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Soil Erosion

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FOREWORD

This publication is derived from a 10 week course on soil erosion given within the graduate Diploma in Irrigation and Soil Conservation of the Faculty of Agriculture, University of Nairobi. Interest in soil erosion and conservation in Kenya, and Africa, appears to be increasing. A Workshop on Soil and Water Conservation was recently held in Nairobi, the proceedings of a Conference on Soil Conservation and Management in the Humid Tropics have just been published and Wenner has produced a handout for technical assistants on Soil Conservation in Kenya.

Although several good texts on soil erosion and conservation exist, many are either out-of-date or are mainly concerned with North American situations. This paper aims to complement the above publications by giving a broad view of soil erosion, the erosive processes and the environmental and agronomic factors which affect the processes. It draws upon material from a variety of disciplines, attempts to compile the results of recent studies and pays particular attention to examples applicable to East Africa. I hope that it will be of use to students, researchers and other interested parties.

I would like to thank the Department of Soil Science for making available publication facilities.

TABLE OF CONTENTS

1. INTRODUCTION

Soil erosion is a topic which has gained popularity and prominence in the last three or four decades. Kovda (1977) has recently reviewed soil erosion and man's influence on it, and has called for an increase in research and application, such as through the creation of an International Conservation Decade. It must be realised, however, that soil erosion is a naturally-occurring process, and that much of the contemporary concern arises from the "accelerated" rates of soil erosion under modern land use systems, compared to lower "normal" or "geological" rates. The term accelerated is not easily defined, and some soil erosion is tolerable and probably necessary. The definition of the tolerable limits of erosion will be examined later.

Accelerated rates of soil erosion first became apparent in old civilizations. The decline of civilizations, such as the Greeks and the Romans, has occasionally been ascribed, in part, to poor land use management and the removal of the forest (Stallings, 1957). However, Vita-Finzi (1974) has cautioned against almost automatic identification of accelerated rates of erosion with man's mismanagement of the land, and suggests that changes in climate, and particularly rainfall amount and distribution, are important contributory factors. Moreover, ancient civilizations such as those around the Tigris, Euphrates and Nile depended on rich alluvial materials. The degradative effect of man's impact on the landscape depends to a large extent on the ability of the environment to resist changes in the erosive forces (its "buffering capacity"). In any assessment of the impact of soil erosion on man's activities, the benefits and costs must be carefully analysed for the whole of society (Muthoo, 1976). Nevertheless, extensive soil erosion has caused a significant decrease in the productivity of the soil, through chemical, physical, and biological changes in the soil body. The best documented examples are from the United States, where the Soil Conservation Service was established after an appreciation of the importance of accelerated rates of soil erosion and land degradation associated with the drought of the early 1930's. In 1934, it was estimated that, from a total of 167 million ha of arable land in the United States, 20 million ha were totally ruined, 20 million ha were almost ruined, 40 million ha had lost over half of their topsoil and another 40 million ha had lost more than a quarter of their topsoil (Hudson, 1971). Thus, nearly three-quarters of the arable land was seriously damaged by soil erosion.

The U.S. Soil Conservation Service was established to combat this problem and recent reviews of its work have been published by Ackermann (1976) and Held and Clawson (1965).

From the conservation viewpoint, the soil may be regarded as an exhaustible but renewable resource under normal conditions (Held and Clawson, 1965). The soil can be exhausted by the extraction and removal of plant nutrients, and by the erosion of the soil particles. The nutrients can be replaced by the application of fertilizers or organic matter and the soil particles can be replaced by the weathering of bedrock. The problem of soil erosion can be viewed as the conflict between short- and long-term accounting periods, characteristically used in many "western" agricultural activities. Short-term accounting seeks to maximize profits in the near future, whereas long-term accounting aims to maintain productivity and profits at a stable level over a period of time. The former often causes a pronounced degradation of the soil resource. Soil conservation, then, often involves the sacrifice of short-term benefits for the long-term maintenance of productivity (Held and Clawson, 1965; Warren, 1974).

Quantification of the economic aspects of soil erosion and conservation is often complicated and difficult. Studies in the U.S., however, suggest that even in the 1950's the annual cost of soil erosion was well over 100 billion dollars. This represents the combined effects of reduced crop yields and increased fertilizer requirements, downstream damage by floods, damage to water supplies and fisheries by increased sediment load, siltation of harbours and reservoirs, and increases in malarial areas (Stallings, 1957).

What are observed values of soil erosion rates? The problem facing this question is that soil erosion is often quoted in a variety of different units, such as cm^3/cm^2, tons/ha, m^3/ha and mm, and both Imperial and metric units.

On a global scale, Judson (1968) estimates that the world soil erosion rate has increased through man's activities from 9 to 24 billion tons per year, based on measurements of suspended sediments, and this observation is supported by the measurements of Gregor (1970). Using a variety of data sources, Young (1969) has calculated a median erosion rate of 46 mm/1,000 yr for normal relief and 500 mm/1,000 yr for steep relief, with interquartile ranges of 20-81 and 92-970 mm/1,000 yr, respectively.

One of the major factors affecting the erosion rate is climate, and through it, vegetation. Douglas (1967) and Langbein and Schumm (1958) have shown that there is a relationship between soil erosion rate and climate, which is best expressed as the amount of runoff per year. This relationship is shown on the following diagram, (Fig. 1.1), and reveals that the most erosive areas are those with an annual runoff of about 20 mm, equivalent to an annual rainfall of 200 to 500 mm. The discrepancy between the two curves reflects the fact that the Langbein and Schumm curve was based on drainage basins in the U.S., which have been affected by agricultural activities, whereas the Douglas basins were essentially undisturbed by man. The discrepancy suggests, again, that man has caused an increase in soil erosion rates by at least twofold.

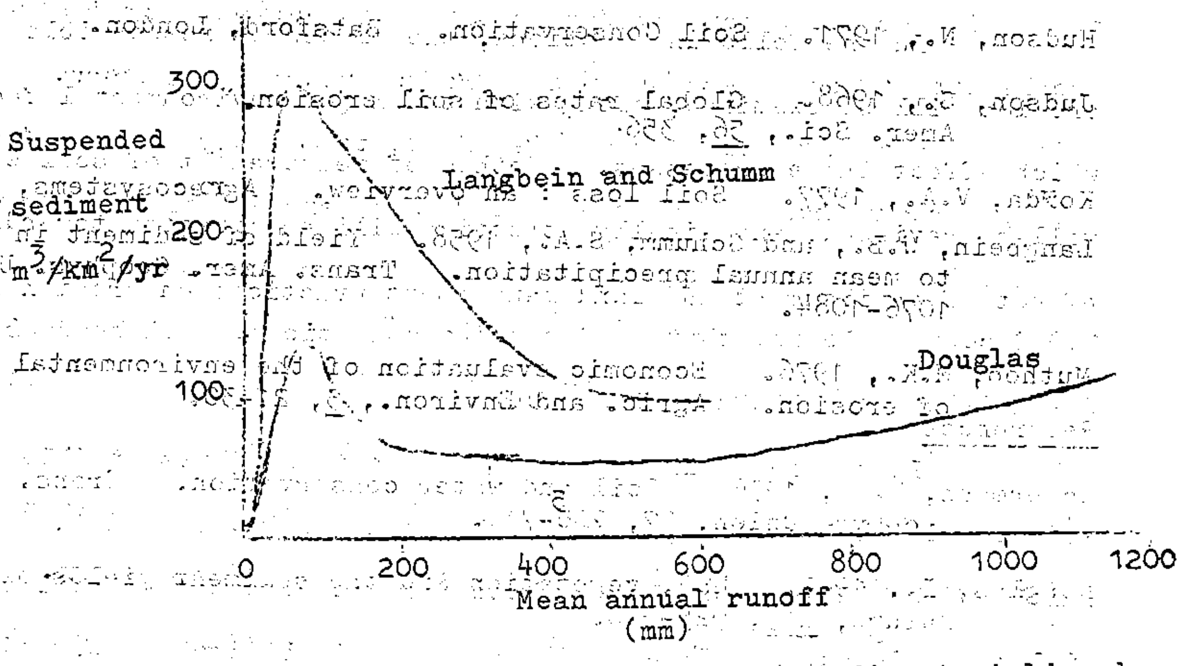

Fig. 1.1 The relationship between sediment yield and mean annual runoff.

In Kenya, Dunne et al. (1977) have calculated historical erosion rates from geomorphological relationships. In the Cretaceous and Tertiary periods, the climate was wetter than at present and much of Kenya was occupied by forests, with woodland/savanna in the drier areas and at drier times. Overall erosion rates appear to be 0.0007 to 0.0012 cm/yr, equivalent to an average of 24 tons/km^2/yr. This is considerably lower than the values quoted by Young (1969). The Quaternary period was accompanied by more pronounced oscillations of climate and vegetation, with, in particular, cooler and drier periods associated with the glaciations in temperate and polar regions. The overall erosions rate in the Quaternary appears to be about 0.0029 cm/yr (equivalent to 77 tons/km^2/yr), but the drier periods were probably accompanied by erosion rates of 0.0075 cm/yr (equivalent to 200 tons/km^2/yr).

At the present time, studies of Kenyan catchments indicate erosion rates of 18 to 26 tons/km^2/yr from undisturbed, forested land,

and 50 to 140 tons/km^2/yr from lightly grazed, semi-arid lands. Where heavy grazing occurs, and in catchments which have undergone intensive agriculture without adequate soil conservation practices, erosion rates are well over 1,000 tons/km^2/yr or 0.1 to 1.0 cm/yr. It is these accelerated rates of erosion which are the cause for concern. With soils 1m thick the soil has a life of 100 to 1,000 years, and the most important fraction of the soil will be eroded first, leaving an essentially unproductive residue. Rates of soil formation may be only 0.002 cm/yr, showing that, under these conditions, soil becomes an essentially non-renewable resource.

In the following sections, the processes responsible for soil erosion will be examined, along with the major environmental factors which affect these processes. Method of measurement of soil erosion will be reviewed, and finally attention will be directed towards the effect of environment and land management practices on erosion rates.

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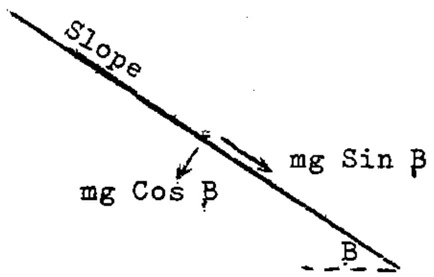
2. THE BASIC PRINCIPLES

An understanding of the processes of soil erosion demands an appreciation of the forces which act to move soil particles, and the resistances that the soil body presents to these forces.

Forces

Force requires energy, and all the energy will be ultimately derived from either gravity or climate.

Gravity forces are expressed in terms of the weight of an object, and this force acts vertically downwards. Most soil particles rest on a slope, and in this case the gravity force can be resolved into two components, a downslope component which tends to move the soil particle downslope, and a perpendicular component which tends to hold the particle on the slope. These can be represented, (Fig. 2.1).



downslope component: mg Sin β
perpendicular component: mg Cos β

Fig. 2.1 Gravity forces acting on a slope.

As the slope angle β increases, the downslope component increases and the perpendicular component decreases, so that the force tending to move the particle effectively increases.

Water is a major force moving soil particles. When water moves in response to gravity, potential energy is converted into kinetic energy. Much of this kinetic energy is dissipated as heat within the body of water and along the sides of the water channel, but some of the energy may be used to transport soil particles. The speed of the moving body of water is important and represents the balance between the downslope component of the weight of water and the frictional forces acting on the channel bed and banks. The flow of water exerts forces on the particles at the base of the channel, and when these are larger than the weight of the particles and frictional and cohesive forces holding the particle on the channel bed, the flowing body of water can transport the particle downslope.

In addition to water movement along the soil surface or in a stream channel, water movement can also occur through the soil body. Rates of movement, however, are generally slow so that little overall transport of soil particle is effected, though small clay particles can be eluviated from the upper to lower parts of the soil profile.

Water is also important in that rainfall possesses kinetic energy, and this energy is transferred from the water drop to the soil surface, though some is preserved in water movement after raindrop impact. The impact force can be absorbed by the soil through compaction or crater formation, but some is also absorbed by upward movement of soil and water particles. Where the slope is appreciable, net downslope transport of soil particles will occur.

The water in the soil also exerts a force on the soil body. Beneath the water table surface, gravity exerts a pressure which is equal to the overlying weight of water, thereby subjecting the soil body to an upward thrust equal to the weight of water which is displaced. Thus, the effective gravity force, or weight of the soil, is reduced, and this can cause instability in the soil body.

Above the water table, water also occupies the soil pore space, but not all the pores are filled with water. This water exerts a suction, often called the soil moisture tension, which increases as the water content decreases. This tension force, called the capillary cohesion, increases the effective weight of the soil body.

Expansion and contraction of the soil body can cause the downslope movement of soil particles, and this expansion/contraction is related to changes in temperature and especially water content, the latter acting through the capillary cohesion forces.

Although tremendous forces are available for the transport of soil particles, the efficiency of these forces in transporting material is very low. The following table gives an indication of the magnitude and efficiency of the transport forces (Table 2.1).

Table 2.1 Orders of magnitude of geomorphic forces and their efficiency in transporting debris (from Carson and Kirkby, 1972).

Force	Total Work Done By Force $J/m^2/yr$	Total Work Done In Downward Transport of Particles $J/m^2/yr$	Efficiency %
Gravity	1-100	1-100	100
Water flow rivers	10^5-10^6	10-100	0.01
Rainfall impact: grass	2000	0.02	0.002
bare	500	0.5	0.1
Heaving moisture	2000	0.4	0.02
frost	5×10^7	0.1	2.5×10^{-7}
temperature	2×10^8	0.02	1.5×10^{-9}

Resistances

The earth's surface would be a plain were it not for the strength, or resistance, of the soil body against the potentially-erosive forces examined in the preceding section.

There are several properties of soils which contribute to this strength:

(a) Particle size is important in soils. The conventional size limits for the major particle size classes are:

Boulders	> 256 mm
Cobbles	64 - 256 mm
Gravel	2 - 64 mm
Sand	0.05 - 2 mm
Silt	0.002 - 0.05 mm
Clay	< 0.002 mm

(b) Pore space is the volume of the soil that is not occupied by solid particles, i.e. it is occupied by water or air. The total porosity of soils can vary considerably depending on cultivation practices, organic matter content and so on, but common values are:

Gravel	25 - 40%
Sand	30 - 40%
Silt	20 - 50%
Clay	45 - 60%

By comparison, most rocks have very low porosities, such as 10^{-4} to 1% for most igneous and metamorphic rocks, 5 to 20% for many sedimentary rocks (e.g. shales, sandstones, limestones), but up to 80% for some volcanic rocks, for example tuff (Gregory and Walling, 1973).

Equally important is the pore size distribution. An adequate number of relatively large pores are required for water drainage, aeration and root development, whilst the small pores are able to retain water against drainage, and, hence, act as a reservoir of soil water.

(c) Shear strength is important in that this is the direct resistance against the erosion forces. The shear strength of a soil or rock is derived from a number of components. One is the plane friction when one soil grain attempts to slide past another; a second is the interlocking of the soil grains. These are the internal friction of the soil or rock, which is usually expressed by the parameter ϕ, the angle of internal friction, or tan ϕ, the coefficient of internal friction. The total force developed by this frictional strength is the product of the coefficient of internal friction and the normal stress attempting to push the soil or rock together, i.e. compressing. In addition to the above frictional components, cohesion aids to pull particles together.

Thus, the shear strength can be represented:

$s = c + \sigma \tan \phi$ where s = shear strength (kgf/cm^2)

c = cohesion

σ = normal stress

ϕ = angle of internal friction

The internal angle of friction is dependent on a number of properties of the soil. For sandy materials, it decreases as particle size increases, and the porosity increases. The shear strength increases as the applied normal stress increases.

Cohesion becomes important when the clay-size fraction forms an important component of the soil or rock body. Cohesion is the bonding of soil particles due to attractive forces between them. These forces consist of:

(a) electrostatic attraction between positively and negatively charged sites on particle surfaces, particularly clay minerals;

(b) van der Waal's forces, those which are a function of the gravitational pull between particles;

(c) cationic bridges between particles, linking adjacent negatively charged sites;

(d) cementation of particles by organic matter, iron and aluminium oxides, hydroxides, and carbonates;

(e) surface tension of water at water contents below saturation; (Baver et al., 1972).

Thus, the more reactive and the finer the particles, the greater the cohesion between them. Also, the higher the water content, the lower the cohesion in a soil body.

There are a variety of methods for measuring shear strength.

Some commonly-observed values for soil and rock materials are quoted in the following table (Table 2.2 from Carson and Kirkby, 1972).

Table 2.2 Commonly-observed shear strengths of various soil-forming materials.

Material	Shear Strength (kgf/cm^2) c	ϕ^o
Chalk	9	21
Sandstone	350	44
Limestone	45-350	37-58
Granite	97-406	51-58
Talus (granite)	0-8	18-36
Sandy soil	0	33-43
Sandy granitic soil	0	36-37
Clay loam	0.1	25
Clay loam	0.4-0.5	29-30
Clay	0.15	20
Quartz (100% clay)	-	35
Quartz (0% clay)	-	35
Muscovite (100% clay)	-	24.5
Muscovite (0% clay)	-	20.0
Ca-montmorillonite (100% clay)	-	10.0
Na-montmorillonite (100% clay)	-	4-10

Thus, it can be seen that there is a considerable range in the shear strength of materials. Cohesion is high in intact rocks, such as sandstone and granite, decreases to very low values in coarse weathered material and then increases again in fine clays. The more reactive the clay, the higher the cohesion, but also the cohesion depends on the type of absorbed cation. The angle of internal friction generally decreases as the size of the material decreases.

(d) The other important property of soil which lends resistance to change is its <u>plasticity</u>, which allows a soil to change shape without cracking or rupture - a deformation in response to an applied stress. In this characteristic, the cohesion is an important control on the plasticity characteristics of the soil, and also the water content of the soil.

There are two points which are commonly measured as indicative of the plasticity characteristics of a soil, and these are known as the Atterberg limits.

The upper plastic limit (or liquid limit) is the moisture content at which the soil will barely flow under an applied force. The lower plastic limit is the moisture content at which the soil can barely be rolled into a wire. The plasticity number (or index) is the differenc between the two moisture contents. The method of measurement of these two parameters is discussed by Baver et al. (1972).

The change in the soil strength between these limits represents the influence of differences in moisture content on the cohesion forces in the soil. From a dry soil, as water is added, the cohesion increases until the lower plastic limit is reached, when the soil can be moulded into a wire. This is equivalent to a pF value of 2.8 to 3.3. As further water is added the cohesion decreases until the liquid limit is reached (equivalent to a pF value of 0.5), when the cohesion is reduced such that the soil can flow under an applied stress. The plasticity index is the amount of water which must be added to the soil to change the maximum cohesion to flow, and thereby is an indirect measure of the force required to mold the soil. The plasticity characteristics are important, then, in resistance to erosion and farm practices, such as ploughing.

Several soil properties affect the plasticity characteristics; these are mainly clay content and mineralogy, adsorbed cation type and organic matter content.

As plasticity is a function of the finest fractions of the soil, the plastic limit, liquid limit and plasticity index all increase with an increase in the clay content. The plastic limit increases only slightly, but there are pronounced increases in liquid limit, and, hence plasticity index.

The type of mineral affects the plasticity measurements. Quartz and feldspar clay-size minerals are non-plastic. Within the clay minerals, the plasticity generally increases in the series kaolinite illite montmorillonite, representing increasing surface area and the interlayer swelling of montmorillonite.

The type of exchangeable cation on the clay minerals is important, through the cations' influence on the amount of water adsorbed on the clay surface. In general, Na-saturated soils have generally low plastic and liquid limits, but often have high plasticity indices. K also introduces low plastic and liquid limits, but low plasticity indices. On the other hand, the divalent cations, Ca and Mg, increase the plastic and liquid limits. The specific effect of different cations depends to a large extent on the type of clay mineral present in the soil, and its interaction with the cation.

The fourth important characteristic, especially from an agronomic viewpoint, is organic matter content. An increase in organic matter content generally causes an increase in the plastic and liquid limits, though there is little change in the plasticity index. These results can be explained by the high absorptive capacity for water possessed by the organic matter; this absorptive capacity occurs up to the plastic limit and thereafter little further water is absorbed by the organic matter, resulting in little change in the plasticity index.

These plasticity characteristics are important, then, in that they control the response of the soil to changes in water content, and the ability of the soil to resist deformation by applied stresses.

Associated with these measures of soil strength is the structure of the soil, which is the three-dimensional arrangement of soil particles. Soil structure is important in that it, to a large extent, controls the infiltration rate of the soil, and offers resistance to movement of soil particles, by binding together the particles into larger, less mobile aggregates. Soil structure is caused by swelling and shrinking of the soil associated with wetting and drying cycles, and the pressures exerted by plant roots and soil organisms. The stability of the structure is important, and this depends on the type of cation adsorbed onto the exchange complex, the cementing effect of iron and aluminium oxides and hydroxides, cohesion between clay particles and the binding effect of organic matter and products of microbial activities. A more detailed discussion of soil structure is given in Baver et al. (1972), and will be referred to again in the consideration of soil erodibility.

(e) Last, but perhaps not least, the vegetation cover plays an important role in increasing the resistance of the soil body to erosive forces. One component of this role is through vegetation and its associated surface litter reducing the energy of raindrops, and, hence, rainsplash of soil. In addition, plant roots help strengthen and bind the soil together through their decomposition products and exudates. The importance of vegetation in reducing soil erosion will become obvious later.

It appears, then, that there are several forces acting in the environment which tend to move soil materials downslope, and that these forces are opposed by strengths possessed by the soil body. At equilibrium, the forces promoting movement are balanced by the forces opposing it; when the movement forces are increased, then the opposition forces will be overcome and soil will move downslope.

Although the above discussion has involved a number of measures of movement forces and soil strength, it is difficult to apply these measurements to specific cases of soil erosion. As will be seen later, most studies of soil erosion still involve empirical relationships between the driving forces and the soil resistances. This does not detract, however, from the principles involved. One can expect an increasing number of studies involving the fundamental forces and resistances.

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3. TYPES OF SOIL EROSION

From a geomorphological viewpoint, two major types of processes can be recognized that involve the movement of soil material. There are, firstly, mass movements, which involve soil movement under gravity, without significant assistance from other outside forces, such as flowing water and wind. Secondly, there are surface processes, which are dominated by the action of water and wind. These two sets of processes are obviously of central importance in geomorphology in that they control, to a large extent, the shape and distribution of slopes.

(A) Mass movements

This category involves a range of processes, including rockfalls, landslides, earthflows, and soil creep.

Rockfalls are spectacular and occur when the transporting agent (gravity) is greater than the ability of the rockface to supply weathered material, resulting in a coarse scree or talus at the bottom of the rockface. As the phenomenon is of little importance to soil erosion and conservation in agricultural areas, it will not be discussed further, though a detailed discussion can be found in chapter six of Carson and Kirkby (1972).

A more serious problem from the soil conservation viewpoint is instability in the soil mass, such as exhibited by landslides and earthflows. Landslides can be either deep-seated or shallow, depending on the depth of the soil mantle, the relationship between shear strength and stress with depth, and the resulting depth of the failure surface. A major property promoting landslide development is high pore water pressure resulting from a high water table and downshope movement of soil water. Landslides can also occur when the regolith is removed from beneath a laterite layer, resulting in a slumping of the surface (Thomas, 1974).

Landslides appear to be restricted to slopes of 35 to 60° in the tropics, although they can occur on slopes of 20° or less which have a high clay content. The distribution of landslides follows steep slopes, often related to stream incision and land dissection, and high pore water pressures associated with rains. Earthquake activity is often the triggering mechanism, and the removal of forest vegetation can promote landslide development, through loss of surface strength and precipitation interception (Thomas, 1974).

Earthflows can also affect agricultural land, the movement occurring as a viscous flow. They often occur in clay-rich sediments which have a high sensitivity (i.e. high undisturbed:disturbed strength ratio), a natural water content greater than the liquid limit, and again are related to heavy rainfall and tremors (Carson and Kirkby, 1972).

Two interesting studies on landslide activities and soil conservation in tropical areas have recently been published.

The first examines erosion in the stiff, overconsolidated Joe's River Muds of the Scotland District in eastern Barbados (Carson and Tam, 1977). Landslides and gullying are extensive and produce a "badland" landscape, which has increased in area by about 45% between 1951 and 1968. The increase in these landforms appears to reflect the change in land use from forest and grassland into cropland (particularly sugar cane and banana) since the area was colonized. Over half the area shows signs of erosion and instability, with deep ravines and landslides and slumps on slopes of 20° or more. Many slopes are steeper than the frictional angle of the regolith, and mass movements are likely to occur in the future.

Increases in overland flow and stream discharge have caused the gully development, and this has been aided by the relative fall in sea level. The landslides are triggered by high pore water pressures, associated with intense rains and the movement of water into tension cracks caused by drying and shrinkage of the muds. The soil important in the formation of soils (e.g. Nye, 1954, 1955), but which is unimportant as a method of accelerated soil erosion.

(B) Surface water-induced movements

Within this category, several processes can be recognized, though, once again, they are often intimately related in nature. These categories are rainsplash erosion, sheetwash or overland flow erosion, and channel erosion, such as the formation of rills and gullies.

(i) Rainsplash erosion

Raindrops falling at their terminal velocity of close to 8 m/sec possess energy, which must be dissipated upon their impact with the soil. Water moving through the soil or along its surface possesses a relatively small amount of energy, so that much of the raindrop's energy must be absorbed by the soil. The magnitude of this energy change can be shown (Hudson, 1971) :

$$\text{Kinetic energy of rain} = \frac{1}{2} R\,(8)^2 = 32\,R$$

$$\text{Kinetic energy of water moving through the soil and as overland flow} = \frac{1}{2} R\,\frac{3}{4}\,(0.0001)^2 + \frac{1}{2}\,\frac{R}{4}\,(1)^2$$

$$= \frac{R}{8}$$

where R is the mass of the water; $\frac{3}{4}$ of the rain passes into the soil, with a velocity of 0.0001 m/sec; and $\frac{1}{4}$ runs off as overland flow with a velocity of 1 m/sec.

The pronounced decrease in the energy of water can be absorbed by a variety of mechanisms. The soil can become compacted, craters can form around the raindrop impact, water droplets can be splashed upwards and soil particles can also be splashed outwards. These latter two processes are well illustrated by photography in most textbooks.

If the soil surface has zero slope, then the soil particles will be splashed in a random fashion, and there will be no net movement of soil particles in any one direction. Where a slope exists, then there will be a net movement of particles downslope, caused by the downslope component of the raindrop energy, and the longer downslope trajectories of splashed soil particles.

The amount of splash erosion will depend on a number of factors. Many laboratory or field studies have shown that there is a good correlation between energy parameters of the rainfall (Particularly momentum and kinetic energy) and the amount of splash (Bisal, 1960; Bubenzer and Jones, 1971; Ekern, 1951; Ekern and Muckenhirn, 1947; Ellison, 1944; Free, 1960; Hudson, 1971). In particular, Rose (1960) studied five East African soils, and found that momentum was a better predictor of splash erosion than kinetic energy, though, as pointed out by Hudson (1971), they are very similar for natural rainstorms.

The soil characteristics also affect the splash erosion process. Mazurak and Mosher (1968, 1970) studied the influences of different rainfall intensities on the movement of soil particles and aggregates through rainsplash. They found that particles with diameters of 400 to 50 um were most readily detached; the larger sizes were too heavy and the smaller sizes were not moved so much because of cohesive forces increasing their effective weight. Aggregates treated with Krilium to increase their stability behaved in a similar manner to the individual particles, over the range of sizes examined, with a maximum detachment at sizes in the range 297 to 210µm. Untreated aggregates, however, showed a maximum splash at diameters of 2400 to 1700 µm, suggesting that they were broken down by the raindrop impact. At these large sizes, untreated aggregates splash erosion was about four times that of the treated aggregates. Total splash and the size range of splash erosion increased with an increase in rainfall intensity. Farmer (1973) obtained maximum detachments with soil particles in the diameter range 800 to 300 µm, but found that increases in rainfall intensity did not cause much increase in the detachability of the particles. Working with clods, Moldenhauer and Koswara (1968) found that large clods of a silty clay loam reduced the amount of splash erosion, mainly through a delay until the clods are broken down into detachable sizes. Further studies into rainsplash are reported by Young and Wiersma (1973).

The water content of the surface also affects splash erosion. Dry soils often have a friable, very easily removed surface layer, but then detachment decreases as the surface becomes wetter and more compact layers are produced or exposed. The presence of a thin film of water on the soil surface increases the splash effect, perhaps through a decrease in cohesive forces between particles, but then the water layer decreases splash when the water layer is thick enough to absorb much of the raindrops' energy (Palmer, 1964).

An important by-product of the splash process is the formation of a thin crust or cap of compacted particles on the soil surface. This cap may be only 1 to 3 mm thick, but is important in reducing infiltration of water (and hence increasing overland flow), though it may also reduce the erodibility of the soil (McIntyre, 1958 a, b). The crust or cap is less likely to be important where the slope is steep or where there is a rough microtopography.

Splash erosion movements can be up to 150 cm (Ellison, 1944), and stones 4 mm in diameter can move up to 20 cm. The overall effect is to remove the finer particles downslope, leaving a residual layer of gravel and stones. Carson and Kirkby (1972) show that the downslope movement is linearly related to the sine of the slope angle. On a slope of 10%, Ellison (1944) found that the downslope component of rainsplash erosion was three times that of the upslope component.

Vegetation cover is important in that it reduces the impact of the raindrops, though this may not be effective in all cases. For example, tropical rain forests may intercept a large proportion of the rainfall, but drops falling from the canopy as throughfall may be large, reach terminal velocity and fall onto a soil surface with a thin litter layer and sparse understory, causing considerable splash erosion (Thomas, 1974).

(ii) Overland flow

When the intensity of rainfall exceeds the infiltration capacity of the soil, water will accumulate on the soil surface, and if there is a sufficient slope, water will move downslope across the soil surface. This is known as Hortonian overland flow, after the man who first examined it in detail. Overland flow can also occur in lowlying areas where the soil becomes saturated through subsurface seepage.

Thus, the rainfall intensity and infiltration capacity of the soil are critical in determining the spatial and temporal occurrence of overland flow. The infiltration capacity of the soil is dependent on a number of factors:

(a) the grain size distribution and the pore size distribution. Movement of water into and through the soil is generally restricted to the non-capillary pore spaces, and thus the infiltration rate depends on the non-capillary porosity. For texture classes, the infiltration capacity decreases in the order sand > silt > clay. Also well structured soils will have higher infiltration capacities than poorly structured ones;

(b) the infiltration rate decreases with time as more of the soil pores become occupied by water and the rate decreases to a constant value, which is the ability of the soil to transmit water downwards - the saturated hydraulic conductivity;

(c) since the water content of the soil increases through infiltration, the wetter the soil prior to the storm, the smaller the amount of water that can be absorbed by the soil;

(d) the storm intensity may also affect the infiltration capacity, especially on bare or poorly covered soils. As noted above, raindrop impact can break down the soil structure and produce a relatively impermeable cap or crust, effectively reducing the infiltration capacity.

The variation of infiltration through a storm shows the following pattern (Fig. 3.1)

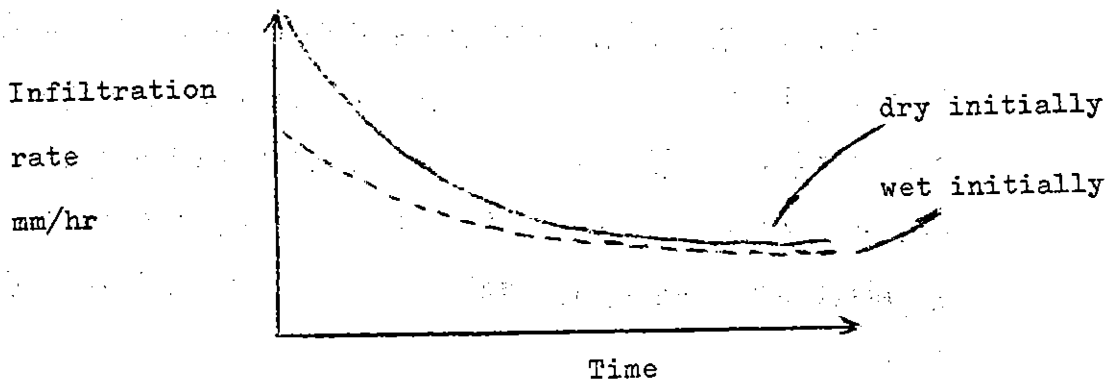


Fig. 3.1 Variation in infiltration rate through time.

Average infiltration capacities are (Kohnke and Bertrand, 1959):

Clay loam	2.5 - 5.0 mm/hr
Silt loam	7.5 - 15.0 mm/hr
Loam	12.5 - 25.0 mm/hr
Loamy sand	25.0 - 50.0 mm/hr
Sand/gravel	50.0 - 100.0 mm/hr

The infiltration rate of a soil can be increased by disturbing the surface layer, by adding organic matter, and by encouraging the development of strong root systems.

Another major component which affects the amount of infiltration and overland flow that will occur is the interception of rainfall by vegetation, whereby the intercepted water is evaporated back into the atmosphere, without reaching the soil surface. The interception rate depends on numerous factors such as the size, shape, and character of the vegetation, the intensity and duration of the rainstorm and the evaporative power of the atmosphere. Interception can be quantitively important in the hydrologic cycle, as can be seen from the following average losses through interception of rainfall (Gregory and Walling, 1973):

Forests	10 - 30%
Grass	30 - 60%
Corn	15%
Oats	7%
Clover	40%

The data are derived mainly from temperate areas, and the interception rates are probably lower in tropical areas where rainfall is generally more intense.

A final hydrologic component which reduces the occurrence and effectiveness of overland flow is depression storage. This involves the collection of water in small depressions, the water slowly infiltrating into the soil or being evaporated into the atmosphere. Depression storage may account for 2 to 5 mm of a rainstorm, depending on surface roughness and configuration.

Therefore, the water not intercepted, not infiltrated and not caught by depression storage will flow across the surface as overland flow, where the slope allows. When the flow occurs as a fairly thin and uniform layer across the soil surface, it is known as sheetwash. When the flow becomes more concentrated, however, the flow occurs in rills, gullies and streams.

All of these water movements are controlled by the interaction of the downslope component of the water weight and the frictional drag exerted by the soil surface or the channel walls. The velocity of the flow can be calculated (Carson and Kirkby, 1972):

$$v^2 = \frac{2g}{f} \cdot r \cdot s$$

where v = velocity
g = gravity
r = hydraulic radius (mean depth of flow)
s = slope
f = friction factor

Therefore, an increase in depth of water or slope angle increases the velocity of flow, whilst an increase in the frictional factor (dependent mainly on the size and configuration of surface particles) causes a decrease in velocity.

The type of flow is also important, in terms of the ability of the water body to detach and transport soil particles. Where the viscous forces dominate, flow is laminar; where the inertial forces dominate, then flow will be turbulent. The ratio between viscous and inertial forces is defined by the Reynolds Number - laminar < 500, turbulent > 2000.

In thin flows on steep, smooth slopes, laminar flow generally dominates. As the thickness of the water increases or the surface roughness increases, then the flow will become more turbulent. Where the surface is very rough, such as where a field has grass or stones, then, although the flow is very slow, the roughness diverts the flow such that a sinuous turbulent flow is produced.

Flowing water has the ability to detach and transport particles. Several forces operate on the bed to raise and move the particle, and once entrained, the particle can be transported in flow by turbulence within the water body. When the flow rate or turbulence decreases, some of the soil particles will be deposited.

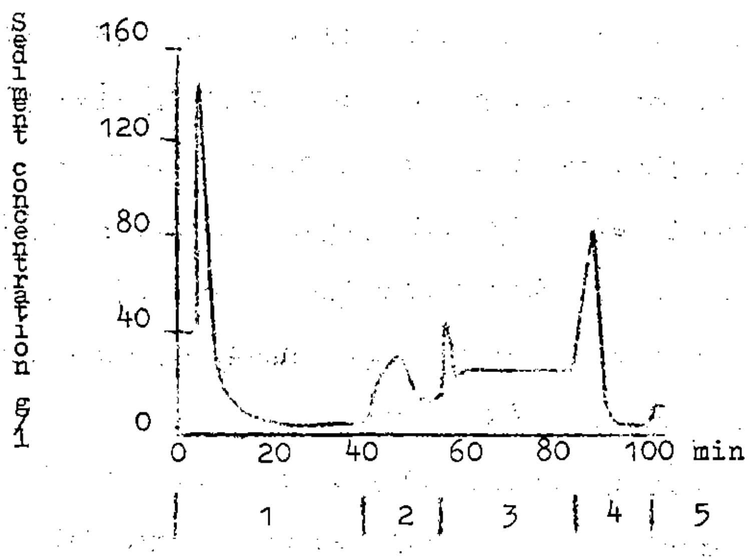
In most cases, rates of overland flow are very small - often 1 - 100 cm/sec, and generally $<$ 10 cm/sec (Ellison, 1947a), so that the tractive force exerted by the water on the soil surface is low, and few particles can be detached from the surface. The flow can, however, readily transport fine particles, when they have been detached from the soil surface.

Yoon and Wenzel (1971) have shown that rainfall causes a considerable increase in the turbulence of overland flow, particularly when the Reynolds Number is low (about 500), so that flow is laminar, as is common with overland flow (Emmett, 1970; Pearce, 1976). Under more turbulent conditions (Reynolds Numbers > 1500), increased turbulence occurs only in the upper layers of the flowing water.

Thus an important aspect of soil erosion by overland flow processes is that the rainsplash impact is needed to help detach the particles from the soil surface, and the overland flow is needed to transport the detached particles downslope. The need for both these processes to produce effective soil erosion has been shown clearly by Ellison (1945) in an experiment in which he subjected a bare soil surface to different combinations and sequences of rainsplash and overland flow and rainsplash (Fig. 3.2).

Changes in the erodibility of the soil are also important. In stage 1 the erosion rate is high because a dry, friable surface layer is readily removed by overland flow; as soon as this layer is removed a more compact, less detachable layer is encountered and the erosion rate decreases. The increase in soil moisture content also tends to reduce the ease with which particles can be removed (Grissinger, 1966). The rate of erosion rises again when the rainsplash is added, as the raindrop impact detaches more particles (stage 2), but then declines. A brief increase occurs after overland flow is removed, probably because the impact of raindrops is more effective when the surface film of water is reduced in thickness (stage 3). There is another rapid increase when overland flow is re-instated, again reflecting the removal of a detached surface layer (stage 4), but decreasing soon afterwards, to rise yet again when both rainsplash and overland flow occur (stage 5). The general sequence show, however, that the amounts of soil removed become smaller, suggesting that the soil is becoming less erodible and that a drying period would be needed to reproduce the high initial erosion rates.

These aspects are emplasised in the early studies of soil erosion by Ellison (1947, a-g).



Time elapsed.

Fig. 3.2 Variations in sediment concentration in runoff as a function of time and combinations of rainsplash and overland flow.

Thus, overland flow's major role in soil erosion is that of transporting particles which are already detached. The difference between detachment and transport must be realised, and that both are necessary in soil erosion. Ellison (1947a) speculated on the detachability and transportability of different soils and came to the following general ranking:

	Detachability	Transportability
decrease ↓	fine sand	clay
	loam	loam
	clay	fine sand

A dramatic example of the need for rainsplash is given by Hudson (1971). He covered one bare plot of soil with gauze, another he left bare and another he covered with grass. Erosion rates from the bare, uncovered plot were about 100 times as large as from the other plots. This Hudson (1971) ascribes to the reduction in rainsplash caused by the gauze netting. The grass is equally effective in reducing rainsplash and increasing infiltration and decreasing overland flow. Both Ellison (1947e) and Hudson (1971) emphasise the importance of rainsplash in soil erosion. Rainsplash has been found to be important in some areas, such as bare sites in temperate forests (Kwaad, 1977), but a recent study by Morgan (1977) of erosion processes on bare sandy soils in England, showed that overland flow was a more important contributor to soil erosion than rainsplash.

It is extremely difficult to separate out the overland flow and rainsplash components of soil erosion. For example, in Hudson's (1971) study with the gauze netting, the increase in erosion on the unprotected plots is partly due to compaction of the soil surface under raindrop impact. Ellison (1947g) also observed that turbid overland flow passing through a parking lot deposited a layer of sediment beneath parked cars. This suggests that the raindrop impact is also important as a mechanism for increasing the turbulence of overland flow and increasing its transporting capacity. Thus, the separation of rainsplash and overland flow erosion is a rather academic exercise since they are so synergistically inter-related in nature.

(iii) Rills and gullies

Overland flow is the unconcentrated form of surface runoff; where topographic variations exist across a slope, then the surface runoff is likely to become concentrated into small depressions. As such, the erosive ability of the flowing water is concentrated into a smaller area, and the channel is likely to expand. There is a continuum of sizes from very small channels through to streams and rivers.

The smallest member of this continuum is the rill, which can be defined as:

"localised small washes in defined channels which are small enough to eliminate by normal cultural methods"; gullies are somewhat larger:

"when they are so large and well established that they cannot be crossed by agricultural implements" (Hudson, 1971).

The above definition is really an agricultural separation based on their impedance to agricultural machinery. Gregory and Walling (1973) suggest that rills and gullies can be separated by criteria of 0.3m width and 0.6m depth. The transition into streams and rivers is not so well defined; an important characteristic is that rills and gullies are ephemeral, in that they carry water only after heavy stirms, whereas streams and rivers flow for a much longer period and possess a base flow derived from local ground water sources.

Rills, gullies, and streams possess an erosive ability through the energy possessed by the flowing water. The rate of water movement is dependent on the downslope component of gravity acting on the water body, and the resistances offered by the channel sides. The velocity of flow can be calculated from the following formula:

$$V = \frac{R^{2/3} \quad S^{1/2}}{n}$$

where V = velocity (m/sec);

R = the hydraulic radius of the channel, i.e. the cross-sectional area of water ÷ the wetted perimeter;

S = the slope of the water surface, expressed in m/m;

n = the roughness of the channel bed and sides.

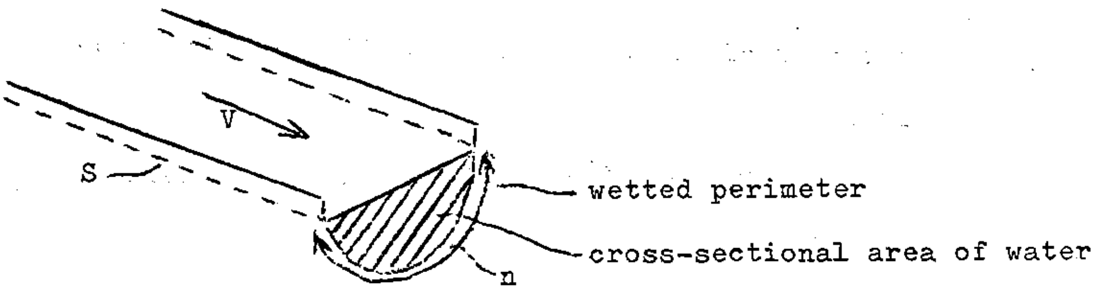


Fig. 3.3 Hydraulic characteristics of a river channel.

This empirically-based formula is known as Manning's Equation. Commonly-found values of 'n' are given in Table 3.1 (Gregory and Walling, 1973; Hudson, 1971).

Table 3.1 Some commonly observed values of Manning's "n"

		n
Streams:	sluggish, weedy, deep pools	0.07
	clean, straight, no pools, fine sediment, flat bed	0.02
	gravel/cobbles on bed	0.04
	cobbles/boulders on bed	0.05
Flood plains:	short grass	0.03
	mature crops	0.04
	dense bush	0.08
Vegetated channels:	short grass	0.045
	medium grass	0.06
	long grass	0.09

Roughness values have also been calculated for overland flow, with depths usually less than 10 mm, and values range from about 0.25 for sparsely vegetated surfaces to about 0.60 for a thick grass cover (Emmett, 1970; Pearce, 1976; Ree et al., 1977).

As the stream gully or rill flows it converts its potential energy (based on its height above the "base level") into kinetic energy. Most of this kinetic energy is dissipated as heat created by internal friction in the waterbody and friction created on the channel bed and sides. However, about 3 to 5% of this energy can be used to transport sediment within the channel.

The flowing body of water can exert stresses on particles on the channel bed, which tend to raise the particle. The higher the velocity of the water, the greater the stresses, and, thus, the more upward movement of the particles. Therefore, as the velocity increases, larger and larger particles can be moved. The term "competence" refers to the largest particles which can be moved by a stream, and is mainly a function of the velocity of the stream, though other factors, such as particle density and cohesive forces between particles will also be important.

The sediment moved in a channel can, rather arbitrarily, be separated into bedload and suspended load. Bedload refers to the material which moves by sliding, rolling and saltating along the channel bed, and, therefore, moves very close to the bed. Suspended load is material which is borne by the upward fluxes of eddies, and, thus, travels away from the bed. The bedload contains the heaviest or largest particles, whilst the suspended load includes the lightest and smallest, though there will be an interchange between the bed and suspended loads as the flow changes through the channel.

The type and amount of material which can be carried in the channel depends on the flow characteristics and the availability of material from the channel sides and bed. Two characteristics are important. One is the ability of the flow to remove particles, the erosive characteristic, and the second is the ability of the water to transport the particle, the sedimentation characteristic. These characteristics are strongly related to the velocity of flow, and this affects the stresses created on bed material, and also the eddies which keep material in suspension.

The relationship between velocity and erosion and sedimentation of various particles of different sizes is represented in Fig. 3.4 (Morisawa, 1968).

It shows that the "erosive" velocity increases as does the particle diameter, except for an increase in the clay and silt-size fractions. This is because these fine particles possess cohesive forces which help bind the particle to the channel bed and so increase the velocity required for erosion. The "sedimentation" velocity shows that, again, the velocity required to transport increases as does the particle size.

The graph is the basis of the identification of critical velocities for cutoff channels, i.e. the maximum velocity that can be allowed without important erosion of the cutoff channel (Table 3.2, from Hudson (1971)).

Table 3.2 Maximum velocities permissible in cutoff channels.

	Maximum or critical velocity m/sec based on channel slope of 0-5%		
	bare	moderate grass cover	very good grass cover
silty sand	0.3	0.8	1.5
coarse sand	0.8	1.3	1.7
firm clay loam	1.0	1.7	2.3
stiff clay	1.5	1.8	2.5
coarse gravel	1.5	1.8	-
hardpan, shale	1.8	2.1	-

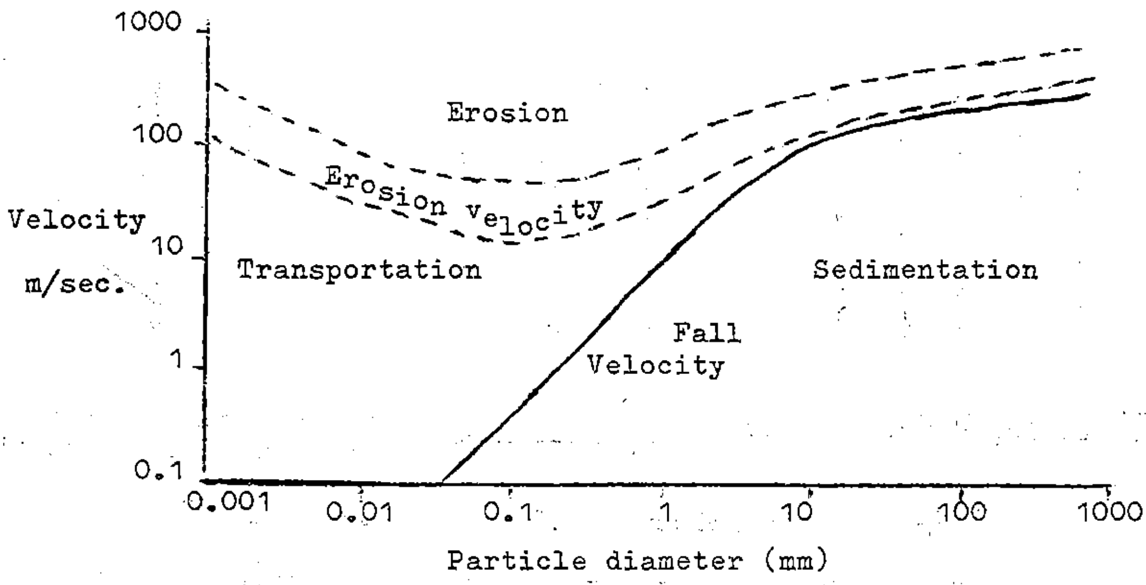


Fig. 3.4 The fate of particles in relation to stream velocity.

As overland flow becomes concentrated into small channels, such as rills and gullies, the hydraulic radius increases and the velocity increases. As the velocity increases, the ability of the water to transport material, both quantitatively and in terms of particle size, also increases, resulting in an incision of the channel bed and an increase in the channel gradient in the upper part of the channel section. Incision in the headwater area occurs because of the steepened gradient, resulting in the "capture" of more overland flow and inferior rills, resulting in more erosion (Fig. 3.5). Thus, a rill or gully network develops on a slope. An important stage in this process is the formation of a master channel which is not destroyed each year by cultivation practices. Where rills are destroyed each year, new rill formation will be slow and there will be little opportunity to concentrate the erosive power into permanent channels. When the major channel is not destroyed by cultivation, a more developed network will occur, with a much more incised rill and gully network. Once formed, gullies tend to be self-perpetuating.

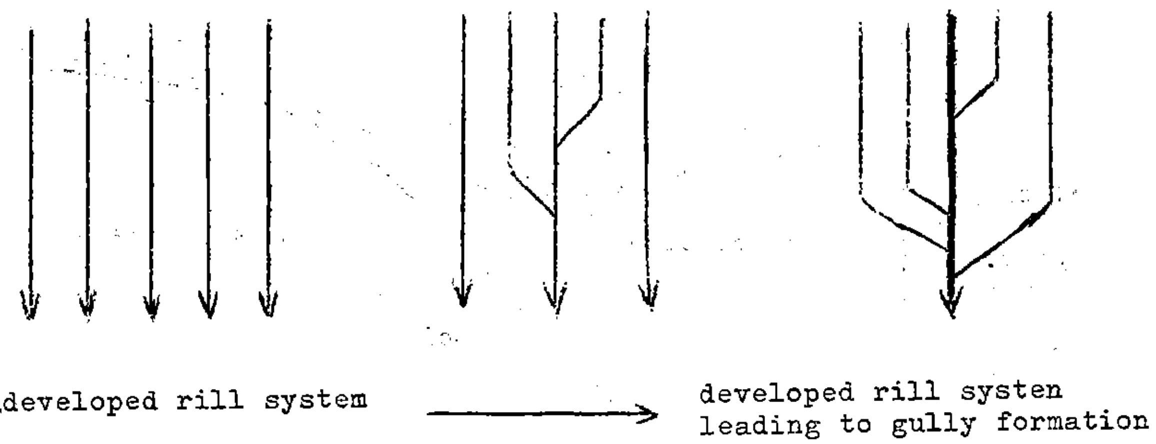


Fig. 3.5 Development of rills into a gully.

What factors control the formation of rills and gullies? Vegetation cover is important, in that it increases infiltration rates, and slows down and diverts overland flow. Most accounts of well-developed gully systems come from areas with semi-arid climates, though they can also be found in temperate areas with a normally denser vegetation cover. The intense rains found in semi-arid areas favour the development of overland flow and rill and gully initiation, and the generally sparse vegetation does not slow down overland flow. Once initiated, gullies can extend through scouring of the bed, maas movements along oversteepened channel walls and headward extension. Headward extension appears to be encouraged by sapping caused by pore water pressures at the head, and a plunge-pool effect during storms.

Seginer (1966) and Thompson (1964) have developed models to predict the rate of gully headward extension, in Israel and U.S.A. respectively. They developed regression equations from observed rates of extension and local characteristics. Seginer (1966) found the area to be the most useful characteristic, and the regression

$$Y = b\,A^{0.50}$$

where Y = headward extension (m)

b = constant (2.1 to 6.0)

A = area in km^2, of basin which gully drains

explained 40 to 70% of the variation in headward extension. Thompson (1964) also found area of catchment to be important, and also included slope of approach channel, depth, a rainfall factor and a soil factor. The combined regression equation explained 77% of the variation in headward extension. The Seginer (1966) study suggests that headward extension increases to a peak and then decreases, presumably as the catchment area decreases, giving a "life-cycle" for gully development.

The most extensively studied gully networks appear to be those in the semi-arid S.W. U.S.A. These "arroyos" increase headwards at rates of several m/yr, and the mechanisms of initiation and development are probably applicable to the semi-arid areas of East Africa. Many gullies can be related to certain man-induced changes, such as conversion from grassland to cultivated land, channeling along roads and tracks, fences and buildings, and cattle tracks. In the S.W. U.S.A., sheep overstocking and consequent decrease in vegetation cover was originally cited as the cause of gully development. However, it also appears that changes in climate, particularly towards a more semi-arid one with lower total rainfall but higher intensity summer rainfalls, contribute to gully development (Denevan, 1967; Malde, 1964; Malde and Scott, 1977; Yi-Fu Tan, 1966). The same combination of land use changes and intensified land use may also be applicable to some East African gully networks, such as in Machakos District, Kenya and Dodoma, Tanzania (Rapp *et al.*, 1972b).

(iv) Relative importance of erosion forms

As can be seen, there are a wide range of methods whereby soil can be eroded. Each is dependent on certain environmental conditions, and, therefore, will be most important in different environments. Unfortunately, there are few published studies which compare erosion rates for all the methods, in well-defined areas.

Gully erosion is the most spectacular form of erosion, and most erosion books contain impressive photographs of gully networks. However, in terms of damage to agricultural lands, it is not very important in that gullies often cover only a small proportion of the land surface, and the gullied soils are generally not very productive (Hudson, 1971). Gullying seems to be most important in semi-arid areas. Rilling can be important, such as noted by Rapp *et al.*, (1972a) in the Morogoro catchment, Tanzania.

Landslides are also mainly restricted to certain areas, generally with steep slopes, unstable soils, and high rainfall amounts. Two cases have been noted above - in Barbados (Carson and Tam, 1977) and the Mgeta area, Tanzania (Temple and Rapp, 1972).

Soil creep, although an ubiquitous process, is also regarded as an ineffectual mechanism of soil erosion. This is due partly to the slow rate of erosion and partly to the fact that the topsoil moves slowly downslope, so that only the uppermost section of the slope is likely to have infertile, exposed subsoils.

As noted above, the differentiation between rainsplash and overland flow is difficult and rather academic. Bennett *et al.*, (1951) regarded overland flow as being more important than rainsplash, whereas Hudson (1971) states that the reverse is the case. Few studies have been able to separate the two, though Morgan (1977) found that overland flow was more important than rainsplash in eroding soil from a bare, sandy soil in England. Part of this confusion probably arises from the spatial properties of the two processes (overland flow is usually discontinuous whilst rainsplash is continuous), and that an important component of rainsplash is the detachment of particles, which can then be transported by overland flow.

The combined effect of overland flow and rainsplash erosion is the most important in agricultural areas and overall denudation rates of 10 mm/yr are not uncommon. Rapp _et al._, (1972b) found this mechanism to be the most important near Dodoma, Tanzania. Glymph (1957) found that sheet and rill erosion contributed between 10 and 100% of the sediment from watersheds in U.S.A., and Hudson (1971) notes that gullying may become more important in semi-arid and tropical climates.

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4. MEASURING SOIL EROSION

There are several ways in which soil erosion may be measured, and these may be conveniently divided into *in situ* methods, the estimation of erosion rates from drainage basin sediment yield measurements, and those involving remote sensing, usually air photographs.

(A) In situ methods

There are numerous *in situ* methods available, depending on the type of erosion encountered and the attributes of the terrain.

One of the simplest methods used to measure the downslope movement of soil is to insert a trough along the contour of the slope. There should be a good contact between the upslope wall of the trough and the soil, and this can be achieved with a plastic sheet or concrete. A poor contact will result in the erosion of the soil near the trough lip, or the diversion of runoff water beneath the trough. The runoff and erosion rates can be converted to an areal basis by measurement of the upslope catchment of the trough. Where this is not feasible, the edge of the plot can be bounded by metal or concrete walls, though this may introduce edge effects into the plot. Gerlach (1967) describes a design suitable for many conditions.

Based on these designs, many permanent plots have been established, though their size varies. In the United States, plots 70 to 90 ft long and 0.01 to 0.20 acre in area were used, and the results from these plots formed the basis of the Universal Soil Loss Equation (Wischmeier, 1959). Othieno (1975) used plots 60 x 12 ft in tea fields near Kericho; Lal (1976) used plots 25 x 4 m in Nigeria; and Roose (1967) and Roose and Lelong (1976) measured runoff and erosion from plots 16 x 6 m or 200 to 600 m^2. Temple (1972) reviewed the results of runoff and soil erosion from four series of plots established under different conservation systems, length and slope in Tanzania. In general, the larger the plot the more readily applicable the results are to field conditions. The Universal Soil Loss Equation does include a conversion factor for plots of different slope and length.

Where soil erosion by overland flow is not very important, but where measurements of soil creep are required, then problems are presented by the very slow rate of movement and the delicate nature of the processes involved. One method is to dig a soil pit and place a narrow strip of aluminium plates into one of the side walls. Bench-marks are established, the position of the plates measured, and the pit refilled. When measurements are required again, the pit is carefully re-excavated and the position of the plates relocated. Rates can then be calculated as the displacement downslope over the time period elapsed (Young, 1960). Flexible tubes or glass beads are an alternative method. Rates of movement are only a few mm/yr, though they can be higher on steep slopes (up to 60 mm/yr).

The effect of rainsplash can be estimated by placing boards about .5 m high at right angles to the slope, and with small troughs at the base, on each side. The troughs are protected from direct rainfall and the upslope and downslope components can be weighed, the difference representing the net downslope movement over the width of the board (Ellison, 1944; Kwaad, 1976). An alternative method has been proposed and used by Morgan (1976), which removes any wind turbulence effects on raindrops around the board. This method involves a 10 cm diameter, exposed soil surface surrounded by a tray with an outer wall 10 cm high; the tray is divided into upslope and downslope compartments, and the amount of soil in each compartment is measured. The net downslope movement is the difference between the two, expressed in g per width of slope, calculated from the area of exposed soil. Radioactive tracers can also be used (Coutts _et al._, 1968), as can fluorescent beads (Young and Holt, 1968).

An alternative method to measuring the soil removed from a plot is to measure the lowering of the land surface. This can be done either by establishing a series of benchmarks near the soil surface or measuring the soil surface against some natural phenomena.

Perhaps the best example of the former approach is the use of erosion pins, which are nails up to 30 cm long, driven into the soil, with a washer between the soil surface and the nail head. The pins can be laid out in a network and the distance between the washer and the nail head measured accurately. After a period of erosion, the distance can be remeasured. Nails with large heads will tend to protect the soil from rainsplash effects, so that the washer may stand above the surrounding surface. In this case, the soil pedestal should be removed so that the base of the washer is flush with the level of the surrounding soil surface. Where deposition is likely to occur (such as on footslopes and flood plains) then the nail should be left with up to 10 cm above the soil surface; in other cases, the nail head should be within 2 cm of the soil surface, to reduce the possibility of damage.

Erosion pins provide a cheap, quick way of obtaining erosion data. For example, Rapp et al., (1972a) were able to measure an average erosion loss of 2.5 mm of soil within a two-week period during the rainy season on a 4° slope near Morogoro, Tanzania. Temple and Murray-Rust (1972) used stakes to measure erosion rates [illegible] hill rice and regenerating bush at Mfumbwe, Tanzania. They found annual soil losses of up to 28 mm/yr under hill rice, while the various conservation practices reduced the erosion rate. Regenerating bush reduced the erosion rate to only a few mm/yr. Frequent measuring of pin levels can allow a relationship to be established between rainfall characteristics and erosion rates.

There are several disadvantages with the method, however. It has to be established that the nail or pin does not seriously affect the erosion rate, nor that disturbances by stock or soil animals or expansion and contraction of the soil affect the pin:soil levels. Pins may also be removed by local people. A further complication is that there is frequently a large difference in erosion rates between pins on the same slope, emphasising the spatial heterogeneity of the erosion processes. Thus, a large number of pins are required to obtain reliable results. On bare, eroding slopes in Canada, Pearce (1976) used pins with a density ranging from $1/m^2$ to $0.4/m^2$. The standard errors calculated are variable, depending on the topographic variation within each plot, but range from 16 to 579% of the mean.

A similar approach for measuring changes in surface elevation is to record the height of certain natural features above the present surface. If these features can be assumed to have been level with the soil surface, then their present height represents the amount of erosion. Stone pedestals can often be used, though the problem of dating exists.

Vegetation which produces a root system or another part of the plant close to the soil surface can also be used. As erosion removes the soil, the old plant markers will be left above the new surface, or the tree or bush will have protected the soil against rainsplash, to produce a mound. When the plant can be aged, the amount of erosion can be converted into an annual rate.

Lamarche (1968) used this technique to measure erosion rates in a mountainous area of California, specifically by exposure of <u>Pinus aristata</u> (bristlecone pine) roots and tree-ring dating. Dunne (1977a) measured root exposures and mounds beneath <u>Acacia drepanolobium</u>, <u>A. tortillis</u> and <u>Cericio comopsis palsidia</u> in semi-arid areas of Kenya. The age of the plants was established from a regression of age on girth; it was assumed that the bimodal rainfall regime produced two vegetative flushes and two rings per year. Haggett (1961) used coffee collars to measure erosion rates on old plantations in Brazil.

Grasses and sisal may also be useful, but the problem of dating is more severe and these have not been tested. Preliminary observations of exposed roots of dead grasses and grass mounds suggest erosion rates of about 2 cm/yr in Machakos. The problems inherent in this method are establishing that the plants have not occupied pre-existing mounds and that the plant morphology does indicate the original surface. Rainsplash accumulation may also increase the relative height of the mounds.

Differences in soil profile morphology can also be used. Haggett (1961) measured depth to the C horizon in forested and coffee plots in Brazil, and estimated that erosion had removed 20 cm of topsoil upon the conversion of forest to coffee. Morphological differences were supported by differences in soil texture. Stone lines may be useful in that they frequently occur at depths of about 1 m in noneroded hillslope soils, and can occur as stone pavements in some sites, leading to a speculation of about 1 m of erosion. Again, the main problems in this approach are establishing a typical non-eroded profile and being able to establish that this profile existed in all locations prior to erosion. Obviously, profile differences caused by topography and "natural erosion" will have to be taken into account.

Routine survey techniques can be used for large scale changes, such as the headward and lateral movement of gullies. Benchmarks close to the erosion features and stakes located in the gully walls have been utilized. Frequent measurements can be used to estimate the volume of sediment removed. An example is given by Malde and Scott (1977). Similarly, Rapp et al., (1972b) measured landslide volumes in the Mgeta area, Tanzania by survey techniques, supplemented by air photo interpretation.

All the above methods depend on measurements of soil erosion which occurs naturally. An alternative approach is to produce erosion experimentally and to relate this to naturally-occurring erosion. This is usually achieved by simulating rainfall in storms of standard intensity, duration and amount. The design and operation of rainfall simulators has been covered in the literature (e.g. Hall, 1969, 1970; Meyer, 1965). The resulting soil erosion can be measured by changes in elevation of the surface soil, or by weighing the soil removed from the plot and collected in a trough. The rainfall simulator approach is very useful in determining the effect of factors such as soil, slope, crop type and conservation practices on soil erosion and surface runoff. The main problems are in the relationship between the artificial rainstorm and natural rainfall, in both type and frequency/duration/intensity. Young and Burwell (1972), however, have been able to show a good agreement between runoff and soil erosion from artificial and natural rainstorms for a soil in Minnesota.

(B) Drainage basin methods

The above *in situ* methods refer to individual plots or areas of soil and thus require replications to achieve an acceptable average for a large area. An alternative approach is to examine erosion rates for larger areas, usually defined catchment basins, by measuring the removal of soil in the stream or river leaving the basin. This method has the advantage that the integrated losses of soil can be recorded at one point, but frequent measurements are required and there are problems in the extrapolation of erosion losses from individual areas in the drainage basin.

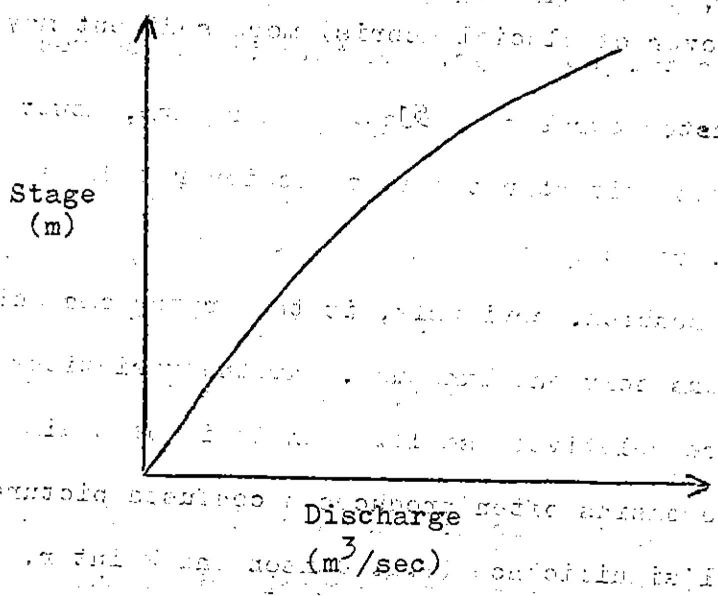
The material removed from a drainage basin by a stream or river can be divided into three types. The solute load represents the material in solution, and is derived mainly from nutrient cycling, leaching, and weathering. The suspended load is the material carried within the body of moving water, usually sand to clay sizes. The bedload is the material carried on the channel floor by sliding, rolling, or saltation, usually as coarse as or coarser than sand. Soil erosion rates refer only to the solid load, so the solute load can be ignored.

The suspended load is calculated through measurement of the discharge in the channel and the sediment concentration. Discharge is the volume of water passing through a section of the channel, usually expressed in units of cubic feet per second (cfs) or cubic metres per second (m^3/sec). It is calculated by measuring the cross-sectional area of the channel occupied by water and the mean velocity of flow:

$$Q = V \cdot A \quad \text{where} \quad \begin{aligned} Q &= \text{discharge} \\ V &= \text{mean velocity of flow} \\ A &= \text{cross-sectional area.} \end{aligned}$$

The area can be measured by survey of the stream channel, and the mean velocity is estimated by measurements of velocity at various depths in and across the channel. This is required because the velocity will be low where frictional drag occurs, such as along the channel bed and sides. A current meter is lowered into the water at distances across the channel, and the velocity measured at various depths. For deep channels, the mean of the velocities at 0.2 and 0.8 of the channel depth can be used, or at depths of 0.15, 0.50 and 0.85, or as an integration of the whole depth profile. In shallow channels, one measurement at 0.6 of the depth can be used (Gregory and Walling, 1973).

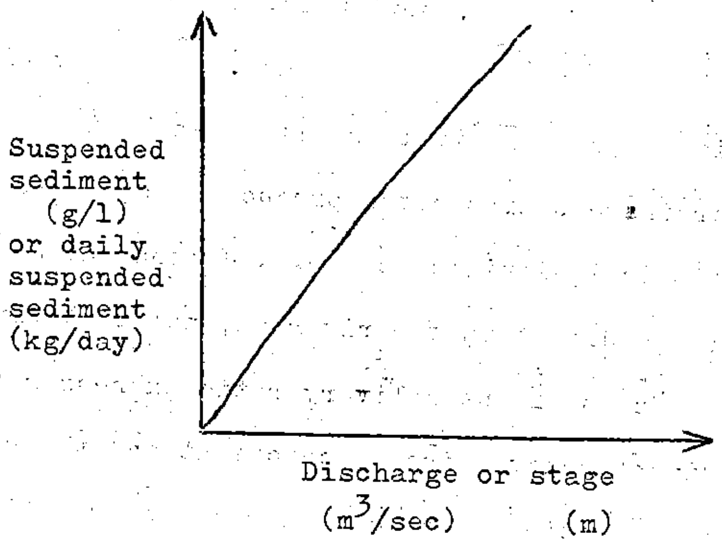
A stage curve can then be produced by repeated measurements of discharge at different times. A convenient method of expression is to plot the discharge against the level of the water surface, as measured against an object on the bank, such as a stake. Rapid discharge estimates can then be made by merely reading the height on the stake, and converting to discharge.



Frequent measurements of discharge allow the construction of a discharge:time curve, known as a duration curve.

Suspended sediment can be measured by collecting water samples and drying or filtering to obtain the weight of the sediment. As for discharge, sediment concentration will vary through the water body, and representative samples need to be taken. A depth-integrated sampler is commonly used (Dunne, 1977b).

Multiplication of discharge and sediment concentration at specific times gives a suspended sediment rating curve:



From this curve, the suspended sediment load can be calculated for a stream on a daily basis. The annual hydrograph or flow duration curve can then be used to estimate the number of days with discharges of certain magnitude and transformed into an annual sediment loss from the drainage basin.

Bedload movement is much more difficult to measure, and usually involves a trap laid into the channel bed. Bedload:discharge curves can be produced to estimate bedload movement on an annual basis, as for suspended sediment (Dunne, 1977b; Gregory and Walling, 1973). Bedload formulae exist, but are not used very frequently in erosion studies.

The suspended load:bedload ratio depends on many factors, e.g. channel size and shape and particle size, but a commonly used estimate is that bedload is about 20% of the suspended load. Summation of suspended and bed loads gives the total load, i.e. the total amount of solid material removed from the drainage basin by the stream or river.

Thus, the sediment loss from a drainage basin may be expressed in m^3/km^2 yr or tonnes/km^2 yr, and this is the commonest way to express soil erosion rates for large areas. There are, however, problems in translating the sediment rates into erosion rates for surfaces within the catchment.

One problem, obviously, is that rates can only be applied when the basin covers only one land use, and the variations in topography and climate are relatively small. Analysis of sediment yields from large drainage basins often produces a confused picture, with low levels of statistical significance (e.g. Janson and Painter, 1974). Furthermore, the sediment in the river may not be derived from the surrounding soils. The sediment delivery ratio expresses the relationship between the sediment removed from a catchment and the soil loss from the slopes. This is a function of catchment size and relative relief, soil erosion and channel scour processes. In some catchments (such as temperate are with a thick cover of glacial debris) more sediment may be removed than soil eroded from the hillsides. In most cases, however, more soil is eroded than sediment removed; the soil eroded but not removed remains on footslopes, debris fans and flood plains.

Dunne (1977b) quotes examples of changes in sediment delivery ratio with drainage basin area and relative relief, based on southeastern U.S.A. data. The ratio decreases from about 80% with drainage basin areas of 0.05 km^2 to about 10% in drainage basins of 1000 km^2. The ratio increases with an increase in relative relief, expressed as the ratio of catchment relief to mainstream length. At height/length ratios of 0.002 the sediment delivery ratio is about 4% and rises to about 80% in basins with a height/length ratio of about 0.05.

Thus, meaningful comparisons of the influence of land use practices on sediment yields and erosion rates can only be made when individual catchment values are converted to a standard basin size and height:length. Rapp _et al._, (1972a) give an example of this approach for a Tanzanian catchment.

Sediment generally accumulates in reservoirs, and these deposits can be used to estimate sediment yield and erosion rates. The volumes of accumulated sediment can be established by survey and comparison with original plans or by location of the original reservoir bed. Dunne (1977b) gives a table of bulk densities whereby sediment volume can be converted into weight. A further correction (in addition to the sediment delivery ratio) needs to be applied as not all the sediment carried by the river will be deposited. This will depend on the particle size of the sediment (clays less likely to be deposited than sands) and on the ratio between the capacity of the reservoir and the annual flow. Dunne (1977b) presents a curve which shows that the sediment trapped increases from only 10% with a capacity:flow ratio of 0.003 to 95% with a ratio of 0.2 and remains constant with further increases in the ratio.

Division of the amount of accumulated sediment by the age of the reservoir allows an estimation of the annual sediment yield. More precise rates, and changes in rates of sedimentation can be estimated from the dating of sediment levels, such as by radiocarbon or lead-210 isotopes (Dunne, 1977b). Rapp *et al.*, (1972b) have measured sediment yields in four reservoir-catchments in semi-arid Tanzania.

(C) Remote sensing

Aerial photography offers advantages to soil erosion measurement in that intensive field work is not required, and photographs can act as a historical record. Thomas (1974) measured land use changes and gully extension by comparing two sets of aerial photographs in Machakos District. The photographs could also be used to delimit areas where tonal changes suggest exposure of subsoils or stone lines, indicative of soil erosion. Temple and Rapp (1972) were able to map and measure landslides from air photographs of the Mgeta area, Tanzania, and a correlation between landslide area and volume can be obtained, allowing the rapid calculation of erosion rates.

Jones and Keech (1966) have discussed methods of identifying soil erosion areas from aerial photographs, and Williams and Morgan (1976) have developed a geomorphological mapping method applied to soil erosion.

The disadvantages with aerial photogtaphs are that they can be of variable scale and distortion and are only suitable for relatively large scale lateral movements of features, such as gullies. Although height measurements can be made with stereoscopic pairs, the accuracy and precision is generally unsuitable for measurements of surface lowering or rill or gully deepening.

Multispectral analysis from space satellites or air photographs may prove to be useful in delimiting eroded areas through their influence on moisture and temperature characteristics (Rango, 1977).

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5. THE MAGNITUDE:FREQUENCY OF EROSIVE EVENTS

A major question that needs to be asked in soil erosion problems and their solution is: does the erosion occur as infrequent, intense events or as frequent, small scale events? The design of conservation measures must take these magnitude:frequency relationships into account.

Magnitude:frequency analyses have been developed for many common and important events. One good example is that for the high discharge of rivers, which cause flooding and economic damage. The magnitude:frequency of floods can be calculated by observing the annual maximum discharge, ranking these discharges and obtaining the recurrence interval of each discharge (flood), which is given by:

$$r.i. = \frac{n+1}{m}$$

where r.i. = recurrence interval (years)
n = number of years of record analysed
m = rank of annual flood.

When plotted on special Gumbel probability paper, the points often occur as a straight line, so that the recurrence interval of any discharge can be predicted, or, alternatively, the magnitude (discharge) of the maximum flood likely to be experienced within any given period. Land use on flood plains and flood design structures are often developed around the "50 year flood" discharge.

A similar approach can be used for precipitation, either on a daily or annual basis, or, more commonly, for specific time periods. An example of this is given by Taylor and Lawes (1971), who have compiled a magnitude:duration:frequency graph for 16 East African stations for periods ranging from 15 minutes to 6 hours. The same method of analysis has been used, except that groups of rainfall amounts have been utilized, rather than individual measurements. Thus, the amount of rain that can be expected within one hour can be predicted for recurrence intervals of 1 to 100 years. It is unwise to extrapolate for recurrence intervals longer than the period of record:

In assessing the magnitude:frequency relationships of erosive events, the characteristics of the area studied and the processes involved will affect the dominant event. The dominant event (or events) is that which, over a period of years, causes the most soil erosion. In particular, the dominant event's magnitude and frequency will be dependent on:

1. the climate, particularly the magnitude:frequency relationships of precipitation;
2. the soil type;
3. the antecedent conditions, particularly the moisture content and soil strength prior to the erosive events;
4. the vegetation type and its density;
5. the dominant type of erosion process.

As Hudson (1971) has noted, studies have produced different results for magnitude:frequency analyses. Many of the differences can probably be explained in terms of the variations of the above five factors. For many rivers in the United States, it appears that it is the bankfull discharge which moves much of the sediment in a river channel and essentially shapes the river channel (Leopold et al., 1964). This bankfull discharge has a frequency of once every one or two years. However, as the river flow becomes more variable and the drainage basin decreases in size or becomes steeper, the extreme flows, less frequent than once per one or two years, become more important in transporting sediment.

Compared to river studies, there are relatively few studies on the magnitude:frequency relationships of erosive events on agricultural land.

In the United States, Wischmeier (1962) found that the overall pattern is for a large part of erosion to be carried out by the frequent, small storms. This is not the case in all areas, as Hudson (1971) quotes an example from Missouri in which 50% of the erosion of a five year period was caused by a single storm. Also for the central United States, Greer (1971) found that excessive rate storms (defined by intensity and duration characteristics) caused over 50% of the soil erosion from small plots in a six year period, yet represented only 6% of the total rainfall.

In bare plots in Canada, Pearce (1976) showed that the dominant erosive events were storms of 1 to 6 hours duration, which occurred about once per year. For shorter durations, the less frequent storms were important. In England, Morgan (1977) found that 99% of the erosion from a cultivated sandy soil over a two year period could be related to only 10 storms, from a total of 240. Over 50% of the erosion was caused by 4 storms.

All of the above analyses deal with erosion by overland flow and rilling in temperate climates. For rainsplash erosion, the Arizona study quoted by Carson and Kirkby (1972) shows that, on vegetated plots, the dominant events had recurrence intervals of 0.5 to 1 year. Also in southwest U.S.A. Malde and Scott (1977) have reported on the headward extension of arroyos (gullies), and, for one gully, the 13m extension over a five year period was related to the major 4 or 5 rain-storms which produced runoff. Antecedent conditions, particularly moist, unstable gully walls, also affect the effectiveness of a runoff event. Both the above studies occurred in a semi-arid environment.

Studies of mass movements, such as landslides and earthflows, suggest that most movements occur infrequently, possibly at intervals of 5 years or more, though this will depend on the climatic characteristics. Landslides in mountainous areas suggest this relationship (e.g. Paeth _et al._, 1971; Temple and Rapp, 1972; Rice and Foggin, 1971). Not surprisingly, in view of its slow rate and delicate mechanism, there appear to be no magnitude:frequency studies of soil creep, though one might expect the frequent, small events (such as wetting and drying cycles and animal activities) to dominate.

Few magnotide:frequency studies of soil erosion have been made in tropical and subtropical environments and many studies are restricted by their short duration. In Rhodesia, Hudson (1971) found that, in almost all years, more than half of the total erosion was caused by the one or two heaviest storms. The results of erosion studies performed by other workers in Africa do permit some estimates of the magnitude:frequency relationships.

Roose (1967) reported soil losses from bare plots in Senegal over a five year period. Expressed in terms of the amount of rainfall in each storm, the results are:

Rainfall amount (mm)	Frequency (no./yr)	% mean annual rainfall	% mean annual runoff	% mean annual soil erosion
0 - 15	16.0	13.4	3.6	4.2
15 - 30	14.2	27.2	19.8	27.1
30 - 60	10.0	39.1	43.3.	41.3
60 - 90	1.6	9.5	11.7	19.7
90 - 120	0.4	3.5	7.0	2.5
120	0.6	7.7	14.6	5.3

The results show that the rainstorms of moderate amount (30 to 60mm) and moderate frequency (10 times per year) cause almost half of the annual soil erosion.

Lal (1976) measured soil erosion from bare fallow plots on slopes of 1, 5, 10, and 15% on an Alfisol near Ibadan, Nigeria. He calculated the erosivity parameter R ($EI_{30} \div 100$) for each storm, and measured the soil loss from each plot. The data can be converted into the following table, based on three years' results:

R value	Frequency (no./yr)	Soil loss 1% slope	(% total) 5% slope	- Soil loss 10% slope	(% total) 15% slope
1 - 5	18.7	23.1	25.7	17.9	14.6
5 - 10	4.0	8.7	10.1	9.2	9.1
10 - 20	6.7	26.8	31.8	27.9	30.3
20 - 30	2.0	9.7	12.1	11.1	7.6
30 - 40	0.3	1.6	3.6	1.6	1.0
40 - 50	0.3	1.6	0.9	1.0	3.3
50 - 100	2.3	25.1	14.4	30.3	26.8
100 - 200	0.0	0.0	0.0	0.0	0.0
200	0.3	3.3	1.4	1.1	7.3

It shows that over half of the erosion is caused by storms which occur with a frequency of 2 to 7 times per year. It is also interesting to note that the heaviest storm (105mm in just one hour, with a mean intensity of 97 mm/hr) caused only 1 to 7% of the total erosion, despite the fact that a further 58 mm of rain had fallen in the previous week. The Senegal and Nigerian experiments occurred in similar climates, with about 1200 mm of rain per year, distributed in one rainy season.

Othieno (1975) and Othieno and Laycock (1977) measured soil losses from plots near Kericho, Kenya. The measurements covered three years, during which time a tea crop was developing, and the intervening bare soil was either kept free of weeds by tilling or by the application of a herbicide (non-tillage). The mean annual rainfall is 2160 mm. Othieno gives both the monthly soil losses and losses for individual storms, as a function of rainfall intensity. Unfortunately, of the six plot-years available, only four show good accordance between the tabulated monthly losses and those shown as a function of individual storms. For these 4 plot-years the results of a magnitude:frequency analysis are:

Year and treatment.	Number of erosive storms	Number of storms to give 50% soil loss
1971-72 non-tillage	32	6
1972-73 tillage	17	3
1972-73 non-tillage	22	5
1973-74 tillage	16	4

Again, the results suggest that it is the events which occur 3 or 6 times a year that cause most of the soil erosion. The tea canopy developed from 1 to 20% in 1971-72 to 60 to 70% in 1973-74, with a subsequent decrease in soil lost (from 160 tons/ha in 1971-72 to about 4 tons/ha in 1973-74). The results for the above plots, and for the plots treated with mulch or oats between the tea, suggest that with an increasing cover the more intense and infrequent rainstorms produce a larger proportion of the soil loss. This is to be expected, as the vegetative or mulch cover would protect the soil from all but the most intense storms.

Unfortunately, the tea study is the only detailed one which allows an analysis of magnitude:frequency in East Africe, and, even then, it is located in a highland area with a high rainfall. A preliminary, indirect analysis of soil erosion could be made in East Africe by using the rainfall intensity:duration:frequency data compiled by Lawes (1974). The following assumptions have to be made:

a. no soil erosion is caused by storms with an average intensity of $\geq$ 25 mm/hr, based on 30 minute duration;

b. kinetic energy can be predicted from rainfall intensity (Hudson, 1971);

c. there is a strong linear correlation between soil erosion and the kinetic energy of storms $>$ 25 mm/hr and the EI_{30} parameter (Moore, 1978).

If these assumptions can be accepted, calculation of the magnitude:frequency of kinetic energy $>$ 25 mm/hr and the EI_{30} for the meteorological stations should be good indicators of the magnitude: frequency of soil erosion. The above calculations for KE 25 and EI_{30} have been performed for 33 East African stations, each with a record of at least 8 years. The results are presented as the percentage of mean annual KE 25 or EI_{30}, which occur in 30 minute rainfall periods with a frequency of less than or equal to 5, 1, 0.5, 0.2 times per year. If there is a strong correlation between the rainfall and erosion parameters, these percentages should also be a good indication of the soil erosion occurring in these rainfall periods.

The results show that at the wetter stations (e.g. Equator, Entebbe, Gulu, Kisumu, Nairobi-Kabete and Tororo), about half the annual soil erosion may be caused by 30 minute rainfall periods which occur with a frequency greater than 5 times/year. High intensity rainfalls, with frequencies of less than once per year, do not cause a large proportion of the soil erosion.

% Property (KE > 25 or EI_{30}) in Storms of Frequency (no./yr) Less Than or Equal to That Indicated:

Station	KE > 25 frequency 5.0	1.0	0.5	0.2	EI_{30} frequency 5.0	1.0	0.5	0.2
Eldoret	73.3	39.5	14.8	4.9	83.1	55.2	24.2	8.9
Equator	40.1	8.0	8.0	0	49.3	13.6	13.6	0
Kisumu	60.3	18.5	6.1	6.1	77.9	32.1	13.5	13.5
Kitale	45.8	18.1	7.8	3.4	60.8	30.0	14.6	7.0
Lamu	62.4	20.1	20.1	0	73.9	33.5	33.5	0
Lodwar	100	69.5	30.9	30.9	100	81.0	46.2	46.2
Makindu	100	32.2	19.0	8.1	100	47.0	32.2	11.7
Malindi	100	23.4	10.7	5.0	100	37.3	19.9	10.7
Mombasa	57.2	18.6	9.3	5.9	69.8	31.6	18.7	13.2
Nbi Airport	100	27.6	27.6	16.1	100	42.5	42.5	28.0
Nbi Kabete	45.1	21.5	2.7	2.7	54.0	30.2	5.2	5.2
Nakuru	100	40.3	23.9	11.7	100	60.2	42.7	22.7
Nanyuki	52.5	28.8	17.3	8.9	65.1	43.2	29.5	17.3
Narok	100	29.1	17.5	8.0	100	43.5	30.1	14.4
Voi	100	25.7	24.2	9.8	100	48.2	39.6	18.6
Dar Es Salaam	73.2	29.2	3.9	3.9	83.2	44.4	7.9	7.9
Dodoma	100	31.9	16.0	10.5	100	48.4	29.5	20.5
Kigoma	100	23.5	5.7	5.7	100	36.8	11.6	11.6
Lyamungu	71.1	23.2	12.1	12.1	80.7	36.0	21.2	21.2
Mbeya	100	32.4	16.0	1.9	100	45.2	26.0	4.3
Mwanza	46.4	27.3	1.6	1.6	60.7	41.1	3.1	3.1
Tabora	73.0	27.5	8.7	1.3	82.8	42.4	16.8	3.3
Zanzibar	46.3	13.2	7.9	1.4	60.0	22.8	14.8	3.0
Entebbe	55.1	12.1	6.6	3.6	69.2	22.1	13.3	8.0
Fort Portal	52.4	8.0	3.9	1.5	66.7	14.6	7.8	3.4
Gulu	37.4	6.3	6.3	2.1	50.6	11.8	11.8	4.4
Jinja	59.5	17.4	1.0	1.0	75.2	30.8	2.1	2.1
Kabale	100	15.5	1.7	1.7	100	23.0	2.9	2.9
Kampala	50.4	15.6	7.8	2.5	64.6	26.5	15.1	5.5
Kasese	100	10.0	10.0	2.0	100	19.9	19.9	4.0
Masindi	54.6	12.5	6.7	1.0	69.4	22.6	12.3	2.3
Mbarara	100	22.0	2.8	2.8	100	32.3	5.0	5.0
Tororo	41.3	10.2	6.1	0	56.9	18.7	11.9	0

At the semi-arid and arid stations (e.g. Lodwar, Makindu, Nakuru, Voi and Dodoma) the pattern has changed, due partly to the smaller number of rainfall events and partly to the increased proportion of intense rainfalls. Almost half of the mean annual EI_{30} value for these stations is caused by 30 minute events with a frequency of less than once per year, and up to 20% by events occurring once in 5 years.

This rather crude analysis of the East African situation does suggest that the drier the climate, the more important the rare events.

As general conclusions, then, soil erosion by overland flow and rilling appears to be dominated by events which occur 5 to 10 times per year in humid areas, and once per year in arid or semi-arid climates. The denser the plant cover, the more important is the extreme event. Where gullying is important, events with return periods of 0.5 to 5 years are probably dominant, due to the cumulative effect of overland flow concentrating in gullies, and the rapid increase in erosive ability of gullies with increases in discharge. Landslide activities are also likely to be dominated by events with recurrence intervals of about 5 years.

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6. THE CONSEQUENCES OF ACCELERATED SOIL EROSION

In the previous sections, the mechanisms of and measurement of soil erosion have been examined. The important question still remains: what effects do accelerated rates of soil erosion have on man's activities? This can most conveniently be dealt with by separating these effects into those which affect the agricultural productivity of the soil, and those which cause downstream changes in rivers and streams.

Agricultural effects

Many examples can be found of the effect of topsoil removal on crop yields. In the United States, a reduction in topsoil thickness from 30 to 0 cm generally produced a reduction in yield of about 50%, for a wide range of crops such as cotton, corn, hay and grapes (Stallings, 1964). Beasley (1972) estimates that reduced productivity from eroded soils in the United States cost about $500 million per year, based on mid-1960's data. In Nigeria, Lal (1976) observed a decrease in about 50% for cowpeas and maize when the top 12 cm of an Alfisol were removed. Not all examples show a decrease in yield associated with topsoil removal; for example, Grosse (1967) noted an increase in productivity when the topsoil was removed from some Parabraunerde soils.

The overwhelming evidence, though, is for a significant decrease in productivity associated with topsoil removal or soil erosion. This decrease depends on a wide range of soil characteristics, and how they vary with depth in the soil. The important characteristics can be divided into two groups, chemical and physical.

Soil erosion affects the chemical properties of soils by selectively removing the finest particles, which often contain the majority of the nutrients available for plant growth. In particular, the eroded sediment often contains a higher proportion of silt, clay and organic matter than the original soil. One common method of expressing this difference is the erosion ratio:

$$\text{Erosion ratio} = \frac{\text{silt \& clay}}{\text{sand \& gravel}} \text{ in eroded sediment} \div \frac{\text{silt \& clay}}{\text{sand \& gravel}} \text{ in original soil.}$$

and the enrichment ratio:

$$\text{Enrichment ratio} = \frac{\text{\% available or total nutrient in eroded sediment}}{\text{\% available or total nutrient in original soil}}$$

Barrow and Kilmer (1963) have reviewed experiments in which the enrichment ratio has been measured, mainly from soils in the Corn Belt of the U.S.A. Available nutrient enrichment ratios are generally higher than that for total nutrients because most of the available nutrients are attached to the erodible particles. General enrichment ratios are:

1.2 to 4.7 for organic matter
1.2 to 5.0 for total nitrogen
1.3 to 3.1 for total phosphorus
3.3 to 6 for available phosphorus
1.1 to 7.3 for total potassium
4.7 to 12.6 for available potassium
1.4 to 2.4 for available calcium and magnesium.

The above figures apply mainly to temperate soils. Lal (1976) measured sediment losses from plots in Nigeria, and found that the overall erosion ratio was 2.3, though it was higher on 1% slopes and lower on 15% slopes, reflecting the ability of overland flow on steep slopes to transport coarser particles. The overall enrichment ratios were 2.4 for organic carbon

1.6 for total nitrogen
5.8 for available phosphorus
1.7 for exchangeable potassium,
1.5 for exchangeable calcium
1.2 for exchangeable magnesium.

The enrichment ratio decreased with increasing slope, as more and more sand was eroded.

Losses of nutrients can also occur in the dissolved form in the runoff water. This particularly applies to recently added fertilisers which may be dissolved by the surface water and pass across the soil surface in overland flow, thereby bypassing the "fixing" properties of the soil. This is especially true for phosphorus which is readily fixed by soil particles, but can be lost through surface runoff. In the Nigerian experiment, Lal (1976) also measured dissolved nutrients in the runoff. He found that, under bare fallow, the dissolved nutrient loss was generally higher than that as available forms on the eroded sediments, except for nitrogen, which was measured only as total nitrogen on the eroded sediments. Both dissolved and adsorbed nutrient losses were reduced when a crop cover was introduced, emphasising the role of vegetation in protecting the soil surface, reducing runoff and increasing nutrient uptake and cycling.

At Samaru, Northern Nigeria, Kowal (1969) measured soil erosion and nutrient losses under different management and crop conditions on a loamy sand soil, bench terraced at 0.4% slope. He found enrichment ratios of 1.8 to 4.8 for total nitrogen, 1.6 to 3.4 for exchangeable potassium, 0.8 to 1.6 for exchangeable calcium and 0.8 to 1.7 for exchangeable magnesium. Thus, significant losses of nitrogen and potassium occurred, and these were reflected in lower soil contents at the end of the experiments. Silt and clay were preferentially removed in erosion, with silt and clay ratios of eroded:original soil of 1.7 to 3.4.

Soil erosion affects the physical characteristics in a number of ways. Compaction of the soil surface by raindrops reduces infiltration rates and increases overland flow. Then, impermeable crusts may form (McIntyre, 1958 a,b) and these can inhibit seedling emergence. Lal (1976) records a decrease in infiltration rate from 35 to 2 mm/min on bare fallow plots on 10% slopes in Nigeria within a 3 year period.

Removal of the finest particles and compaction of the surface layer also reduce the water holding ability of the soil. In the Nigerian study (Lal, 1976), there was a decrease in water held at all soil moisture tensions, and a decrease in the moisture equivalent and available water capacity over a 2 year period on a bare, fallow plot.

Subsoils usually have a lower porosity, especially macroporosity, than the surface horizons, and removal of the topsoil often reduces the effective root volume available to a plant, as well as the rooting depth.

The overall effect of these physical changes, then, is to reduce the amount of water available for plant growth, and to make the crop more susceptible to drought.

It is difficult to assign productivity decreases on eroded topsoils to either nutrient or available water changes. Beasley (1972) quotes a study in Missouri in which the topsoil was removed from one plot, adequate fertilisers added and the yields of corn compared to an equivalent non-eroded, fertilised plot. Over an 18 year period the corn yield from the eroded plot was about half that from the non-eroded plot, suggesting that decreases in available moisture is the important factor. Obviously, where water is potentially limiting to crop growth, decreases in moisture availability will be most important; where the soil is depleted of nutrients, such as some highly leached, tropical soils, then nutrient losses will be more important.

In Kenya, decreases in water availability are probably more important, though nutrient losses may be important in highland areas. In the Kajiado District, Dunne (1977) has estimated that about 5 to 10% of the rainfall is converted into overland flow, and if all of this water was absorbed by the soil and available for transpiration by grasses, then productivity would be increased by 8 to 16%. The soils in the Kajiado area generally show few decreases in available nutrients with depth, so nutrient losses are small compared to water losses.

A similar type of calculation could be performed for maize growth in the semi-arid areas of Kenya. For example, in Machakos District, the yield of maize has been found to be strongly correlated with seasonal rainfall (Dowker, 1971):

$$y = 1514 + 120.06\ (R-27.64) \pm 610 \quad r^2 = 0.72$$

where y = yield in Kg/ha

R = seasonal rainfall (0-120 days) in cm.

Assuming 10% of the seasonal rainfall runs off as overland flow, conservation measures to ensure that all this water would be absorbed by the soil would result in a maize yield increase of 25% (assuming a mean seasonal rainfall of 28 cm). When a 5% increase in soil moisture occurs, a 14% increase in maize is produced. Thus, conservation measures to ensure all the rainfall is absorbed by the soil is a worthwhile practice in these semi-arid regions.

Downstream effects

The main downstream effects of soil erosion relate to the rapid rate and high volume of runoff and the increased sediment load. Most conservation measures reduce the amount of overland flow, though Morgan (1972) has observed overland flow occurring as a result of terracing in Malaya. On deep, permeable soils near Kuala Lumpur, most of the precipitation infiltrates and passes downslope as through flow; terraces intercept this water and transform it into overland flow.

The increased volume and speed of runoff from eroded areas produces a higher and more rapid discharge peak in the downstream areas, often leading to flooding. The reduced infiltration also causes lower base flows; thus the extreme discharges of a river system are extenuated.

The increased high discharges can cause severe erosion of the gully and stream systems, and scour the bed of the main channel. River banks can also be eroded as a response to the increase in peak discharges Beasley (1972) estimates floodplain and stream bank erosion to cost $47 million annually in the U.S.

The increased sediment concentrations and loads can also create problems. For many drainage basins the sediment delivery ratio is low (generally 0.05 to 0.80), that is not all the sediment removed by erosion reaches the river system. Gully erosion probably goes directly into the stream or river, but much of the rill or sheetwash erosion probably accumulates on lower, gentle foot slopes. If an eroding area moves sediment into a relatively flat river area, aggradation of the channel floor can occur, reducing the carrying capacity of the river, as well as depositing sediment on the flood plains. Braided channels are common examples, though not all braided channels are related to high sediment loads in gentle channel sections. Channel responses to changes in discharge and sediment are examined by Leopold _et al._, (1964) and Schumm (1969).

Increased turbidity of river water involves increasec costs in water treatment and purification. Eroded sediment often accumulates in reservoirs and behind dams, reducing the effective life and capacity of the dam. Storage depletion rates of 5% per year are common (Beasley, 1972). The rates are higher for small reservoirs than for large ones. Hudson (1971) gives two methods of estimating storage depletion or sediment trapping by reservoirs.

A final problem, mainly restricted to areas where agricultural practices are intensive, is pollution caused by erosion. The nutrient status of many waterways in Europe and North America has been changed by the removal of nutrients, particularly phosphorus, from fertilised fields through soil erosion and overland flow. The eutrophication of some lakes has been attributed to agricultural practices. Furthermore, normally terrestrially immobile particles such as DDT, can be transferred into water bodies adsorbed onto soil particles, thus bringing potential damage to aquatic ecosystems.

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7. THE PREDICTION OF SOIL EROSION

The accurate measurement of soil erosion is generally an expensive process, so that the construction of appropriate soil conservation measures is dependent on reliable estimates of soil erosion rates, under given environmental and management conditions. The need for such a predictive tool became apparent in the U.S.A. in the 1930's. Much of the research work of the Soil Conservation Service, established at that time, has been directed towards the identification of the important factors controlling soil erosion and their quantification in predictive equations. Soil loss has been measured from small plots under various climates, slopes, lengths, soils, crops and management practices. 10,000 plot-years of data have been collected, and these data form the basis of the universal soil loss equation.

Musgrave (1947) first established the five factors which control soil erosion:

(1) climate (R)

(2) soil erodibility (K)

(3) slope angle and length (LS)

(4) crop and management (C)

(5) conservation practices (P)

Quantification of the above factors was carried out by analysing the soil loss data for the experimental plots under controlled conditions. Modifications are still being made as more data become available (Soil Conservation Service, 1977) and a compendium of papers on the method and its application has recently been published (Soil Conservation Society of America, 1977). Basic information is still lacking for many tropical countries, so the benefits of the method are applicable mainly to North America, though evaluation of the factors has been performed for some other areas (e.g. Roose, 1975).

This empirical approach is useful, in that the effect of each factor can be quantified, allowing a modification in crops or management practices to adjust soil losses. The empricism involved means that the approach can only be used in areas where the fundamental relationships in the erosion process are the same as in the areas from which the equation was derived.

Wischmeier (1976) has recently re-iterated the main limitations of the approach. The equation predicts losses of soil moved off defined slope segments, not erosion within the segment (for example, redistribution by rainsplash). Nor can it predict the amount of soil which may accumulate on the footslope. The differences between soil loss and soil erosion are important. Gully erosion is not included in the equation, which deals with losses due to sheetwash and rilling.

The equation contains predictive errors. Wischmeier (1976) checked the mean annual soil loss of 189 plots against that predicted by the equation, and found that the average error was about 10% (1.4 tons/acre from a mean of 11.4 tons/acre). Over 84% of the plots were within 2 tons/acre of the predicted value. Although these figures are impressive, it must be realised that they have been obtained from carefully controlled, typical plots in the U.S.A., and errors will probably be much larger for areas, soils and crops not included in the original analysis. For African conditions, an accuracy of ±50% might be observed, unless the individual factors are carefully examined. Moreover, the errors increase when calculations are made on the basis of individual storms, in which the soil loss is strongly affected by antecedent conditions; the variations in antecedent conditions are balanced out on an annual basis.

There are two main ways in which the universal soil loss equation can be used. One way is to predict annual or long term soil losses from a field under given management and crop conditions. Differences in crops and management can then be tested for their effect on soil loss.

The second way is the reverse of the first. It is to set a limit to the amount of soil loss which can be tolerated, and then to adapt crops and conservation practices to specific fields to ensure that the soil loss is kept below the tolerable level. This approach requires the establishment of acceptable levels of soil loss or erosion, a topic which has been discussed by Smith and Stamey (1965) and Stamey, and Smith (1964). Soil losses can be regarded as tolerable when they do not detrimentally affect the long-term productivity of the soil. Whilst soil is being lost through erosion, it is simultaneously being gained by chemical and physical weathering of the parent material. Rates of weathering are infrequently measured, but may be between 0.2 and 0.5 tons/ha/yr. The tolerable rate of soil loss will also depend upon the depth and fertility of the soil: deep, fertile soils are more able to absorb soil losses than are shallow, infertile soils.

On the above basis, tolerable upper limits of soil loss have been established (Table 7.1).

Table 7.1 Tolerable soil losses (from Arnoldus, 1977)

Rooting Depth cm	Annual Soil Loss (tons/ha) Renewable Soil[1]	Tolerance Value Non-Renewable Soil[2]
0 - 25	2.2	2.2
25 - 50	4.5	2.2
50 - 100	6.7	4.5
100 - 150	9.0	6.7
150	11.2	11.2

[1]Renewable soil: has favourable substrata that can be renewed by tillage, fertilisers, organic matter and other management practices.

[2]Non-renewable soil: has unfavourable substrata, for example rock, or soft rock that cannot be renewed by economic means.

Whilst the equation is useful in predicting soil losses from fields, it does not examine the fate of the eroded soil once it leaves the field. In the U.S.A., much concern has developed over the fate of eroded soil which enters river channels or reservoirs, and causes pollution. The universal soil loss equation is not competent to accurately predict sediment yields from drainage basins (Beer et al., 1966; Wischmeier, 1976). Partly in response to this inadequacy, and partly to improve the predictive ability of soil erosion models by removing some of their empiricism, research in the U.S.A. has developed a more detailed model of soil erosion.

The starting point of this model is based upon Ellison's (1947) recognition of the two fundamental processes in soil erosion : particle detachment and particle transport, and the roles of rainsplash and overland flow in them. Meyer and Wischmeier (1969) have developed a model which incorporates these components to predict soil movement down a slope segment (Fig. 7.1).

The model requires functions for each of the processes involved, and many of these functions have been derived empirically from laboratory or field measurements (David and Beer, 1975; Foster and Meyer, 1972; Foster et al., 1972a, b; Onstad and Foster, 1975; Young and Mutchler, 1972). This approach, then, still involves some empiricism, but does allow a prediction of erosion and deposition on a slope, and thence an assessment of the eroded soil which reaches the river channel. The model is in the process of being calibrated for different watersheds, and some promising comparisons of predicted and observed sediment yields have been made, but much more work is required before the method will be as effective as the universal soil loss equation.

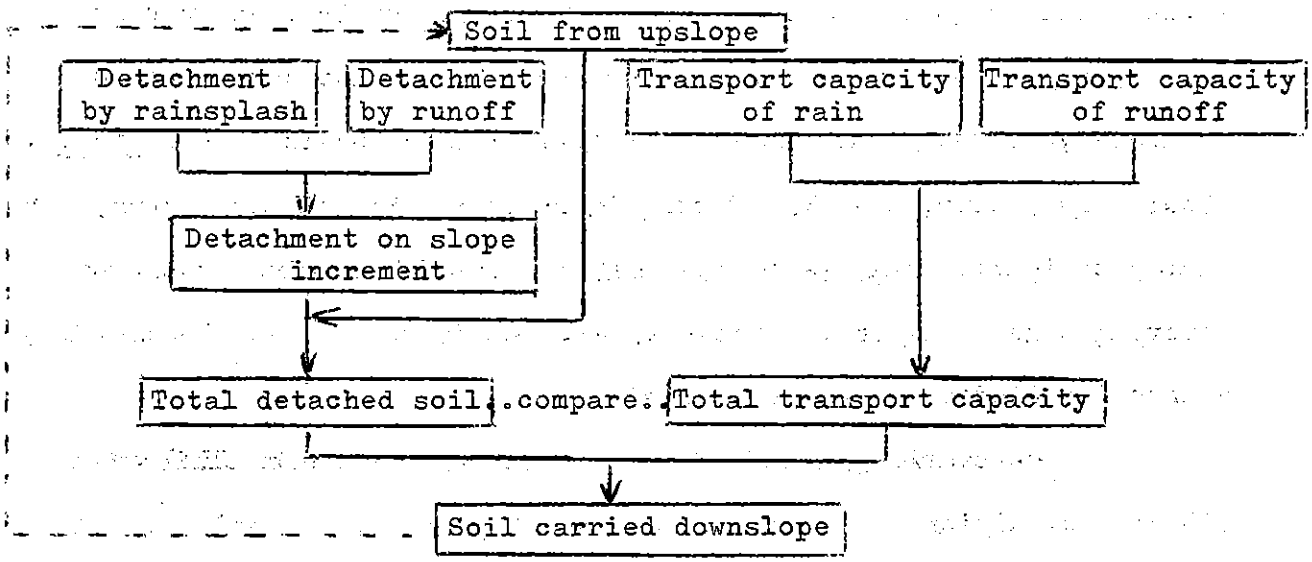

Fig. 7.1 A model of soil movement downslope.

By comparison to sheetwash and rill erosion, gully erosion prediction has been the subject of few studies. Beer and Johnson (1963) and Thompson (1964) examined gully area changes in the U.S.A., as a function of environmental variables such as slope of approach channel, gully depth and length, runoff characteristics, soils and position of the gully with respect to the watershed divide and main river channel. Seginer (1966) carried out a similar study in Israel, using gully head advancement as his dependent variable; air photographs were used to define the change in gully characteristics over a number of years. Logarithmic or linear regressions were found to be the most effective predictive form, and correlation coefficients of between 0.7 and 0.8 were obtained. Malde and Scott (1977) have monitored gully advancement in the southwestern U.S.A. and have found that advancement was closely related to runoff events, but that the relationship was complicated by variations in antecedent conditions.

Based on the above studies, a gully retreat equation has been proposed for areas in the U.S.A. which receive 500 mm of rain annually; it has not been tested in other areas (Soil Conservation Service, 1977b):

$$r = 1.5\, w^{0.46}\, p^{0.20}$$

where r = rate of retreat of gully head (ft/yr)

w = drainage basin area of gully (acres)

p = summation of rainfalls 0.5 in per 24 hr, for the lifetime of the gully (in/yr)

Finally, a predictive equation has also been developed for wind erosion in the U.S.A., again based on empirical relationships (Chepil, 1963; Skidmore et al., 1970; Woodruff and Lyles, 1965). No estimates of the accuracy are given. The equation is:

$$E = f\,(I.K.C.L.V.)$$

where E = soil loss by aeolian processes

I = soil erodibility, based on the proportion of the soil 0.84 mm on dry sieving

K = soil surface roughness

C = climatic factor, based on soil moisture and evapotranspiration and wind velocity

L = field length along prevailing wind direction

V = vegetation cover.

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8. RAINFALL EROSIVITY

Rainfall is the driving force behind most of the "accelerated" erosion rates, through its effects as raindrop impact and soilsplash and its creation of overland flow. Rainfall erosivity is the potential ability of rain to cause erosion, and obviously is an important component in the prediction and control of soil erosion.

Many studies have been conducted in the laboratory and field to assess the relationship between different rainfall parameters and rates of soil erosion, under controlled conditions. General parameters such as annual rainfall and the seasonality of rainfall have been tested, but appear to be applicable only at small scales, such as global or continental (Fournier, 1962; Jensen and Painter, 1974).

Laboratory studies of soil erosion have shown that the energy parameters of rainfall are as important as the total amount. In particular, the rainfall's kinetic energy and momentum have been found to be strongly correlated with rainsplash erosion (e.g. Free, 1960; Hudson, 1971; Rose, 1960). As these two parameters are closely related in nature, they are equally acceptable as predictors of splash erosion.

Rainfall intensity is another parameter which might be expected to be important in soil erosion, as intense rainfalls are more likely to create overland flow. The critical intensity in creating overland flow will obviously depend on characteristics such as soil type, infiltration rate, soil moisture content and vegetation cover. At Kericho, Othieno and Laycock (1977) found that little erosion occurred when the rainfall intensity is less than 20 mm/hr, and Hudson (1971) and Smith and Wischmeier (1962) regard rainfalls with an intensity <25 mm/hr as non-erosive. On the other hand, French workers in Africa regard the maximum rainfall intensity as an important parameter, such as >60 mm/hr for 15 minutes (e.g. Fournier, 1967; Roose, 1967).

The most detailed analysis of rainfall erosivity has been carried out in the U.S.A., where soil erosion from large fallow plots has been compared to various rainfall parameters over a number of years. For each storm, soil loss was correlated with rainfall parameters (Wischmeier and Smith, 1958). For five soils, correlation coefficients increased in the following general order:

> max. 15 minute intensity $<$ max. 30 minute intensity $<$ rainfall amount $<$ rainfall energy $<$ combined rainfall amount, max. 15 minute intensity and max. 30 minute intensity $<$ kinetic energy and max. 30 minute intensity $<$ kinetic energy and max. 30 minute intensity plus antecedent precipitation and total rainfall since last tillage.

The correlation coefficients (r) increased from about 0.65 to 0.95 within this sequence. On this basis, Wischmeier and Smith (1958) recommended that the product of the kinetic energy and max. 30 minute intensity (EI_{30}) be used as the most effective rainfall erosivity parameter. This parameter has been incorporated into the Universal Soil Loss Equation, for individual storms and for the whole year. The summed EI_{30} values are divided by 100 to become the rainfall erosivity parameter, R.

The EI_{30} values can be calculated by examination of hyetograms, which record rainfall amount against time, giving intensity. Thus, the max. 30 minute intensity can be measured for each storm that satisfies the criteria, i.e. a storm must have at least 12.5 mm of rain and be separated from other storms by a period of at least 6 hours with less than 1.3 mm of rain. The kinetic energy cannot be calculated but can be estimated, because several studies have shown that there is a reasonably good correlation between rainfall intensity and kinetic energy (Hudson, 1971). Although some errors will occur in the prediction of kinetic energy from rainfall intensity, these are probably small, especially when compared to other errors in the Universal Soil Loss Equation (Kinnell, 1973; Kowal and Kassam, 1976; McGregor and Mutchler, 1977; Rogers et al., 1967; Stocking and Elwell, 1976).

One factor which can increase clod breakdown, particle detachment, and erosion rates is wind speed. For example, Lyles et al. (1969, 1974) found that 66% more soil was lost when rain was accompanied by a 13 m/sec wind than when there was no wind. However, there appears to be little likelihood of being able to incorporate wind speed within routine estimates of rainfall erosivity.

Although studies have shown a high correlation between EI_{30} and soil loss on an annual basis in the U.S.A. (Wischmeier and Smith, 1962), the correlation is poorer when expressed on the basis of individual storms, and when applied to other areas. For example, Ahmad and Breckner (1974) found correlation coefficients of 0.4 to 0.6 for three soils in Tobago, and Lal (1976) obtained correlation coefficients of 0.7 to 0.9 for an Alfisol in Nigeria. Rhodesian studies have shown that, for vegetated plots, the EI_{30} parameter is less efficient than an EI parameter based on shorter periods of maximum rainfall intensity (e.g. EI_5 and EI_{15}) (Elwell and Stocking, 1973a,b; Stocking and Elwell, 1973). The rainfall erosivity parameters have been discussed in more detail by Moore (1978).

While rainfall erosivity values have been established for North America (Wischmeier and Smith, 1962), few studies have been made in Africa. Roose (1976) has prepared a map showing the distribution of the R parameter in West Africa, based on a strong correlation between mean annual rainfall and R. Values range from over 2000 in the very wet coastal areas to 100 in the Sahel. In Rhodesia, Stocking and Elwell (1976) found R values of over 300 in the high veld and eastern districts, decreasing to below 100 in drier areas.

The paucity of rainfall intensity measurements in East Africa hinder the calculation of R values. However, Moore (1978) has calculated the kinetic energy of rain falling at intensities greater than 25 mm/hr (KE > 25 parameter) for 35 stations in Kenya, Tanzania and Uganda, based on the data contained in Lawes (1974). For 15 minute periods, the kinetic energy values range from about 2000 $J/m^2/yr$ in semiarid northeastern Kenya to over 23000 $J/m^2/yr$ in parts of Uganda. A map shows the distribution of the KE > 25 parameter, based on correlations between KE > 25 and mean annual rainfall, though the accuracy is not very good (average standard errors of about 2000 $J/m^2/yr$).

Although this parameter shows the general pattern of rainfall erosivity in East Africa, it cannot be used in the Universal Soil Loss Equation. Wenner (1977) has calculated R values for 11 Kenyan stations, and there is a strong correlation between these values and the KE > 25 parameter ($r = 0.952$). Thus, the KE 25 values can be converted into R values for the 35 stations, based on the assumption that the same R:KE > 25 relationship applies to the Tanzania, Uganda and remaining Kenya stations. These calculations have been performed (Moore, 1978) and give the following results (Table 8.1).

The estimates show that R values range from less than 100 in semi-arid Kenya, 100 to 200 in much of central Kenya and Tanzania, to 200 to 400 in the Kenyan and Tanzanian highlands and over 400 in much of Uganda. Local variations in topography will obviously affect the R values.

Another important characteristic of rainfall erosivity is its temporal distribution, particularly with reference to the development of a protective crop cover. In East Africa, nearly all the annual R value is concentrated in the rainy season(s). Moore (1978) has analysed the distribution of the most erosive rains,

Table 8.1 Estimated Rainfall Erosivity, R, Values For East African Meteorological Stations.

Station	R	Station	R
Eldoret	258	Dar es Salaam	293
Equator	210	Dodoma	146
Kisumu	476	Kigoma	287
Kitale	316	Lyamungu	227
Lamu	178	Mbeya	147
Lodwar	40	Mwanza	282
Makindu	103	Tabora	252
Malindi	167	Zanzibar	451
Mombasa	248	Entebbe	493
Nairobi Airport	110	Fort Portal	503
Nairobi Kabete	225	Gulu	546
Nairobi Wilson	171	Jinja	467
Nairobi Dagoretti	157	Kabale	197
Nakuru	119	Kampala	340
Nanyuki	163	Kasese	214
Narok	146	Masindi	487
Voi	134	Mbarara	235
		Tororo	646

taken as daily rainfall totals of > 25 or 50 mm, and compared to the onset of the rains, which also follows Fisher's (1977) criterion of starting when the 4-day rainfall total exceeds 50mm. The timing of the erosive rains is most likely to be critical in the lower rainfall areas, where planting and germination cannot start until the dry season is completed. In the wetter areas, planting can start earlier and the vegetation cover can be more protective by the start of the rainy season.

For stations with a pronounced long, dry season(s) the timing of the > 25 and > 50mm daily rainfall totals has been calculated (Moore, 1978). These stations are Kitale, Makindu, Mombasa, Nairobi Airport, Nakuru, Narok, Voi, Dar es Salaam, and Dodoma. In general, about 70% of the > 25 or > 50 mm daily rainfalls occur within the first 30 days after the onset of the rains. This 30 day period would be likely to have a low plant cover, for both arable areas and grazing land (Fisher, 1977; Stocking and Elwell, 1976).

Thus, although the R values for the low rainfall stations in East Africa are low, their effectiveness is increased because most of the erosive rains fall when the soil is exposed.

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9. SOIL ERODIBILITY

The erodibility of a soil is its susceptibility to erosion. The erodibility is dependent on the detachment and transport of soil particles by rainsplash and overland flow, and as such is related to physical and chemical properties and infiltration rate of the soil

Two commonly adopted approaches have been used to estimate soil erodibility. One approach is to measure erosion from field plots under carefully controlled conditions, using either natural or simulated rainstorms. Comparisons of the amount of soil eroded allows a ranking of soil erodibility, and when other factors (rainfall erosivity, slope length and angle, crop type, etc.) can be quantified, the Universal Soil Loss Equation soil erodibility 'K' factor can be calculated.

The second approach is to measure erodibility under controlled field or laboratory conditions, and then to identify important soil characteristics which appear to affect the erodibility of the soils examined. From studies on a number of soils and for different properties, simple or multiple regression equations can be developed to readily predict erodibility based on a few properties. This approach removes the necessity for actual measurements of erodibilty of all soils.

The first approach suffers from the problem that each erodibility value applies only to the particular soil, and cannot be readily transferred to other, different soils. The second approach can involve detailed laboratory analyses, and the best regression analyses are still not perfect in their predictive abilities.

If soils existed as individual particles of sand, silt and clay, the assessment of erodibility would be relatively simple. Mazurak and Mosher (1968, 1970) and Farmer (1973) have shown that the detachment of soil particles by raindrop impact is generally highest in the fine sand fraction. In real soils, however, the problem is complicated by the formation of aggregates which may resist detachment because of their increased size, or which may be broken down into more detachable sizes by raindrop impact. Furthermore, aggregation controls, to a large extent, the infiltration and permeability rates, which, in turn, affect the rate of overland flow. Thus, the number, size and stability of aggregates are an important control of soil erodibility. For example, Moldenhauer (1970), Moldenhauer and Kemper (1969) and Moldenhauer and Koswara (1968) found that large clods reduced rainsplash erosion and increased infiltration rates, mainly through a delay until the clods were broken down by raindrop impact into detachable sizes.

Much attention has been directed towards the measurement of soil aggregates and their stability (e.g. Deshpande et al., 1968; Greenland, 1977; Hamblin and Greenland, 1977). Although the methods employed are very empirical, organic matter and active iron and aluminium oxides and hydroxides appear to be the major agents in stabilising soil aggregates.

Bryan (1968) has reviewed the early studies which searched for properties which explained variations in soil erodibility. Many of these early studies concentrated on the particle size distribution of the soil, and the ease with which the soil could be dispersed. For example, a "dispersion ratio" was proposed based on the ratio of silt and clay contents in the undispersed and dispersed states. The "clay ratio" expressed the ratio of sand to silt plus clay. The proportion of water stable aggregates above certain sizes was measured, as were sesquioxide:silica ratios and type of exchangeable cations. Each of these parameters was found to be useful in distinguishing between "erodible" and "non-erodible" soils under local conditions, but often the criteria were of limited applicability to soils in other areas. This was due, in part, to the arbitrary nature of some of the analyses and their poor reproducibility and, in part, to the complex nature of soil erodibility and differences in the importance of each property between different soils.

Bryan (1968) applied the major erodibility indices to rainsplash erosion from 88 soils, under laboratory conditions. Correlation coefficients between erodibility and indices ranged between 0.1 and 0.8, with the best results being obtained for the surface aggregation ratio and the proportion of water-stable aggregates > 0.5 mm or > 3 mm diameter.

Pereira (1955) also examined various soil properties in relation to two East African soils, a sandy loam from Uganda and a Kikuyu red loam from Kenya, with differences in erodibility being related to different cultivation practices. He found that no one property was very satisfactory in explaining differences in erodibility, based on field observations, and this he ascribed to operator variability in analyses and the insensitivity of the methods to small changes in soil properties significant enough to increase erosion. Ahn (1978) considers micro-aggregation of silt and clay particles to be important, and to impart a "pseudo-sand" texture on the soil.

Bruce-Okine and Lal (1975) have developed a simple method for assessing soil erodibility. Analyses of water-stable aggregates do not incorporate the influence of raindrop impact, so their method measures the number of standard drops of water required to destroy air-dry clods, the results being expressed as the kinetic energy possessed by the raindrops. For Nigerian soils, they found a sand: clay ratio of 0.5 critical; at ratios > 0.5 fewer drops were required to cause destruction of the clods.

All of the above methods are very empirical, and standardisation of analytical conditions is critical for comparisons to be made. Recent research has given more attention to the basic properties of water and soil which affect soil erodibility. For example, North

(1976) has developed a method of measuring aggregate strength by subjecting aggregates to ultrasonic energy. The breakdown in aggregates is measured by determining the suspended clay content, a peak is reached, beyond which additions of energy do not cause further aggregate breakdown. Cruse and Larson (1977) and Sloneker _et al._, (1976) have examined the influence of shear strength and pore water pressure of soils on detachment, and Ghadiri and Payne (1977) have studied the relationship between raindrop stresses and crumb disintegration. At the moment, these studies have a very limited value from the practical viewpoint, but more theoretical studies can be expected to improve our knowledge and understanding of soil erodibility.

Several recent studies have developed regression equations to predict 'K' values from soil parameters. Barnett and Rogers (1966) examined soils in southeastern U.S.A., Dangler and El-Swaify (1976), El-Swaify and Dangler (1977) and Yamamoto and Anderson (1967, 1973) studied Hawaiian soils, Mannering and Wischmeier (1969) analysed 55 soils from U.S.A., and Young and Mutchler (1977) performed experiments on 13 Minnesota soils. Numerous physical, chemical and mineralogical analyses were carried out and used in simple and multiple regressions against measured 'K' values. Individual soil properties rarely possessed correlation coefficients greater than 0.7, but multiple regressions involving 6 or 7 parameters can often explain up to 95% of the variation in 'K'. Although this approach has produced impressive results, it remains to be seen whether a predictive equation can be developed which is applicable to all soils, and which involves easily measured parameters.

From earlier work, Wischmeier _et al._ (1971) identified 5 soil parameters which are most important in controlling soil erodibility. These are per cent silt plus fine sand, per cent coarse sand, organic matter content, structure and permeability. These authors have converted their regression equations into a nomograph which allows a rapid estimate of a soil's 'K' value, given the 5 parameters. The predicted 'K' value should be within ± 0.04 of the observed value for 95% of the samples, but this nomograph was developed mainly from soils in the central U.S.A., and the accuracy will probably be lower for soils of other areas.

Three other soil characteristics can influence the erodibility, and these can be difficult to evaluate. A stone or gravel layer on the soil surface can reduce erosion by increasing infiltration and reducing rainsplash detachment; a stone or gravel layer may reduce the 'K' value by 0.05 to 0.15 (Epstein et al., 1966). A second characteristic is crusting or capping, which occurs when surface aggregates are broken down by raindrop impact and a thin, impermeable surface layer is formed (McIntyre, 1958a,b; Tackett and Pearson, 1965). This layer reduces infiltration and increases overland flow, thereby increasing erosion, though turbulent overland flow may disrupt the layer and cause a temporary decrease in erosion. The mechanisms of the process of capping or crusting are poorly understood, but the phenomenon appears to be most evident in soils with a weak structure and containing significant amounts of fine sand, silt and clay.

The third factor influencing erodibility is the moisture content of the soil. This influences erosion through changes in rainsplash effectiveness and the duration of the period before overland flow commences during a storm. The 'K' factor value increases several-fold from a dry to a wet storm, under the same rainfall conditions (Dangler and El-Swaify, 1976).

Soil erodibility can be decreased by the addition of mulches or conditioners. Numerous studies of the effect of conditioners on soil structure and erodibility have been performed (e.g. Gabriels et al., 1977; Pla, 1977; Soil Science Society of America, 1975). Although some organic conditioners (e.g. polyacrylamide) or bitumen emulsion can be effective, the large quantities required restricts their application to very sensitive, eroding areas which warrant the expensive treatments.

Few measurements of the erodibility factor 'K' have been made for tropical soils. In Hawaii, El-Swaify and Dangler (1977) produced the following values:

Torrox clay loams	0.15 - 0.22
Humult silty clays	0.00 - 0.09
Ustert clay	0.30
Andept clay loams	0.07 - 0.17
Orthid and Andept fine sandy loams	0.35 - 0.55.

In southeastern U.S.A., clayey and fine loamy Paleudults and Hapludults possess 'K' values of between 0.17 and 0.37, a clay Humult of 0.02 and a Chromudert of 0.32 (Barnett, 1977). In West Africa, Roose (1977a) assigns 'K' values of 0.05 to 0.18 for ferrallitic soils and 0.20 to 0.30 for ferruginous tropical soils developed from granite. Roose (1977b) also observes that the nomograph gave satisfactory results, except where the soil possessed a rocky or gravelly surface layer.

In Kenya, Barber _et al_, (1978) have measured 'K' values of 0.05 for a Nitosol (Kabete) and 0.60 for a Luvisol (Machakos). From data given by Dunne (1977) for the Kajiado District, 'K' values of 0.14 for a Vertisol, 0.35 for a clay Rendzina and 0.50 for a sandy clay loam can be calculated.

In the absence of the required analytical and morphological information, approximate 'K' values can be derived from texture classes (Holzhey and Mausbach, 1977):

sandy	0.10 - 0.20
loamy	0.20 - 0.40
silty	0.30 - 0.45
clayey	0.25 - 0.40.

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10. SLOPE ANGLE AND LENGTH AND EROSION-CONTROL PRACTICES

Whereas the rainfall erosivity and soil erodibility factors can be modified in only a small way by the farmer, the angle and length of slope of the field are more readily alterred. Through changes in the slope angle and length and erosion-control practices, soil losses can be reduced to acceptable levels.

The angle of the slope affects soil movement through its influence on downslope rainsplash of particles and the velocity of overland flow. The length of slope affects the discharge of water by overland flow at the bottom of the slope, and, therefore, longer slopes may be expected to have a higher transporting ability.

Zingg (1940) appears to be the first worker to evaluate the influence of slope angle and length on soil loss, and produced the following expression, based on field and laboratory experiments:

$$X = c\ S^{1.4}\ L^{1.6}$$

where X = soil loss
c = constant
S = angle of slope (%)
L = length of slope

Wischmeier *et al.* (1958) separated the angle and length components, and found that the influence of angle could be best represented by a quadratic expression:

$$E = 0.43 + 0.30S + 0.04S^2$$

where E = soil loss
S = slope (%)

Studies of slope angle and soil loss in the Tropics, however, have suggested that soil loss is a function of the square of the slope angle (%), or even a higher exponent (Hudson, 1971; Roose, 1977). Lal (1976), working with plots of 1, 5, 10, and 15% slopes in Nigeria, observed exponents of 0.5 to 1.2 for soil loss and plot angle, though poor correlation coefficients between the two variables were obtained for plots with a crop or a no-tillage treatment. The exponents were generally lowest for the series of plots with mulch or crop treatments, as might be expected if the cover were to slow down overland flow and reduce rainsplash. Roose (1977) also notes that variations in slope angle of only 0.5% can cause considerable variations in soil loss in Senegal.

Zingg's (1940) original study showed that soil loss was related to the 0.6 power of slope length, and an exponent of 0.5 ±0.1 was incorporated into the early versions of the Universal Soil Loss Equation (Wischmeier *et al.*, 1958). The slope length factor's effect seems to be very variable, depending on characteristics such as soil type and vegetation cover. Lal (1976) found a variable but general increase in soil loss with an increase in slope length. Hudson (1971) notes that, under Rhodesian conditions, the exponent is higher than 0.6, and Roose (1977) quotes a study from Benin, West Africa, which indicates that the influence of slope length on soil loss is neither constant nor important.

Part of this problem lies in the relative importance of sheetwash, rill and rainsplash erosion on a plot. Foster *et al.* (1977a,b) suggest that the slope length exponent varies between 0 and 1, for slopes in which inter-rill erosion is dominant to ones in which rill erosion dominates.

In the Universal Soil Loss Equation, the length and angle factors are combined into one 'LS' factor (Soil Conservation Service, 1977):

$$LS = \left(\frac{\lambda}{72.6}\right)^m \frac{(430X^2 + 30X + 0.43)}{6.574}$$

where λ = field slope length (ft)
m = 0.5 if S 5%
  = 0.4 if S = 4%
  = 0.3 if S 3%
$X = \sin\theta$ where θ = angle of slope (degrees)

Graphs have been produced to allow easy calculation of the LS factor. A 9% slope, 72.6 ft long gives a LS value of 1.

All of the above experiments and calculations have been conducted for plots with a uniform slope angle. The question of changes in soil loss when plots are not uniform in slope angle has been examined by Young and Mutchler (1969a,b). They used a rainfall simulator on 75 ft long slopes with an average angle of 8½%, but with either a uniform, concave or convex shape, so that angles varied between 2 and 14%. Runoff increased slightly from the concave to uniform to convex slopes, but the increase was not significant at the 5% level. There was a pronounced change in soil loss, however, from the uniform to concave slopes of about 35% decrease (significant at 1% level). There was an increase of only about 5% from the uniform to convex slopes (not significant at 5% level), possibly due to poor detachment of soil particles on the upper, flatter sections of the convex slope. Soil movements occurred within the plots, with the concave slope tending to become more concave, the uniform retaining its uniform nature, and the convex slope's point of maximum inflection moved upslope. For points on the slopes, the gradient of the upslope 5 m was found to be most strongly correlated with soil loss, though the correlation is not very good ($r^2 = 0.43$).

D'Souza and Morgan (1976) have also examined the effect of slope steepness and curvature on soil erosion.

As a general conclusion, then, soil losses should be less than expected on concave slopes and slightly higher than expected on convex slopes. Positions of maximum erosion on the slopes will also differ.

Foster and Wischmeier (1974) have recently proposed a rather complicated method for evaluating the LS factor for irregular slopes, and the approach has been reviewed by Arnoldus (1977). The method involves splitting the slope into equal length segments, or using a nomograph.

As noted by Roose (1977) the slope length and angle factor

(LS) is the weakest link in the Universal Soil Loss Equation, and for accurate predictions of soil loss, adjustments have to be made to allow for differences in the relative effectiveness of different types of erosion, as affected by soil type, climate, vegetation cover, etc.

Man can change the LS factor by modifying the slope angle and/or length. Contour ploughing and strip cropping or grass stripping can effectively reduce the length of slope by causing depression storage or by slowing down overland flow and allowing water to infiltrate. Similarly, bench terracing reduces both the length and angle of the slope.

The Universal Soil Loss Equation incorporates these changes in its erosion-control practice factor (P). Based primarily on experience in the United States, recommended values are given in Table 10.1 (Arnoldus, 1977; Soil Conservation Service, 1977). Downslope ploughing is assigned a value of 1.00.

Table 10.1 'P' Factors For Contouring, Contour Stripcropping, and Terracing.

Land Slope %	'P' Values contouring	contour strip cropping	terracing[1] a	terracing[1] b
1.1 to 2	0.60	0.30		
2.1 to 7	0.50	0.25	0.10	0.50
7.1 to 12	0.60	0.30	0.12	0.60
12.1 to 18	0.80	0.40	0.16	0.80
18.1 to 24	0.90	0.45	0.18	0.90
24.1	1.00	0.50		

[1] terracing which involves reduction in slope length only, i.e. channel terracing; a refers to off-field sediment load, whereas b refers to the soil loss from the slope, much of which will accumulate in the channel.

The values in the above table should be regarded as only approximate. For bench terracing, or for normal terracing, an alternative approach to the 'P' factor is to calculate individual length and slope angle values. The values in the table indicate that contour ploughing becomes ineffective at slopes $>$ 12%, but strip cropping and contour ploughing combined can reduce soil losses on slopes of up to 24%. Terracing through the introduction of channels on slopes is not very effective, unless closely spaced, though the terrace channels do accumulate eroded material and reduce sediment yields.

There are few published evaluations of other conservation practices. Roose (1977) has studied the effect of grass strips in West Africa; the strips are 2-4 m wide and set in fields 20 to 50 m wide cultivated with crops. A 'P' factor of 0.10 to 0.30 is recommended for this practice, based on the ability of the grass strips to absorb water and entrap eroded soil. Tied-ridging is estimated to give a 'P' factor of 0.10 to 0.20. Roose (1977) recommends the adoption of cultural practices (mulch, rotation, early planting, etc) rather than mechanical practices in humid tropical areas.

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11. CROPS AND THEIR MANAGEMENT

The final factor affecting soil erosion is the type and amount of vegetation cover. The type of crop which is grown on a soil is obviously dependent on numerous factors such as climate, soil fertility, economics and social/cultural attitudes. In addition, crops should also be grown in such a way that they help maintain soil fertility and keep soil erosion down to acceptable levels.

The crop-management factor is the most easily modified of the five factors which control soil erosion, and thus represents the easiest solution to soil erosion problems. This is emphasized by the fact that most mechanical control measures (such as terraces, ridges or contour ploughing) can reduce erosion losses from arable fields by about half (see chapter 10), whereas changes in the type of crop or its management can reduce soil erosion by a tenth.

The vegetation cover affects soil erosion in a number of ways:

(1) a dense cover absorbs most of the kinetic energy of raindrops, and the rain that reaches the soil as canopy drip or stemflow does not cause much detachment of soil particles or structural deterioration. There are exceptions, however. In tropical rain-forests, raindrop size and velocity may increase after canopy absorption of the rainfall, and this can cause much soil detachment, as well as localized overland flow resulting from stemflow (Ruxton, 1967).

(2) a dense, permanent cover can keep infiltration rates high, so reducing overland flow. There are numerous examples of the increase in infiltration rates through a grass ley, and the decrease upon the reversion to an arable crop. The beneficial effects of a grass cover on infiltration can be pronounced in tropical areas, but they are often short-lived, with low infiltration rates being recorded after only one or two years of an arable crop (e.g. Pereira _et al._,

1954, 1967; Wilkinson, 1975a; Wilkinson and Aina, 1976).

(3) the presence of a permanent cover or the return of crop residues to the soil can result in an improvement in soil structure and stability. The beneficial effect appears to be less pronounced and more short-lived in tropical than in temperate soils, possibly because of increased rates of organic matter decomposition and weaker structural bonds in the former.

(4) a layer of litter or trash on the soil surface affords protection against raindrop impact and detachment.

(5) the presence of litter or vegetation retards the velocity of overland flow, and thereby decreases soil erosion. This may not always be the case, as De Ploey et al. (1976) have observed increased erosion rates from grass covered, as opposed to bare, laboratory plots at angles $> 8\frac{1}{2}^{\circ}$. This appears to be related to the creation of turbulent eddies on the downslope side of grass blades.

Thus, a vegetation cover generally reduces erosion losses compared to a bare soil. The magnitude of this reduction depends on factors such as soil type, climate, type and density of vegetation cover, preceding management practices and the types of soil erosion involved.

Much work has been carried out in the U.S.A. to quantify the crop-management factor for use in the Universal Soil Loss Equation. Wischmeier (1960) has developed crop-management factor values for 100 different crop rotations. The method is based on the recognition of five crop-period stages:

F rough fallow
1 seed bed
2 establishment
3 growing crop
4 residue or stubble

A crop-management factor value is assigned to each of these stages, expressed as a percentage of the soil loss compared to a bare fallow soil. The annual or seasonal crop-management factor ('C') is calculated by assessing the percentage of the rainfall erosivity (R) which falls within each stage, multiplying the crop and the R factors for each stage, and summing them.

As an example, Wenner (1977) has estimated the crop-management factor for a continuous maize crop near Nairobi:

Crop Stage	Dates	Soil Loss Ratio (%)	Annual R (%)	C-Factor
1	15/03 - 15/04	92	14	0.13
2	15/04 - 15/05	80	27	0.22
3	15/05 - 1/08	50	17	0.09
4	1/08 - 1/09	85	2	0.02
F	1/09 - 1/11	85	5	0.04
1	1/11 - 1/12	92	13	0.12
2	1/12 - 1/01	80	8	0.06
3	1/01 - 1/03	50	12	0.06
4	1/03 - 15/03	85	2	0.02
				0.76 Total

For this example, the annual C factor value would be 0.76, whereas a bare fallow would be assigned a value of 1.00, i.e. the maize crop reduces the soil loss by about 25%. The protection the maize affords the soil varies from 50% (crop stage 3) to 92% (crop stage 1).

Although the crop-management factors have been established for North American conditions, many of the crops and their management are inapplicable to tropical conditions. Table 11.1 gives some C factor values that have been established for tropical crops, based mainly on experiments in West Africa and Hawaii. The values range considerably for each crop, depending on the precise crop type, management practice and rainfall erosivity pattern involved. These data, combined with the more appropriate values given by Wischmeier (1960), should enable an initial attempt to apply the Universal Soil Loss Equation to different East African crops, though only approximate estimates will be produced.

Given the wide variety of crops, management practices and rainfall erosivities present in tropical areas, it will be a considerable time before reliable, local C factor values are available. Elwell and Stocking (1976) have proposed an alternative method, based

Table 11.1 Some Characteristic Crop-Management Factor ('C') Values for Tropical Crops (from Brooks, 1977; Lal, 1976; and Roose, 1977)

Crop Type	C Value
forest, dense shrub	0.001
good grassland	0.01
poor grassland	0.1
corn, sorghum, millet	0.4 - 0.9
fertilized, dense rice	0.1 - 0.2
cotton, tobacco (second cycle)	0.5 - 0.7
peanuts	0.4 - 0.8
cassava, yam	0.2 - 0.8
palms, cocoa, coffee	0.1 - 0.3
pineapples: surface mulch	0.01
residue buried	0.1 - 0.3
residue burnt	0.2 - 0.5
-	0.17 - 0.31
irrigated sugar cane	0.29 (planted before rains)
	0.11 (ratoon)
maize - maize, mulched	0.04 - 0.05
maize - maize	0.29 - 0.35
maize - cowpea, no tillage	0.05 - 0.10
cowpea - maize	0.16 - 0.40

on the cover density (percentage of the soil covered) of crops. They recommend that crop protection classes be established, based on similarities in cover density and plant morphology, and those classes could be substituted for the crop-management factor. The method is simple and cheap and warrants further investigation (e.g. Wilkinson, 1975b).

Although quantitative evaluations of crop-management factors are lacking in tropical areas, there are numerous studies of the influence of different crops, management practices and rotations on soil losses from experimental plots. The recent Conference on Soil Conservation and Management in the Humid Tropics produced reviews of work in East and West Africa (Ahn, 1977; Okigbo, 1977), South Asia (Lal, 1977b; Panabokke, 1977), Latin America (Lal, 1977c) and the Caribbean (Ahmad, 1977). Jones and Wild (1975) also examine soil erosion in the West African savanna, and Lal (1974) has reviewed the influence of shifting cultivation on soil erosion.

Temple (1972) has collated data from old experimental plots in Tanzania. These involved crops such as millet, sorghum, coffee, bananas, maize and grass, with different management practices such as clean weeding, mulching, grass strips and ridges.

The tea experiment at Kericho appears to be the only modern Kenyan evaluation of the influence of management practices on soil erosion (Othieno, 1975; Othieno and Laycock, 1977). Tea fields treated with herbicide or hand weeded produced high (up to 160 tons/ha/yr) soil losses for the first two years, but in the third year dropped to less than 5 tons/ha/yr when the tea crop had established a 60 to 70% cover. Inter-cropping with oats or placement of a lovegrass mulch kept soil losses less than 5 tons/ha/yr in each of the three years.

Most of the attention of crop-management influences has been directed towards arable crops. Recently, more attention has been paid to grasslands and woodlands, where overgrazing and overstocking can cause major soil erosion problems. The Soil Conservation Service (1977) has proposed various C values for grasslands and woodlands, ranging from 0.01 with 95-100% ground cover to 0.2 to 0.4 with 0% ground cover and variable litter and bush canopy. Arnoldus (1977) reproduces the table.

Elwell and Stocking (1975, 1976) have examined soil erosion from various grazing lands in Rhodesia, and express the crop factor as percentage ground cover. They found that there was little reduction in erosion from plots as the ground cover rose to above 30%, but erosion losses were large on plots with a sparser vegetation cover.

Dunne (1977) studied erosion from small rangeland plots under simulated rainfall in Kajiado District, Kenya. He found that erosion losses deceased as the vegetation cover increased, and that a basal cover value of 20-30% may be regarded as critical in preventing excessive erosion. The C factor values presented by the Soil Conservation Service (1977) also suggest that relatively little further reduction in erosion occurs when the ground cover increases above

20 to 40%. Dunne (1977) also noted that overstocking can increase soil erosion by destroying vegetation, compacting the soil surface and degrading the soil structure. He speculates on the impact of different stocking densities on vegetation cover and soil erosion.

Changes in the crop-management factor can be used to reduce soil erosion losses from agricultural fields. Some of the ways in which this might be acheived are examined below.

(1) Changing the crop type can reduce erosion, but this is a rather drastic change which may involve agronomic, social and economic aspects of more importance than soil conservation.

(2) Increases in crop density reduce soil erosion. Hudson (1971) quotes an example of this approach from Rhodesia. Increases in maize densities from 25000 to 37000 plants/ha, plus more fertilizer and ploughing-in of residues, results in a doubling of the maize yield and a reduction in soil erosion from 12.3 to 0.7 ton/ha.

Problems with this approach are that increased plant densities are only feasible where there are adequate supplies of water and nutrients. Where population pressures are high on infertile soils or in dry years, close spacing may result in much reduced yields. It is interesting to note that Hudson's (1971) example was the response in a wet year, with over 1000 mm of rain in the season.

(3) The most sensitive period for soil erosion in arable crops is during seedbed preparation and in the early stages of growth. In East Africa this is compounded by the fact that much of the erosive rain falls during this period. The erosion hazard can be reduced by ensuring that the crop be planted as soon as possible, though this may be difficult in areas where the rainfall is erratic and the soil too hard to till prior to the first rainstorm.

(4) Rotations and grass leys have been introduced in North America, and their beneficial effect on the C factor for succeeding arable crops is shown by Wischmeier (1960). For example, the C factor

for crop stage periods F and 1 for the first year of maize after meadow are about half those in continuous maize. However, the second year values are 80 to 90% of the continuous maize values.

This points out a major failing of leys, particularly in tropical areas. Improvements in infiltration and soil structure, and hence decreases in soil erodibility, are often slow to develop, yet the beneficial effects last for only one or two years afterwards. This has been shown for East Africa by Pereira et al (1954, 1967) and for West Africa by Wilkinson (1975a) and Wilkinson and Aina (1976). Increased rates of organic matter decomposition and higher rainfall erosivities may be the causes of this rapid deterioration.

(5) Inter-cropping may improve the cover by introducing a crop which matures more rapidly and offers a more effective protection early in the growing season. For example, Fisher (1977) has shown that beans establish a ground cover more rapidly than maize, and could be suitable as an inter-row crop. Alternatively, prostrate plants such as sweet potato or legumes could also be used. Othieno (1975) found that oats between tea rows was very effective in reducing soil erosion whilst the tea cover was being established. Lal (1977a) also notes reduced erosion from maize-cassava mixed cropping, compared to pure stands.

Problems with this approach are that establishment of the crops may be difficult in drier areas and that competition between the crops could significantly reduce overall yields. Fisher (1975) found at Kabete that maize-beans mixed cropping can give higher yields than pure stands in good rainy seasons, but can give lower yields in poor rainy seasons. Competition for moisture and nutrients is the major limiting factor, so the technique may be applicable to the more humid areas of Kenya, but ineffective in the drier areas.

(6) Numerous studies have shown that the application of a mulch or leaving the previous crop's residue on the surface can reduce

soil erosion. Othieno (1975) has shown this for tea mulched with lovegrass, and Temple (1972) has given examples from Tanzania. Application of a rice straw mulch at 2 tons/ha reduced erosion losses to negligible values from an Alfisol on gentle slopes in Nigeria; higher application rates are required on steeper slopes. Roose (1975, 1977) gives further examples from West Africa, and there are a large number of American studies (e.g. Lattanzi et al, 1974; Meyer et al, 1970).

Mulch reduces soil erosion by decreasing the soil detachment process by raindrop impact, by decreasing surface sealing, by increasing depression storage, by slowing overland flow and by improving the soil structure through biological activities (Lal, 1976).

Problems with this approach lie in the cost of establishing a mulch. Roose (1975) quotes costs of $150-250/ha for heavy mulches in Ivory Coast. The maximum amount of plant material should be left on the soil surface from the previous crop, but this can interfere with harvesting techniques (e.g. sugar cane). Adequate amounts of organic matter must be available, but this may not be feasible in drier areas where grazing and land pressures are high. Furthermore, crop residues may be rapidly attacked by termites and ants, so that the residues or mulch are ineffective during the following rains (Fisher, 1977).

(7) Zero-tillage (or no- or minimum-tillage) practices have been found to be effective in reducing soil erosion in some areas. Lal (1976) reports negligible soil losses from no-tilled maize and cowpea plots in Nigeria, and several studies have shown its effectiveness in North America. On the other hand, Othieno (1975) found that soil losses were large on tea plots treated with herbicide (as high as manual weeding) and Kalms (1977) quotes an example from Ivory Coast where no-tillage created larger soil losses than conventional ploughing.

Gurnah (1975) has reviewed the benefits to be gained by tillage and no-tillage practices. Tillage increases soil porosity, aids seed germination and root development, and reduces weed germination, insects and diseases in the crop residue. Detrimental effects of tillage can be increased rates of organic matter decomposition, leaching of nutrients, creation of impermeable pans, increased water evaporation and late planting related to problems of dry season ploughing.

Application of a herbicide followed by minimum-tillage has been found to produce good crop yields in a variety of temperate and tropical areas. Whilst minimum-tillage may be feasible for large scale farming in East Africa, there are problems facing its application to small scale farms. The costs of herbicides and their application may outweigh any benefits from labour saving and improvements in soil fertility, water conservation and soil erosion. Incomplete sod kills can result in drastically reduced crop yields, and thus careful applications are required. Gurnah (1975) comments that more work is needed on the effect of zero-tillage practices in East Africa. This approach cannot be recommended until more is known about its effect on soil erosion.

Conclusion

Modern studies of soil erosion problems have emphasised the importance of good crop and management techniques in reducing soil losses from agricultural systems (e.g. Greenland and Lal, 1977; Hudson, 1971). Certainly, improvements in cropping systems through mulching, mixed cropping and increasing plant densities can go a long way in controlling soil erosion in humid areas.

It must be realised, however, that crop-management techniques are of more limited value in areas where the soils are steep, shallow or infertile, the rainfall is low and erratic and the population pressures are high. For example, Fisher (1977) and Thomas (1974) conclude that there is no effective, readily-available crop-management practice to control erosion from the Machakos and Kitui areas of Kenya. In these cases, effective and well laid out mechanical measures, such as terraces, ridges and cut-offs are the major solution to the soil erosion problem.

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