\,mm\,h^{-1}$ and a maximum value of $63.5\,mm\,h^{-1}$ for the I_{30} term (Wischmeier & Smith 1978).

As an alternative erosivity index, Hudson (1965) proposed $KE > 25$, which, to compute for a single storm, means summing the kinetic energy received in those time increments when the rainfall intensity equals or exceeds $25\,mm\,h^{-1}$ (Table 3.3). When applied to data from Zimbabwe, a better correlation was obtained between this index and soil loss than between soil loss and EI_{30}. Stocking and Elwell (1973a) reworked Hudson's data, taking account of more recent data, and suggested that EI_{30} was the better index after all. But, since they computed EI_{30} only for storms yielding 12.5 mm or more of rain and with a maximum 5-minute intensity greater than $25\,mm\,h^{-1}$, they removed most of the objections to the original EI_{30} index and produced an index that is philosophically very close to $KE > 25$. Hudson's index has the advantage of simplicity and less stringent data requirements. Although somewhat limiting for temperate latitudes, it can be modified by using a lower threshold value such as $KE > 10$.

By calculating erosivity values for individual storms over a period of 20–25 years, mean monthly and mean annual data can be obtained. Since the EI_{30} and $KE > 25$ indices yield vastly different values because of the inclusion of I_{30} in the former, the two indices cannot be substituted for each other.

3.1.3 Wind erosivity

The kinetic energy (KE_a; $\mathrm{J\,m^{-2}\,s^{-1}}$) of wind can be calculated from:

$$KE_a = \frac{\gamma_a u^2}{2g} \tag{3.5}$$

where u is the wind velocity ($\mathrm{m\,s^{-1}}$) and γ_a is the specific weight of air defined in terms of temperature (T) in °C and barometric pressure (P) in kPa by the relationship (Zachar 1982):

$$\gamma_a = \frac{1.293}{1+0.00367T}\cdot\frac{P}{101.3} \tag{3.6}$$

For $T = 15$ °C and $P = 101.3$ kPa, $\gamma_a = 0.0625\,u^2\,\mathrm{kg\,m^{-2}}$, which converts to $227\,u^2\,\mathrm{J\,m^{-2}\,s^{-1}}$. Energy values for wind storms can be obtained by summing the energies for the different velocities weighted by their duration.

In practice, kinetic energy is rarely used as an index of wind erosivity and, instead, a simpler index based on the velocity and duration of the wind (Skidmore & Woodruff 1968) has been developed. The erosivity of wind blowing in vector j is obtained from:

$$EW_j = \sum_{i=1}^{n} \overline{V}t_{ij}^3 f_{ij} \tag{3.7}$$

where EW_j is the wind erosivity value for vector j, $\overline{V}t$ is the mean velocity of wind in the ith speed for vector j above a threshold velocity, taken as $19\,\mathrm{km\,h^{-1}}$, and f_{ij} is the duration of the wind for vector j at the ith speed. Expanding this equation for total wind erosivity (EW) over all vectors yields:

$$EW = \sum_{j=0}^{15}\sum_{i=1}^{n} \overline{V}t_{ij}^3 f_{ij} \tag{3.8}$$

where vectors $j = 0$ to 15 represent the 16 principal compass directions beginning with $j = 0 =$ E and working anticlockwise so that $j = 1 =$ ENE and so on.

3.2 Erodibility

Erodibility defines the resistance of the soil to both detachment and transport. Although a soil's resistance to erosion depends in part on topographic position, slope steepness and the amount of disturbance, such as during tillage, the properties of the soil are the most important determinants. Erodibility varies with soil texture, aggregate stability, shear strength, infiltration capacity and organic and chemical content.

The role of soil texture has been indicated in Chapter 2, where it was shown that large particles are resistant to transport because of the greater force required to entrain them and that fine particles are resistant to detachment because of their cohesiveness. The least resistant particles are silts and fine sands. Thus soils with a silt content above 40 per cent are highly erodible (Richter & Negendank 1977). Evans (1980) prefers to examine erodibility in terms of clay content, indicating that soils with a clay content between 9 and 30 per cent are the most susceptible to erosion.

The use of the clay content as an indicator of erodibility is theoretically more satisfying because the clay particles combine with organic matter to form soil aggregates or clods and it is the stability of these that determines the resistance of the soil.

Soils with a high content of base minerals are generally more stable, as these contribute to the chemical bonding of the aggregates. Wetting of the soil weakens the aggregates because it lowers their cohesiveness, softens the cements and causes swelling as water is adsorbed on the clay particles. Rapid wetting can also cause collapse of the aggregates through slaking. The wetting-up of initially dry soils results in greater aggregate breakdown than if the soil is already moist because, in the latter case, less air becomes trapped in the soil (Truman et al. 1990). Aggregate stability also depends on the type of clay mineral present. Soils containing kaolinite, halloysite, chlorite or fine-grained micas, all of which are resistant to expansion on wetting, have a low level of erodibility, whereas soils with smectite or vermiculite swell on wetting and therefore have a high erodibility; soils with illite are in an intermediate position.

In detail, however, the interactions between the moisture content of the soil and the chemical composition of both the clay particles and the soil water are rather complex. This makes it difficult to predict how clays, particularly those susceptible to swelling, will behave. The erodibility of clay soils is highly variable (Chisci et al. 1989). Although most clays lose strength when first wetted because the free water releases bonds between the particles, some clays, under moist but unsaturated conditions, regain strength over time. This process, known as thixotropic behaviour, occurs because the hydration of clay minerals and the adsorption of free water promote hydrogen bonding (Grissinger & Asmussen 1963). Strength can also be regained if swelling brings about a reorientation of the soil particles from an alignment parallel to the eroding water to a more random orientation (Grissinger 1966). The strength of smectitic clays is largely dependent upon the sodium adsorption ration (SAR). As this increases, i.e. the replacement of calcium and magnesium ions by sodium increases, so do water uptake and the likelihood of swelling and aggregate collapse. High salt concentrations in the soil water, however, can partly offset this effect so that aggregate stability is maintained at higher levels of SAR (Arulanandan et al. 1975). Sodic and saline-sodic soils, where the exchangeable sodium percentage (ESP) exceeds $15\,\mathrm{cmol\,kg^{-1}}$ or the SAR of the pore water exceeds 13, are highly erodible.

The shear strength of the soil is a measure of its cohesiveness and resistance to shearing forces exerted by gravity, moving fluids and mechanical loads. Its strength is derived from the frictional resistance met by its constituent particles when they are forced to slide over one another or to move out of interlocking positions, the extent to which stresses or forces are absorbed by solid-to-solid contact among the particles, cohesive forces related to chemical bonding of the clay minerals and surface tension forces within the moisture films in unsaturated soils. These controls over shear strength are only understood qualitatively, so that, for practical purposes, shear strength is expressed by an empirical equation:

$$\tau = c + \sigma \tan\phi \tag{3.9}$$

where τ is the shear stress required for failure to take place, c is a measure of cohesion, σ is the stress normal to the shear plane (all in units of force per unit area) and ϕ is the angle of internal friction. Both c and ϕ are best regarded as empirical parameters rather than as physical properties of the soil.

Increases in the moisture content of a soil decrease its shear strength and bring about changes in its behaviour. At low moisture contents the soil behaves as a solid and fractures under stress, but with increasing moisture content it becomes plastic and yields by flow without fracture. The point of change in behaviour is termed the plastic limit. With further wetting, the soil will reach its liquid limit and start to flow under its own weight. The behaviour of a compressible soil when saturated depends on whether the water can drain. If drainage cannot take place and the soil is subjected to further loading, pressure will increase in the soil water, the compaction load will not be supported by the particles and the soil will deform, behaving as a plastic material. If drainage can occur, more of the load will be supported and the soil is more likely to remain below the plastic limit and retain a higher shear strength. As seen in Chapter 2, shear strength is used as a basis for understanding the detachability of soil particles by raindrop impact. Since the soils are usually saturated and the process is virtually instantaneous, there is no time for drainage and undrained failure occurs. Bradford et al. (1992) found that soil strength measured with a drop-cone penetrometer after one hour of rainfall was a good indicator of a soil's resistance to splash erosion. The drop-cone apparatus simulates the same kind of failure mechanism, in terms of compression and shear, as the impact of a falling raindrop.

The mechanism of soil particle detachment by surface flow involves different failure stresses on the soil surface compared with those generated by raindrop impact. Rauws and Govers (1988) show that these can be represented by measurements of the strength of the soil at saturation, made with a torvane. Equations 2.30 and 2.31 predict the critical shear velocity (u_{*c}; $cm\,s^{-1}$) for rill initiation on a smooth bare soil surface as a function of the strength or apparent cohesion of the soil measured at saturation by a torvane and a laboratory shear vane respectively.

Infiltration capacity, the maximum sustained rate at which soil can absorb water, is influenced by pore size, pore stability and the form of the soil profile. Soils with stable aggregates maintain their pore spaces better, while soils with swelling clays or minerals that are unstable in water tend to have low infiltration capacities. Although estimates of the infiltration capacity can be obtained in the field using infiltrometers (Hills 1970), it was seen in Chapter 2 that actual capacities during storms are often much less than those indicated by field tests. Where soil properties vary with profile depth, it is the horizon with the lowest infiltration capacity that is critical. For sandy and loamy soils, the critical horizon is often the surface where, as described in Chapter 2, a crust of 2 mm thickness may be sufficient to decrease infiltration capacity enough to cause runoff, even though the underlying soil may be dry.

The organic and chemical constituents of the soil are important because of their influence on aggregate stability. Soils with less than 2 per cent organic carbon, equivalent to about 3.5 per cent organic content, can be considered erodible (Evans 1980). Most soils contain less than 15 per cent organic content and many sands and sandy loams have less than 2 per cent. Voroney et al. (1981) suggest that soil erodibility decreases linearly with increasing organic content over the range of 0–10 per cent, whereas Ekwue (1990) found that soil detachment by raindrop impact decreased exponentially with increasing organic content over a 0–12 per cent range. These relationships cannot be extrapolated, however, because some soils with very high organic contents, particularly peats, are highly erodible by wind and water, whereas others with very low organic content can become very hard and therefore stronger under dry conditions. The role played by organic material depends on its origin. While organic material from grass leys and farmyard manure contributes to the stability of the soil aggregates, peat and undecomposed haulm merely protect the soil by acting like a mulch and do little to increase aggregate strength (Ekwue et al. 1993). Thus peat soils have very low aggregate stability.

Chemically, the most important control over erodibility is the proportion of easily dispersible clays in the soil. As seen above, a high proportion of exchangeable sodium can cause rapid deterioration in a soil's structure on wetting, with consequent loss of strength, followed by the formation of a surface crust and a decline in infiltration as the detached clay particles fill the pore spaces in the soil. The addition of sodium-containing fertilizers to support crops such as tobacco can sometimes lead to quite small increases in exchangeable sodium yet result in very marked structural deterioration of a previously stable soil (Miller & Sumner 1988). Excess calcium carbonate within the clay and silt fractions of the soil also leads to high erodibility (Barahona et al. 1990; Merzouk & Blake 1991).

Many attempts have been made to devise a simple index of erodibility based on either the properties of the soil as determined in the laboratory or the field, or the response of the soil to rainfall and wind (Table 3.4). In a review of the indices related to water erosion, Bryan (1968) favoured aggregate stability as the most efficient index. Unfortunately, there is no agreement between researchers on the most appropriate method to evaluate aggregate stability. Indices like the instability index (*Is*) and the pseudo-textural aggregation index (*Ipta*) (Table 3.4) are based on breaking up the aggregates by wet-sieving a sample of the soil. But some researchers consider that wet-sieving does not adequately simulate the processes of breakdown as they occur in the field and prefer to measure the proportion of aggregates that can be destroyed by the impact of water drops (Bruce-Okine & Lal 1975). Different researchers also follow different methods for the duration and speed of oscillation of the sieves in wet-sieving tests, and for the size and height of fall in water-drop tests. Further work on developing an appropriate test probably needs to take account of the factors that contribute to the stability of the aggregates. These are, respectively: for aggregates >10 mm in size, the binding and adhesive effects of plant roots; for aggregates of 2–10 mm, the calcium carbonate and organic matter content; for aggregates of 1–2 mm, the network of roots and hyphae; for aggregates of 0.105–1.0 mm, organic matter, roots and hyphae; and for aggregates <0.105 mm, clay mineralogy and cementing agents derived from microbiological activity (Boix-Fayos et al. 2001).

Given the above, it is not surprising that attempts have been made to develop a more universally accepted index. The one most commonly used is the *K* value which represents the soil loss per unit of EI_{30} as measured in the field on a standard bare soil plot, 22 m long and at 5° slope. Estimates of the *K* value may be made if the grain-size distribution, organic content, structure and permeability of the soil are known (Wischmeier et al. 1971; Fig. 3.1). Soil erodibility has been satisfactorily described by the *K* value for many agricultural soils in the USA (Wischmeier & Smith 1978) and for ferrallitic and ferruginous soils in West Africa (Roose 1977). Where *K* values have been determined from field measurements of erosion, they are valid. Difficulties arise, however, with attempts to predict the values from the nomograph (Fig. 3.1). Where it is applied to soils with similar characteristics to those in the USA, a close correlation exists between predicted and measured values, but poorer predictions are obtained where it is necessary to extrapolate the nomograph values. This applies to soils with organic contents above 4 per cent, swelling clays and those where resistance to erosion is a function of aggregate stability rather than primary particle size.

The resistance of the soil to wind erosion depends upon dry rather than wet aggregate stability and on the moisture content, wet soil being less erodible than dry soil, but is otherwise related to much the same properties as affect its resistance to water erosion. Chepil (1950), using wind tunnel experiments, related wind erodibility of soils to various indices of dry aggregate structure, but little of the work was tested in field conditions. Nevertheless, the results were extrapolated to give an index in $t\,ha^{-1}\,yr^{-1}$ based on climatic data for Garden City, Kansas. Values range from 84–126 for non-calcareous silty clay loams, silt loams and loams to 356–694 for sands. Similar research by Dolgilevich et al. (1973) in western Siberia and northern Kazakhstan yields values in $t\,ha^{-1}\,h^{-1}$ (Table 3.5). Both indices are closely related to the percentage of dry stable aggregates larger than 0.84 mm.

The indices described above treat soil erodibility as constant over time. They thus ignore seasonal variations on agricultural land associated with tillage operations, which alter the bulk density and hydraulic conductivity of the soil. Erodibility is four times higher in summer than in

Table 3.4 Indices of soil erodibility for water erosion

Static laboratory tests		
Dispersion ratio	$\frac{\%silt + \%clay\ in\ undispersed\ soil}{\%silt + \%clay\ after\ dispersal\ of\ the\ soil\ in\ water}$	Middleton (1930)
Clay ratio	$\frac{\%sand + \%silt}{\%clay}$	Bouyoucos (1935)
Surface aggregation ratio	$\frac{surface\ area\ of\ particles > 0.05mm}{(\%silt + \%clay\ in\ dispersed\ soil) - (\%silt + \%clay\ in\ undispersed\ soil)}$	André and Anderson (1961)
Erosion ratio	$\frac{dispersion\ ratio}{colloid\ content/moisture\ equivalent\ ratio}$	Lugo-Lopez (1969)
Instability index (*Is*)	$\frac{\%silt + \%clay}{Ag_{air} + Ag_{alc} + Ag_{benz}}$ where *Ag* is the % aggregates >0.2 mm after wet sieving for no pretreatment and pretreatment of the soil by alcohol and benzine respectively	Hénin et al. (1958)
Instability index (*Is*)	$\frac{\%silt + \%clay}{(\%aggregates > 0.2mm\ after\ wet\ sieving) - 0.9(\%coarse\ sand)}$	Combeau and Monnier (1961)
Pseudo-textural aggregation index (*Ipta*)	$\frac{MWDw - MWDt}{X - MWDt} \times 100$ where *MWDw* is the mean weight diameter of the wet-sieving grain-size distribution (mm), *MWDt* is the mean weight diameter of the primary particle grain-size distribution (mm) and *X* is the maximum average grain-size diameter of the particles in the given grain-size distribution	Chischi et al. (1989)
Static field tests		
Erodibility index	$\frac{1}{mean\ shearing\ resistance \times permeabilty}$	Chorley (1959)
Soil cohesion	direct measure of soil cohesion at saturation using a torvane	Rauws and Govers (1988)
Dynamic laboratory tests		
Simulated rainfall test	Comparison of erosion of different soils subjected to a standard storm	Woodburn and Kozachyn (1956)
Water-stable aggregate (*WSA*) content	%*WSA* > 0.5 mm after subjecting the soil to rainfall simulation	Bryan (1968)
Water drop test	% aggregates destroyed by a pre-selected number of impacts by a standard raindrop (e.g. 5.5 mm diameter, 0.1 g from a height of 1 m)	Bruce-Okine and Lal (1975)
Erosion index	$\frac{dh}{a}$ where *d* is an index of dispersion (ratio of % particles >0.05 mm without dispersion to % particles >0.05 mm after dispersion of the soil by sodium chloride); *h* is an index of water-retaining capacity (water retention of soil relative to that of 1 g of colloids); and *a* is an index of aggregation (% aggregates >0.25 mm after subjecting the soil to a water flow of $100\ cm\ min^{-1}$ for 1 h)	Voznesensky and Artsruui (1940)
Dynamic field tests		
Erodibility index (*K*)	mean annual soil loss per unit of EI_{30}	Wischmeier and Mannering (1969)

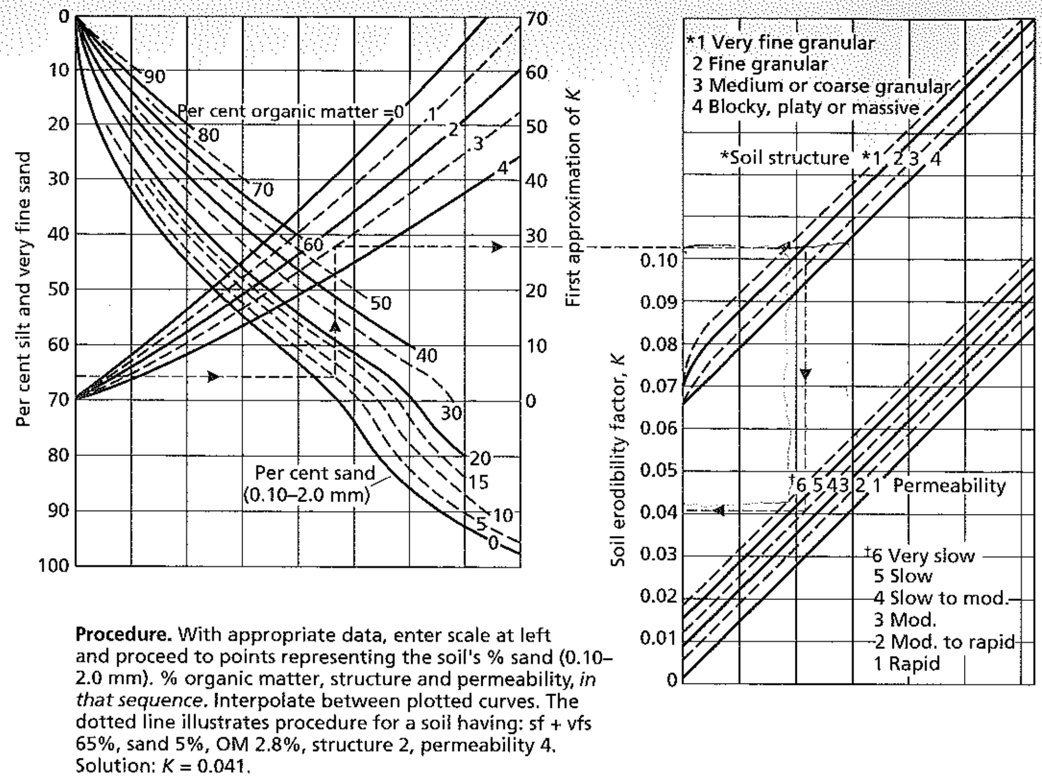

Fig. 3.1 Nomograph for computing the K value (metric units) of soil erodibility for use in the Universal Soil Loss Equation (after Wischmeier et al. 1971). Divide values by 0.13 to obtain K values in the original American units.

Table 3.5 Assessments of soil erodibility by wind

% dry stable aggregates >0.84 mm	>80	70–80	50–70	20–50	<20
Erodibility ($t\,ha^{-1}\,h^{-1}$)*	<0.5	0.5–1.5	1.5–5	5–15	>15
Erodibility($t\,ha^{-1}\,yr^{-1}$)†	<4	4–84	84–166	166–220	>220

* After Dolgilevich et al. (1973) for windspeeds of 20–25 $m\,s^{-1}$.
† After Chepil (1960) for Garden City, Kansas.

winter on bare, uncultivated sandy soil in Bedfordshire, England (Martin & Morgan 1980) and two times higher on silts and silt loam soils in Limbourg, The Netherlands (Kwaad 1991). There are also more frequent changes in erodibility related to changes in moisture content during and between rainstorms. While the expectation is that most soils become more erodible when they are wet because of aggregate destruction during the wetting-up process and the loss of cohesion, some soils are also very erodible when dry and more susceptible to detachment by raindrop impact (Martínez-Mena et al. 1998) and rilling (Govers 1991). A key factor for coarser soils is

their tendency to become hydrophobic when dry, which then leads to an increase in runoff and, until the depth of water becomes sufficient to absorb their impact, an increase in soil particle detachment by raindrops (Terry & Shakesby 1993; Doerr et al. 2000). Freezing and thawing also alter the erodibility of the soil. Conditions of low bulk density and high soil moisture during periods of thaw produce a surface that is highly erodible. The erodibility of agricultural soils in Ontario, Canada, is 15 times higher in winter thaw conditions than in summer (Coote et al. 1988).

Under natural conditions, the seasonal activity of burrowing animals is important, giving rise to considerable disturbance of the soil. Earthworms bring to the surface as casts as much as 2–5.8 $t\,ha^{-1}$ of material on agricultural land (Evans 1948). Rates of 15 $t\,ha^{-1}$ have been measured in temperate woodland in Luxembourg (Hazelhoff et al. 1981) and 50 $t\,ha^{-1}$ in tropical forest in the Ivory Coast (Roose 1976). Other animals and their annual rates of production of sediment at the surface include: ants, with 4 to 10 $t\,ha^{-1}$ observed in Utah (Thorp 1949); termites, with 1.2 $t\,ha^{-1}$ in tropical forest in the Ivory Coast (Roose 1976); voles and moles, with 19 $t\,ha^{-1}$ in temperate woodland in Luxembourg (Imeson 1976) and 6 $t\,ha^{-1}$ from Pyrenean mountain voles in the Spanish Pyrenees (Borghi et al. 1990); and isopods and porcupines, with 0.3 to 0.7 $t\,ha^{-1}$ on stony land in the Negev Desert, Israel (Yair & Rutin 1981). On coastal sand dunes in the western Netherlands, rabbits displaced locally between 0.9 and 5.1 $t\,ha^{-1}$ of sediment from their burrows (Rutin 1992). In many cases, the material brought to the surface comprises loose sediment with low bulk density and cohesion, which is rapidly broken down by splash erosion. The material contained in earthworm casts, however, consists of soil aggregates that are more stable under raindrop impact than the surrounding top soil, probably as a result of their higher organic content and secretions from the gut of the worm. Thus, earthworms have a positive effect on the stability of soil aggregates and the hydraulic conductivity of the soil (Glasstetter & Prasuhn 1992).

3.3 Effect of slope

Erosion would normally be expected to increase with increases in slope steepness and slope length as a result of respective increases in velocity and volume of surface runoff. Further, while on a flat surface raindrops splash soil particles randomly in all directions, on sloping ground more soil is splashed downslope than upslope, the proportion increasing as the slope steepens. The relationship between erosion and slope can be expressed by the equation:

$$E \propto \tan^m \theta L^n \qquad (3.10)$$

where E is soil loss per unit area, θ is the slope angle and L is slope length. Zingg (1940), in a study of data from five experimental stations of the United States Soil Conservation Service, found that the relationship had the form:

$$E \propto \tan^{1.4} \theta L^{0.6} \qquad (3.11)$$

To express E proportional to distance downslope, the value of n must be increased by 1.0. Since the values for the exponents have been confirmed in respect of m by Musgrave (1947) and m and n by Kirkby (1969b), there is some evidence to suggest that eqn 3.11 has general validity. Other studies, however, show that the values are sensitive to the interaction of other factors.

3.3.1 Exponents for slope steepness

Working with data from experimental stations in Zimbabwe, Hudson and Jackson (1959) found that m was close to 2.0 in value, indicating that the effect of slope is stronger under tropical conditions where rainfall is heavier. The effect of soil is illustrated by the laboratory experiments of Gabriels et al. (1975), who showed that m increases in value with the grain size of the material, from 0.6 for particles of 0.05 mm diameter to 1.7 for particles of 1.0 mm. The value also changes with slope, decreasing from 1.6 on slopes of 0–2.5° to 0.7 on slopes between 3 and 6.5°, to 0.4 on slopes over 6.5° (Horváth & Erődi 1962). On steeper slopes, the value may be expected to decrease further as soil-covered slopes give way to rock surfaces and soil supply becomes a limiting factor. In a detailed study of soil loss from 33 road-cut slopes on the Benin–Lagos Highway in Nigeria, Odermerho (1986) found values of $m = 1.09$ for slopes between 1.4 and 6°, 1.80 for slopes between 6 and 8.5°, −2.18 for slopes between 8.5 and 11° and −1.39 for slopes between 11 and 26.5°. Combining the results of these studies suggests a curvilinear relationship between soil loss and slope steepness, with erosion initially increasing rapidly as slope increases from gentle to moderate, reaching a maximum on slopes of about 8–10° and then decreasing with further increases in slope. Such a relationship would apply only to erosion by rainsplash and surface runoff; it would not apply to landslides, piping or gully erosion by pipe collapse.

The exponents in eqn 3.11 also vary in value with slope shape. D'Souza and Morgan (1976) obtained values of $m = 0.5$ on convex slopes, 0.4 on straight slopes and 0.14 on concave slopes. No studies have been made of the effect of changes of slope in plan, but Jackson (1984) found from erosion surveys and laboratory experiments that discharge varies with an index of contour curvature to the power of 5.5. If soil loss is assumed to vary with the square of the discharge, the value of m becomes 3.5. Contour curvature is here defined as the proportion of a circle centred on a point on a hillside that lies at a higher altitude than that point. The index ranges from 0 to 1 in value with values <0.5 indicating diverging slopes, a value of 0.5 a straight slope and values >0.5 converging slopes.

Few studies have looked at the effect of variations in plant cover. Quinn et al. (1980) investigated the change in the value of m for soil loss from 1.2 m long plots with slopes of 5–30° under simulated rainfall in relation to decreasing grass cover brought about by human trampling. They found that m was 0.7 for fully grassed slopes, rose to 1.9 in the early stages of trampling and fell to 1.1 when only about 25 per cent of the grass cover remained. Lal (1976) obtained a value of $m = 1.1$ for both bare fallow and maize on erosion plots in Nigeria; use of a mulch, however, reduced m to 0.5.

The exponent values also vary with the process of erosion: $m = 1.0$ for soil creep, ranges between 1.0 and 2.0 for splash erosion and between 1.3 and 2.0 for erosion by overland flow, and may be as high as 3.0 for rivers (Kirkby 1971). It was shown in Chapter 2 that m was about 0.3–1.0 in value for rainsplash and about 0.7 and 1.7–2.0 respectively for detachment and transport of soil particles by overland flow.

Increases in slope steepness may also cause an increase in the intensity of wind erosion on windward slopes and on the crests of knolls. Data from Chepil et al. (1964) and Stredňanský (1977) show that $m = 0.4$ for slopes up to 2° and 1.2 for slopes from 2 to 15°. The increase in value is attributed to increases in wind speed, shear and turbulence close to the ground as the air moves upslope (Livingstone and Warren, 1996).

3.3.2 Exponents for slope length

The value of 0.6 for exponent n applies only to overland flow on slopes about 10–20 m long, with steepnesses greater than 3°. Wischmeier and Smith (1978) propose values of $n = 0.4$ for slopes of 3°, 0.3 for slopes of 2°, 0.2 for slopes of 1° and 0.1 for slopes of less than 1°. Kirkby (1971) suggests that $n = 0$ for soil creep and splash erosion, ranges between 0.3 and 0.7 for overland flow and rises to between 1.0 and 2.0 if rilling occurs. This implies that the value of n will vary with distance along a hillside as, for example, soil creep close to the summit gives way first to overland flow and then to rill flow. Without rills, n may become negative on slopes longer than about 10 m. The increasing depth of overland flow downslope protects the soil from raindrop impact so that, even though the transporting capacity of the overland flow increases, erosion becomes limited by the rate of detachment, which is decreasing with slope length (Gilley et al. 1985). Once rills form, soil loss will either increase with slope length (Meyer et al. 1975), particularly if the density of rills is very high, or decrease because, as the flow becomes concentrated, there is no longer sufficient flow on the interrill areas to remove all the material detached by rainsplash (Abrahams et al. 1991). Erosion may also decrease with increasing slope length if, as the slope steepens, the soil becomes less prone to crusting and infiltration rates remain higher than on the gentler-sloping land at the top of the slope (Poesen 1984). Similarly, if the slope declines in angle as length increases, soil loss may decrease as a result of deposition. Clearly, with such a great range of possible conditions, a single relationship between soil loss and slope length cannot exist.

3.4 Effect of plant cover

Vegetation acts as a protective layer or buffer between the atmosphere and the soil. The above-ground components, such as leaves and stems, absorb some of the energy of falling raindrops, running water and wind, so that less is directed at the soil, while the below-ground components, comprising the root system, contribute to the mechanical strength of the soil.

The importance of plant cover in reducing erosion is demonstrated by the mosquito gauze experiment of Hudson and Jackson (1959), in which soil loss was measured from two identical bare plots on a clay loam soil. Over one plot was suspended a fine wire gauze, which had the effect of breaking up the force of the raindrops, absorbing their impact and allowing the water to fall to the ground from a low height as a fine spray. The mean annual soil loss over a ten-year period was 126.6 t ha^{-1} for the open plot and 0.9 t ha^{-1} for the plot covered by gauze.

Although numerous measurements have been made of erosion under different plant covers for comparison with that from bare ground, there is little agreement on the nature of the relationship between soil loss and changes in the extent of cover. Elwell (1981) favoured an exponential decrease in soil loss with increasing percentage interception of rainfall energy and, therefore, increasing percentage cover. Such a relationship was suggested by Wischmeier (1975) as applicable to covers in direct contact with the soil surface and has been verified experimentally for crop residues (Laflen & Colvin 1981; Hussein & Laflen 1982) and grass covers (Lang & McCaffrey 1984; Morgan et al. 1997a). The relationship can be described by the equation:

$$SLR = e^{-j.PC} \tag{3.12}$$

where SLR is the ratio between the soil loss with the plant cover and that from bare ground, PC is the percentage cover and j varies in value from 0.025 to 0.06, with 0.035 taken as typical (Fig. 3.2). Foster (1982) attributes the exponential form of the relationship for covers in proximity to the ground to the ponding of water behind the plant elements, which reduces the

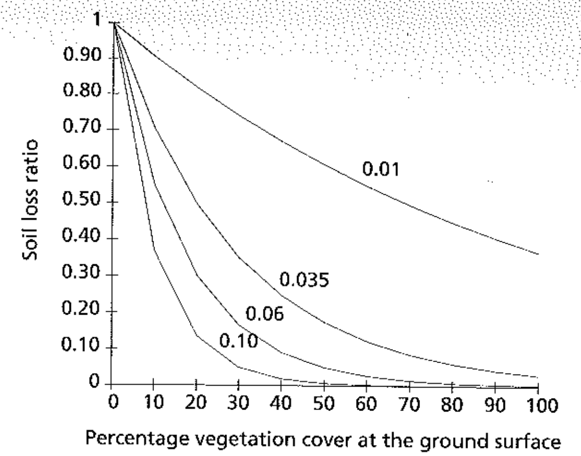

Fig. 3.2 Relationship between soil loss ratio (*SLR*) value and percentage vegetation cover at the ground surface. Curves represent different values of j in eqn 3.12.

effectiveness of the raindrop impact. For plant covers where the leaves and stems are not in contact with the soil but form a canopy at some height above the soil surface, the soil loss ratio is conventionally considered to reduce in a linear relationship with increasing percentage cover (Wischmeier & Smith 1978; Dissmeyer & Foster 1981) but more research is needed to confirm this. Overall, it is generally recognized that for adequate protection at least 70 per cent of the ground surface must be covered (Elwell & Stocking 1976) although reasonable protection can sometimes be achieved with between 30 and 40 per cent cover. The effects of vegetation, however, are far from straightforward and, under certain conditions, a plant cover can exacerbate erosion depending on how it interacts with the erosion processes.

3.4.1 Effect on rainfall

The effectiveness of a plant cover in reducing erosion by raindrop impact depends upon the height and continuity of the canopy, and the density of the ground cover. The height of the canopy is important because water drops falling from 7 m may attain over 90 per cent of their terminal velocity. Further, raindrops intercepted by the canopy may coalesce on the leaves to form larger drops, which are more erosive. Brandt (1989) showed that, for a wide range of plant types, leaf drips have a mean volume drop diameter between 4.5 and 4.9 mm, which is about twice that of natural rainfall. In contrast, Hall and Calder (1993) found that *Pinus caribaea* and *Eucalyptus camaldulensis* forest in southern India produced median volume drop diameters of only 2.3 and 2.8 mm, very similar to those of natural rainfall. For many types of vegetation canopy, raindrop-size distributions are characteristically bimodal with peaks around 2 and 4.8 mm, corresponding to the direct throughfall and the leaf drainage respectively. The effects of these changes in drop-size distribution have been studied mainly in relation to forest canopies. Chapman (1948) under pine forest in the USA, Wiersum et al. (1979) under *Acacia* forest in Indonesia, Mosley (1982) under beech forest in New Zealand and Vis (1986) under tropical rain forest in Colombia all show that while interception by the canopy reduces the volume of rain reaching the ground surface, it does not significantly alter its kinetic energy, which may even be increased compared with that in open ground. As a result, rates of soil particle detachment by rainsplash under the forest canopies can be between 1.2 and 3.1 times those in open ground, unless the soil surface is protected by a litter layer. It is possible that kinetic energy underplays the importance of leaf drips in contributing to soil detachment because it emphasizes drop velocity over drop size. Indices such as the square of the rainfall momentum (Styczen & Høgh-Schmidt 1988) or the product of momentum and drop diameter (Salles et al. 2000), which assign greater importance to drop size, may describe the process better.

Fewer investigations have been made to assess the effects of lower growing canopies. McGregor and Mutchler (1978) found that while cotton reduced the kinetic energy of rainfall by 95 per cent under the canopy and 75 per cent overall, it was locally increased between the rows where the leaf drips were concentrated. Armstrong and Mitchell (1987) found that the detachment under soya bean was about 94 per cent of that in the open despite the very low canopy height. Somewhat lower detachment rates were measured in the field by Morgan (1985b), who found that, with a 90 per cent cover, detachment under soya bean was 0.2 times that in open ground for a 100 mm h^{-1} rainfall intensity and 0.6 times for a 50 mm h^{-1} intensity. Finney (1984) showed in a laboratory study that detachment rates from leaf drip were 1.7 and 1.3 times those in open ground for Brussels sprouts and sugar beet at 23 and 16 per cent cover respectively. In another laboratory study under Brussels sprouts, Noble and Morgan (1983) found that the average detachment rate was the same as that in open ground. In a field study under maize with 88 per cent canopy cover at a height of 2 m, detachment was 14 times greater than that in open ground for a rainfall intensity of 100 mm h^{-1} and 2.4 times greater for an intensity of 50 mmh (Morgan 1985b).

In addition to modifying the drop-size distribution of the rainfall, a plant canopy changes its spatial distribution at the ground surface. Concentrations of water at leaf drip points can result in very high localized rainfall intensities, which can considerably exceed infiltration capacities and play an important role in the generation of runoff. Under mature soya bean, Amstrong and Mitchell (1987) found that half of the rainfall reaching the ground in a storm of 25 mm h^{-1} did so at intensities greater than those in open ground and that 10 per cent of the rain had an intensity of 385 mm h^{-1}. Stemflow also concentrates rain at the ground surface. De Ploey (1982) found that the effective intensity of stemflow beneath a canopy of tussocky grass was 150–200 per cent greater than the rainfall intensity in open ground. Herwitz (1986) recorded local stemflows of between 830 and 18,878 mm h^{-1} in a rainstorm of 118 mm h^{-1} over a six-minute period in a tropical rain forest in northern Queensland.

3.4.2 Effect on runoff

A plant cover dissipates the energy of running water by imparting roughness to the flow, thereby reducing its velocity. In most soil conservation work, the roughness is expressed as a value of Manning's n, which represents the summation of roughness imparted by the soil particles, surface microtopography (form roughness) and vegetation, acting independently of each other. Typical values of Manning's n are given in Table 3.6 (Petryk & Bosmajian 1975; Temple 1982; Engman 1986). The level of roughness with different forms of vegetation depends upon the morphology and the density of the plants, as well as their height in relation to the depth of flow. When the flow depth is shallow, as with overland flow, the vegetation stands relatively rigid and imparts a high degree of roughness, represented for grasses by n values of 0.25 to 0.3. As flow depths increase, the grass stems begin to oscillate, disturbing the flow and causing n values to increase to around 0.4. With further increases in flow depth, the vegetation is submerged; the plants tend to lie down in the flow and offer little resistance, so that n values decrease rapidly (Ree 1949).

Table 3.6 Guide values for Manning's n

Land use or cover	Manning's n
Bare soil	
Roughness depth <25 mm	0.010–0.030
Roughness depth 25–50 mm	0.014–0.033
Roughness depth 50–100 mm	0.023–0.038
Roughness depth >100 mm	0.045–0.049
Bermuda grass – sparse to good cover	
Very short (<50 mm)	0.015–0.040
Short (50–100 mm)	0.030–0.060
Medium (150–200 mm)	0.030–0.085
Long (250–600 mm)	0.040–0.150
Very long (>600 mm)	0.060–0.200
Bermuda grass – dense cover	0.300–0.480
Other dense sod-forming grasses	0.390–0.630
Dense bunch grasses	0.150
Kudzu	0.070–0.230
Lespedeza	0.100
Natural rangeland	0.100–0.320
Clipped rangeland	0.020–0.240
Wheat straw mulch	
2.5 t ha^{-1}	0.050–0.060
5.0 t ha^{-1}	0.075–0.150
7.5 t ha^{-1}	0.100–0.200
10.0 t ha^{-1}	0.130–0.250
Chopped maize stalks	
2.5 t ha^{-1}	0.012–0.050
5.0 t ha^{-1}	0.020–0.075
10.0 t ha^{-1}	0.023–0.130
Cotton	0.070–0.090
Wheat	0.100–0.300
Sorghum	0.040–0.110
Concrete or asphalt	0.010–0.013
Gravelled surface	0.012–0.030
Chisel-ploughed soil	
<0.6 t ha^{-1} residue	0.006–0.170
0.6–2.5 t ha^{-1} residue	0.070–0.340
2.5–7.5 t ha^{-1} residue	0.190–0.470
Disc-harrowed soil	
<0.6 t ha^{-1} residue	0.008–0.140
0.6–2.5 t ha^{-1} residue	0.100–0.250
2.5–7.5 t ha^{-1} residue	0.140–0.530
No tillage	
<0.6 t ha^{-1} residue	0.030–0.070
0.6–2.5 t ha^{-1} residue	0.010–0.130
2.5–7.5 t ha^{-1} residue	0.160–0.470
Bare mouldboard-ploughed soil	0.020–0.100
Bare soil tilled with coulter	0.050–0.130

Source: after Petryk and Bosmajian (1975), Temple (1982), Engman (1986).

Greatest reductions in velocity occur with dense, spatially uniform, vegetation covers. Clumpy, tussocky vegetation is less effective and may even lead to concentrations in flow with localized high velocities between the clumps. When flow separates around a clump of vegetation, the pressure exerted by the flow is higher on the upstream face than it is downstream, and eddying and turbulence occur immediately downstream of the vegetation. Vortex erosion is induced both upstream and downstream (Babaji 1987). Detailed observations during laboratory experiments on overland flow (De Ploey 1981) show that for slopes above about 8°, erosion under grass is higher than that from an identical plot without grass until the percentage grass cover reaches a critical value. Beyond this point, the grass has the expected protective effect.

3.4.3 Effect on air flow

Vegetation reduces the shear velocity of wind by imparting roughness to the air flow. It increases the roughness length, z_0, and raises the height of the mean aerodynamic surface by a distance, d, known as the zero plane displacement (Fig. 2.7). Estimates of d and z_0 can be obtained from the relationships:

$$D = HF \tag{3.13}$$

$$z_0 = 0.13(H - d) \tag{3.14}$$

where H is the average height of the roughness elements and F is the fraction of the total surface area covered by those elements (Abtew et al. 1989). From this, it follows that the key plant parameter is the lateral cover (L_c), defined as:

$$L_c = \frac{NS}{A} \tag{3.15}$$

where N is the number of roughness elements per unit area (A), and S is the mean frontal silhouette area of the plants, i.e. the cross-sectional area of the plant facing the wind (Musick & Gillette 1990). An increase in the value of L_c results in an exponential decrease in the proportion of the shear velocity of the wind exerted on the soil surface (Wolfe & Nickling 1996). This, in turn, causes an exponential decrease in sediment transport. Al-Awadhi and Willetts (1999) showed from wind tunnel experiments with cylinders that sediment transport levels off to very low levels when L_c exceeds 0.18 in value. When L_c reaches 0.5, sediment transport ceases (Gillette & Stockton 1989; Nickling & McKenna Neuman 1995). However, low densities of vegetation can sometimes increase the rate of erosion over bare ground through the development of turbulent eddies in the flow between individual plants (Logie 1982).

The effect of the vegetation can be described by a frictional drag coefficient (C_d) exerted by the plant layer in bulk and computed from:

$$C_d = \frac{2u_*^2}{u^2} \tag{3.16}$$

where u is the mean velocity measured at a height z, which equals 1.6 times the average height of the roughness elements. The coefficient generally decreases in value from about 0.1 in light winds to 0.01 in strong winds for a wide range of crops (Ripley & Redman 1976; Uchijima 1976; Morgan & Finney 1987) but both Randall (1969) in apple orchards and Bache (1986) with cotton canopies found that C_d could also increase with windspeed. When C_d exceeds 0.0104 in value, no regional scale wind erosion occurs (Lyles et al. 1974b).

Instead of considering these bulk drag coefficients, more insight can be gained by examining conditions close to the ground surface. Drag coefficients within the plant layer (C_d') can be calculated from:

$$C_d' = \frac{2u_*^2}{\int_0^h u^2 A(z) dz} \tag{3.17}$$

where h is the height of the vegetation, $A(z)$ is the leaf area per unit volume for the vegetation at height z and dz is the difference in height between z and the ground surface. For a wide range of crops, values of C_d' within the lowest 0.5 m of the plant layer decrease from about 0.1 in low windspeeds to about 0.001 in high windspeeds. However, when the wind is moderate to strong, consistent over time, and the crops are at an early stage of growth, the drag coefficient is found to increase with windspeed (Morgan & Finney 1987). This is probably due to the waving of the leaves in the wind, which disturbs the surrounding air, creating a wall effect that acts as a barrier to the air flow. The result is that the windspeed is reduced close to the ground surface but remains the same or even increases at the canopy level, thereby increasing the drag or shear velocity and enhancing the risk of erosion. The effect is particularly marked in crops of young sugar beet and onions. Similar increases in the drag coefficient with windspeed within a crop have been reported for maize (Wright & Brown 1967).

3.4.4 Effect on slope stability

It was shown in Chapter 2 that forest covers generally help to protect the land against mass movements partly through the cohesive effect of the tree roots. The fine roots, 1–20 mm in diameter, interact with the soil to form a composite material in which root fibres of relatively high tensile strength reinforce a soil matrix of lower tensile strength. In addition, soil strength is increased by the adhesion of soil particles to the roots. Roots can make significant contributions to the cohesion of a soil, even at low root densities and in materials of low shear strength. Increases in cohesion in forest soils due to roots can range from 1.0 to 17.5 kPa (Greenway 1987), although local variability in this may be as high as 30 per cent (Wu 1995). Grasses, legumes and small shrubs can reinforce a soil down to depths of 0.75–1.0 m and trees can enhance soil strength to depths of 3 m or more. The magnitude of the effect depends upon the angle at which tree roots cross the potential slip plane, being greatest for those at right angles, and whether the strain exerted on the slope is sufficient to mobilize fully the tensile strength of the roots. The effect is limited where roots fail by pull-out because of insufficient bonding with the soil, as can occur in stony materials, or where the soil is forced into compression instead of tension, as can occur at the bottom of a hillslope, and the roots fail by buckling.

Following observations on the forested slopes of the Serra do Mar, east of Santos, Brazil, De Ploey (1981) proposed that trees could sometimes induce landslides through an increase in loading (surcharge) brought about by their weight and an increase in infiltration which allows more water to penetrate the soil, lowering its shear strength. Bishop and Stevens (1964) showed that large trees can increase the normal stress on a slope by up to 5 kPa but that less than half of this contributes to an increase in shear stress and the remainder has the beneficial effect of increasing the frictional resistance of the soil. While, generally, surcharge enhances slope stability, under certain circumstances it can be detrimental. Trees planted only at the top of a slope can reduce

stability, as can trees planted on steep slopes with shallow soils characterized by low angles of internal friction. In the Serra do Mar, the landslides occurred in a soil with an angle of internal friction of less than 20°, on slopes greater than 20°, after two days on which respectively 260 and 420 mm of rain fell.

A vegetation cover should theoretically contribute to slope stability as a result of evapotranspiration producing a drier soil environment so that a higher intensity and longer duration rainfall are required to induce a slope failure compared with an unvegetated slope (Greenway 1987). Further, since soil moisture depletion can affect depths well below those reached by the roots, increases in slope stability should extend some 4–6 m below ground level. In practice, however, as found by Terwilliger (1990) in southern California, soil moisture levels after a few storms reach similar levels in both vegetated and unvegetated soils, so that under the conditions when the risk of mass movement is highest, the drying effect of vegetation is unlikely to play a role. Despite this, overall increases in the factor of safety arising from vegetation are generally in the range of 20–30 per cent (Greenway 1987; Wu 1995).

Box 3

Scale and erosion processes

Erosion processes and the factors that influence them vary according to the scale at which they are studied.

Micro-scale (mm² to 1 m²)
At this scale, erosion is controlled largely by the stability of the soil aggregates. Soil moisture, organic matter content and the activity of soil fauna, particularly earthworms, are major influences. Since aggregate breakdown is largely a result of raindrop impact, the frequency and erosivity of individual rainstorms control the rate of erosion through the rate of soil particle detachment. Soil type, slope and land cover are reasonably uniform over areas of this size so that differences in these can be used to demarcate different micro-scale units. For example, Cammeraat (2002) distinguished between areas of bare crusted soils and areas covered by tussocks of *Stipa tenacissima* when describing erosion processes at this scale in the Guadalentin Basin, Spain. In woodland in the Luxembourg Ardennes, he distinguished between bare soil areas where earthworms remove and digest the freshly fallen litter, and litter-covered areas with lower biotic activity.

Plot-scale (1 m² to 100 m²)
Erosion at the plot scale is controlled by the processes that generate surface runoff. These include the infiltration characteristics of the soil and changes in the microtopography of the surface related to crust development and surface roughness. The spatial distribution of crusted and uncrusted areas or vegetated and bare soil areas determines the locations of runoff and the patterns of flow and sediment movement over the soil surface. Depending on the size of the plot, interrill erosion will dominate, but, on steep slopes or in areas with highly erodible soils, the flow may exceed the critical conditions for rills to develop. It is at this scale that rock fragments can increase rates of soil erosion compared with bare ground if the surface between the stones becomes sealed (Poesen et al. 1994).

Field scale (100 m² to 10,000 m²)
At the field scale, there is usually a reasonably well defined spatial pattern of runoff pathways in locations such as swales and valley bottoms, separated by areas of either interrill erosion or no erosion. The extent of interrill erosion depends on the severity of the rainfall event so that the size of the area contributing

Continued

runoff is quite dynamic. The direction of runoff pathways is often controlled by tillage. There may be spatial variations in soil erodibility and slope within a field. The normally expected increases in runoff and erosion downslope can sometimes be offset by soils with greater infiltration rates at the bottom of the field, resulting in a decrease in runoff, or by gentler slopes, resulting in deposition of sediment.

Catchment scale (>10,000 m²)

Depending on the spatial distribution of soils, slope and land cover in a catchment, it is possible to recognize different types of process-domains (De Ploey 1989b):

I: areas dominated by rainsplash erosion and infiltration, typical of soils with high infiltration rates and high aggregate stability.

II: areas dominated by interrill erosion but with sediment deposition on the lower slopes.

III: areas dominated by rill erosion, again with deposition on the lower slopes.

IV: areas of ephemeral gullies, particularly along the valley floors.

The pathways of runoff and sediment movement through the catchment are influenced by the patterns of field boundaries, gateways, tracks and roads, as well as the natural topography. The effectiveness of these pathways and the spatial extent of the process-domains depend on the magnitude of the erosion event. In low magnitude events, erosion is generally limited to local slope wash, but with higher rainfall, runoff pathways develop over the whole hillside with local discharges into the river; with more extreme events, overland flow and slope wash may be widespread, with many points of discharge into the river channel (Święchowicz 2002). Thus, in order to understand the sources of sediment and associated pollutants in water bodies, the dynamic nature of the process-domains and their connectivity must be investigated.

The way in which the effectiveness of erosion events depends on the scale of analysis was demonstrated by Carver and Schreier (1995) for the Jhikhu Khola watershed, Nepal (Table B3.1). They examined three storms, one of 49.5 mm occurring in the pre-monsoon period, one of 35.8 mm during the monsoon and an extreme event of 90.6 mm in the transition period just before the onset of the monsoon season. On a 70 m² area of a terraced hillslope, erosion in the pre-monsoon storm when the cultivated land was bare was higher than that from the monsoon storm when the land was protected by a crop cover; the extreme event was only half as damaging as the pre-monsoon event. In a sub-watershed of 540 ha, the pre-monsoon and monsoon storms had less overall effect but the extreme event became very important as a result of erosion of the stream bed. At the scale of the whole watershed, 11,141 ha, the differences between the pre-monsoon and monsoon periods disappeared and the effect of the extreme event was much reduced because the heavy rain did not extend beyond the area of the sub-watershed. Thus, what was an extreme event at a small watershed scale was no larger than the annual event at the larger watershed scale.

Table B3.1 Effectiveness of erosion events ($t\,ha^{-1}$) with changes in scale

Season	Rainfall total (mm)	Peak rainfall intensity ($mm\,h^{-1}$)	Terraced hillside (70 m²)	Sub-watershed (540 ha)	Jhikhu Khola watershed (11,141 ha)
Pre-monsoon	49.5	109	20	2	0.1
Monsoon	35.8	63	0.02	0.4	0.1
Extreme event	90.6	103	10	40	2

Source: after Carver and Schreier (1995).