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Land erosion by water in different climates

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Abstract

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Five major factors affect erosion and transportation processes by water: climate, relief, soil, vegetation and man's activities. These major factors consist of elements whose effects on erosion and transport are considered and whose interdependence is dealt with. In order to discern climatic differences in erosion, soil loss models and erosion maps are analysed. Relationships between precipitation and sediment yield are examined. An attempt to assess the importance of the major factors in different climates is made. Global and regional maps of sediment yield are reproduced and the construction methods and the reliability of the maps are considered. Fournier's and Strakhov's global maps are compared with regional maps in order to evaluate the reliability of the global maps. Other types of erosion maps which do not show sediment yield and which do not give numerical values for the erosion rates are also studied as well as maps which indicate the relative erosion potential or which give numerical values of potential erosion.

The main conclusions of this study are that rainfall erosivity is great in Af, (Am), Aw and Ca climates and lower in other climates. Low sediment yields are found in Db, Dc, and Dd climates. High values occur in Cs, Ca and, in many cases, in Da climates. Graphs of relations between precipitation and sediment yield indicate high sediment yield in climates which experience seasonal rainfall and in BS climates. The map analysis indicates both high and low sediment yields in BS climates. Graph and map analysis also indicates increased erosion in mountains (with the exception of the eastern USSR), especially in agricultural, pasture and loess soil areas.

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SHORT REVIEW OF MULTIPLE REGRESSIONS OF SEDIMENT YIELD WITH A
DISCUSSION OF THE VARIABLES

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Preface

There is abundant literature dealing with erosion problems in both small and large drainage basins, but there is great difficulty in comparing the results, for there is no general frame of reference on which to bare a comparison. One purpose of this study has been to determine the conditions under which erosion and transport processes operate in different climates and, thereby, to examine the effectiveness of the various methods of representing those conditions that have been used in the past.

The main object of this study is to investigate the global distribution of erosion under present conditions. Erosion under natural conditions is termed geological erosion, while erosion which is accelerated by man's activities is termed accelerated erosion (cf. Hudson 1971). It is especially interesting to discern the effect of climatic differences upon the magnitude of geologic erosion

and accelerated erosion. It was not possible for a single investigator to compile and process a comprehensive amount of raw data from different climatic settings. Instead, in order to achieve its main purpose, this study has focused on analysing and discussing the results obtained by other researchers, and on trying to draw some general conclusions.

I should like to extend my deepest gratitude to Professor Åke Sundborg, Department of Physical Geography, University of Uppsala, for his guidance of my work and his constructive criticism of my many rough drafts. I should also like to thank Dr. Thorsten Stenborg for his scrutiny of and comments on this study. Thanks are also due to Dr. Valter Axelsson for his encouragement and helpful discussion. Finally, I should like to thank Kjerstin Andersson for her many helpful suggestions concerning the maps.

1 General introduction

In this study "land erosion by water" has been defined as interfluvial and fluvial erosion of solid material. Dissolved material is not considered, nor is abrasion by marine action or erosion by wind and glaciers.

Processes of erosion and transport by water are affected by many factors such as climate, relief, soil, vegetation and man's activities. The general character of these processes is the same all over the world but the intensity of the processes is different and depends upon the characteristics of the enumerated factors.

Part I of this study contains a presen-

tation of the influence of the above-mentioned factors on the erosion and transport processes with a view to clarifying the effect of climatic differences. It is based on a large number of publications but does not include complete coverage of all available references. *Part II* contains an examination of mathematical models in order to throw light upon climatic differences in erosion and transport and to determine the significance of variables selected to represent the main factors of climate, relief, soil, vegetation and land use in different climates. *Part III* contains a presentation of erosion maps and the characteristics of such

maps which arise from the method of construction. It also includes a comparison of rates of erosion in different climates as inferred from the map analysis.

In a general or detailed survey of land erosion, certain common problems arise and they are accentuated when surveys are compiled as they are in this study. The following key problems were encountered:

1. *THE AREA PROBLEM.* One fundamental difficulty is that different values of the rate of erosion per unit area are obtained depending upon whether the study area is small or large. This is, among other things, due to the different transport lengths involved. Plot studies give a measure of the effect of splash, sheet and rill erosion and the transport of the eroded material to the end of the plot. Sediment yield from small drainage basins implies that the material has been transported over greater distances, first in interfluvial areas to a small channel and then through the drainage net to the point of measurement. Material from gullies, channels and mass movements is included. Within drainage basins, there is always some deposition at the end of slopes, in depressions, in lakes and reservoirs, on floodplains and in channels. These conditions are still more pronounced in large drainage basins. Therefore, it is impossible to compare directly rates of erosion for different regions and to use plot values in one case and values from drainage basins in another. An answer must be found as to which area should be used for the determination of rates of erosion: plots, slopes, small drainage basins or large drainage basins.

2. *THE SOURCE PROBLEM.* It is difficult to determine where in a basin most sediment originates. The main sediment source may be the interfluvial areas or the channels. Interfluvial erosion may be further divided into sheet, rill and gully erosion and is usually con-

centrated in part of the interfluvial areas. Material is also brought to streams by creep and rapid mass movements. These sources of sediment must therefore be considered and discussed. Material from all sources is measured at sediment gauging stations. One particular source may, however, contribute much more sediment in one basin than in another. Only a few types of sources are normally represented in plot studies.

Different sources of sediment have varying significance in different climates. Moreover, changes in land use or soil conservation measures, dam construction or channel training can change the rate of sediment transport and channel erosion. The influence of non-equilibrium slopes, caused by, for instance, tectonic movements, volcanic activity and former glaciation, is not treated in this study. Nor are periglacial processes discussed.

3. *INTERCORRELATION PROBLEMS.* It is difficult to analyse the influence of one particular variable on a certain process, as the rate of the process may depend on many variables which are mutually related to each other. As an example we may look at the peak water discharge of a storm flow. The peak flow may be correlated with climatic conditions but also with soil and land use variables, relief, drainage density, and some variable expressing the shape of the basin. It is difficult to determine the primary cause of the magnitude of peak discharge, and to discern the interrelation between all variables involved.

The soil moisture content may be chosen as another example. High soil moisture has both positive and negative effects on erosion processes. It increases the chances of surface runoff, but decreases the detachment rate because of higher cohesion, more secondary aggregation and less destruction of aggregates due to rapid wetting. The relative effect of soil moisture conditions on ero-

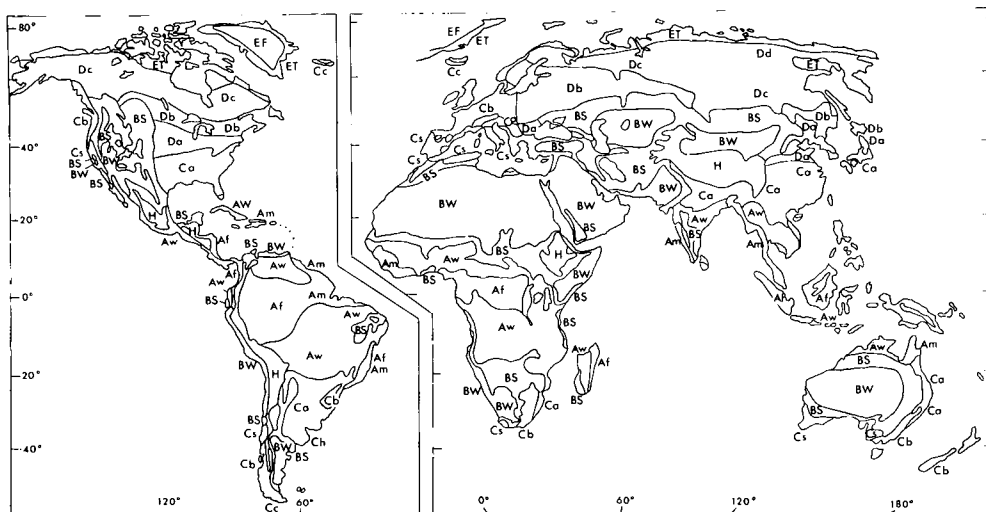


Fig. 1. World distribution of the climatic types of the modified Köppen classification found in Table 1.

sion may be different in different climates, because other factors that affect the detachment processes and runoff may be of relatively greater importance compared to soil moisture.

4. PROBLEMS OF REPRESENTATIVITY. The distribution of measurements or samples in time and space is often a major problem in evaluating how representative the results are. In investigations of a particular drainage basin, it is important that the time distribu-

tion of measurements also include events of high magnitude and low frequency. This is of especially great importance in arid and semi-arid areas. It must also be stressed that a few "point measurements" of erosion can be misleading, if they are not representative of the whole area or of a sufficiently long period.

Due to the varying availability of data, global studies are often based on data which have a biased distribution in favour of particular climates. A similar

Table 1. The Köppen climatic classification as modified by Trewartha (Finch *et al.* 1957) cf. Fig. 1

Groups of climate	Types of climate
A. Tropical humid climates (all months over 18°C)	Af Constantly wet tropical Am Monsoon variety of wet tropical Aw Tropical wet and dry BS Semi-arid or steppe BW Arid or desert
B. Dry climates (potential evaporation > precipitation)	
C. Humid mesothermal climates (coldest month 18°C – -3°C)	Ca Humid subtropical, warm summer Cb Marine, cool summer, (≥ 4 months $> 10^\circ\text{C}$) Cc Marine, cool summer, (< 4 months $> 10^\circ\text{C}$)
D. Humid microthermal climates (coldest month under -3°C, warmest month over 10°C)	Da Humid continental, warm summer Db Humid continental, cool summer Dc Subarctic, cool summer Dd Subarctic, cold winter
E. Polar climates (warmest month under 10°C)	ET Tundra climate EF Ice cap
H. Undifferentiated highlands	

problem of representativity is encountered when studies, equations and maps are compiled from all climatic zones and all parts of the world. The paucity of maps of erosion has been a special problem in this study. The distribution of equations is also regionally biased in favour of the United States and the Soviet Union.

5. **SCALE PROBLEMS.** As a rule, it is difficult to compare large scale and small scale maps. The detail and precision of a particular map often depends on its scale. However, the degree of generalization may also be due to a lack of basic information.

Differences regarding the definition of the erosion intensity intervals is another

difficulty that arises with map comparisons. Small scale maps often have broad intervals. It is inconsistent to compare rates of erosion from, e.g., one map showing 60–600 tons/km² with another map showing 20–100 tons/km² for the same region. Such comparisons may lead to inexact results.

In spite of the problems discussed, this study has attempted to obtain as complete a picture as possible of land erosion by water in different climates.

The climatic classification used in this book is the Köppen system as modified by Trewartha (Finch *et al.* 1957). The symbols found in Table 1 and Fig. 1 will be used throughout the text.

Part I FACTORS AFFECTING EROSION AND TRANSPORTATION PROCESSES BY WATER

2 Introduction

This study is concerned with comparisons of water erosion and transportation in different climates. Therefore, it is important to understand the influence of different elements on overland flow generation and the detachment and transport of material in interfluvial areas and in channels. The elements can be grouped under five major headings: climate, relief, soil, vegetation and land use.

Part one consists of a description of how the elements influence erosion and transport during a short (steady) time span and how they influence other ele-

ments in a short or a long (steady or graded) perspective. Because of this attempt to elucidate interrelationships between the elements and because of an effort to consider all or most effects of one element in the same section, some information is repeated in two or more sections. A third main objective has been to describe the variation in the elements with climate. This implies a transfer of interpretation from a plot scale to a global scale (cf. Schumm & Lichty 1965, Morgan 1973).

3 Climate

3.1 Precipitation

3.1.1 General remarks

The main agents of erosion and transportation are the impact of raindrops and the surface flow of water. The kinetic energy of raindrops and the shear stress of overland flow accomplish the erosional and transportation work. These agents are discussed below. Precipitation intensity is responsible for the kinetic energy with which much of the work is done. In order to investigate climatic differences in water erosion, it is necessary to clarify the

variations in rainfall intensity that are associated with different climates. This is followed by a discussion of the distribution of overland flow within river basins, and of climatic and topographic effects upon the amount of overland flow. Finally, some rainfall parameters are discussed.

3.1.2 The impact of raindrops

Raindrop impact affects erosion and transportation in a number of ways. Drop impacts break down soil aggre-

gates, detach particles and splash material in all directions. As particles are splashed a longer distance in the downhill direction than in the uphill direction, there is some net transport downhill. The net transport is a function of slope inclination. Ekern (1953) used a small target of standard soil in experimental studies of the splash process. He found that a certain minimum impact was necessary to initiate splash erosion, which for fine sand was approximately equal to the impact of a raindrop of 1 mm diameter, falling at terminal velocity. With the soil factor remaining constant, Ekern found splash erosion to be a function of:

$$\text{rainfall intensity} \times \text{duration} \times \text{drop mass} / \text{drop area at impact} \times \text{drop velocity}^2$$

In similar experiments by Ellison (1947), the amount of soil splashed during a 30-minute period was proportional to:

$$KV^{4.33}d^{1.07}I^{0.65} \text{ where}$$

K = constant for the soil type

V = velocity of raindrops

d = diameter of raindrops

I = rainfall intensity

The exponents of both rainfall intensity and drop velocity vary between the two studies. The amount of splashed material has also been correlated with kinetic energy (Free 1960) and momentum (Rose 1960).

A high speed photographic study of splash greatly illuminated the process of splash erosion (Mutchler & Young 1975). The depth of the film of surface water has a bearing on the geometry of splash shape and the total amount of water splashed. Raindrop impact is most erosive when a very thin layer of surface water is present, approximately one-fifth of the drop diameter, and is relatively non-erosive when the soil is covered with water with a depth of three drop diameters or greater.

Raindrops have been shown to have a great potential for detaching soil but a limited ability to transport soil directly. In a laboratory study of a 1.52 m × 4.52 m plot, most of the soil movement to the rills was accomplished by surface flow (Young & Wiersma 1973). However, the ability of the surface flow to transport the detached soil particles was greatly enhanced by drop-impact-imparted velocities. The horizontal velocity of water moving away from the impact site along the surface was a function of the maximum splash crater width and the time taken for the splash to attain its maximum height (Mutchler & Young 1975). Splash width and time vary with the surface water depth and the drop diameter.

It has been observed that soil-bearing water, with a depth of more than 13 mm and a velocity of more than 275 m/h (7.6 cm/s) deposited soil on a city pavement at each point where it flowed under a parked car (Ellison 1947). The flow was not highly charged with soil, but under each car a deposit was left which completely covered the concrete. There was no deposition between the cars. This observation is a striking example of the effect of raindrop impacts on the transport capacity of surface flow.

Ellison (1947) stated that when splashing raindrops break down clods and aggregates and release humus and clay materials, the water becomes so turbid that colloids and some fines approaching colloidal dimensions are carried down into the soil profile. Thus, the soil's permeability is decreased. The turbid water also clogs the pores at the surface. In addition, the impact of raindrops seals the pores at the surface through compaction. The size and the velocity of the drops are thus important for the degree of compaction. The sealing of the pores reduces infiltration and thus causes greater surface runoff than would otherwise be the case.

Hudson (1958) tried to demonstrate the result of splash erosion by comparing two 1.5 m × 27.5 m plots. Both plots were kept free of vegetation by hand weeding. Two layers of mosquito gauze, 5 cm apart, were suspended over one plot, 15 cm above the soil. Measurements by rain-gauges in each plot showed that all the rain passed through the gauze. However, the mesh broke up the falling raindrops so that they reached the soil surface as small droplets and thus had less power to compress and detach the soil and to disturb the surface water. The 10-year mean annual soil loss from the gauze-covered plot was 94 tons/km²/yr compared to 12,657 tons/km²/yr for the unprotected plot. The major cause of the great difference in resulting soil loss was therefore assumed to be the splitting of the raindrops.

The effects of raindrop impact are also apparent when a bare soil containing stones is exposed to a heavy storm. The stones may be found resting on miniature earth pedestals several centimeters high because of the lowering of the surface of the surrounding unprotected soil.

In conclusion it can be said that raindrop impact detaches soil particles which are then transported by a thin film of flow to more concentrated flows in rills. The amount of overland flow is increased by the sealing effect of raindrop impact. Raindrop impact on overland flow increases its transport capacity.

3.1.3 The kinetic energy of rainfall

Raindrops accomplish much erosional and transportational work. Therefore it is essential to know the amount of erosive power of the rain. The erosivity, i.e. the potential of rain to cause erosion, is dependent upon the kinetic energy of the

rain. The equation for kinetic energy is:

$$KE = \frac{1}{2} m v^2 \dots\dots\dots (1)$$

where KE = kinetic energy, m = mass, v = velocity.

Thus, if we know the rainfall intensity, the drop size distribution of the rain and the velocity of each drop size class, we can calculate the kinetic energy.

The terminal velocity of a raindrop depends on the drop mass which is a function of drop size. Drop size distribution is related to intensity (Mutchler & Young 1975 and Lal 1977, after Laws & Parsons 1943, Ekern 1953, after Gunn & Kinzer 1949 and Best 1950). Thus kinetic energy might be expressed as a function of rainfall intensity. Using the drop size distribution of rainfall determined by Laws & Parsons, Wischmeier & Smith (1958) derived an empirical kinetic energy equation. In metric units the equation becomes

$$KE = 210.2 + 89 \log I \dots\dots\dots (2)$$

where KE is kinetic energy in ton×m/ha per cm of rain and I is rain in cm/h. One ton×m/ha = 0.9807 Joules/m² (1 tm/ha = 0.9807 J/m²).

Other relations between intensity of rain and kinetic energy have been derived from different parts of the world (Fig. 2). It is not surprising that such relations do not coincide, as drop size distribution varies in different types of rainfall. Some of the variation may also be due to different measurement techniques. In Fig. 2, the curves from India show considerable divergence for similar intensities with different types of rain. Hudson's curve (Fig. 2) from Rhodesia shows a very steep increase in energy with an increase in intensity up to 25 mm/h, followed by a somewhat less steep increase up to 75 mm/h and then nearly no increase in kinetic energy/mm of rain for higher intensities. However, it should be noted that the total energy increases as the amount of rain per hour increases.

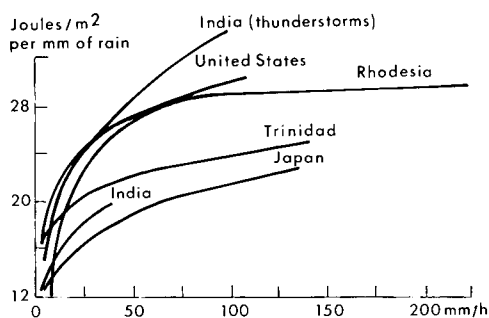


Fig. 2. The relation between the kinetic energy of rainfall and rainfall intensity as observed in different countries. Each curve extends to the highest recorded rainfall intensity (from Hudson 1971).

Kowal & Kassam (1977) studied the kinetic energy of rain at Samaru in the savanna region of northern Nigeria. They recorded the number and amplitude of the electrical pulses originating from the impact of raindrops in order to classify drops according to size. Records of 18 rainstorms with 6.6–60.6 mm of rain in 1973 and 1974 were analysed. Kinetic energy per mm of rain ranged between 21.8 and 39.5 J/m²/mm with a mean of 32.2 J/m²/mm. This is a higher value than that quoted by Elwell & Stocking for Rhodesia, where the seasonal average during a period with 910 mm rainfall was of the order of 19 J/m²/mm. Kowal & Kassam suggested that this was because the rainfall in the study area in Rhodesia was predominantly composed of drops less than 2.0 mm diameter. This is similar to the drop size distribution pattern found in England. Another possibility is that Kowal & Kassam's calibration was not quite accurate.

At Samaru, 59 % of all drops are larger than 3 mm in diameter. The modal value of the drop diameter in Rhodesia rises to 2.5 mm at approximately 80–100 mm/h but decreases again at still higher intensities (Hudson 1971). This occurs because there is apparently an upper limit to the maximum drop size. It has been shown in wind tunnel experiments that drops are stable up to 4.6 mm diameter

and unstable above 5.4 mm, with a transitional range in which drop disintegration is dependent upon the amount of turbulence (Hudson 1971 and Ekern 1953, after Blauchard 1950).

The wind velocity accompanying a storm may influence the kinetic energy. This is especially true for small intensities. However, the acute angle of the rainfall may counteract this effect to some extent. Slope direction, in relation to the direction of the storm, can also significantly alter the kinetic energy per unit area.

Hudson (1971) made a rough comparison of kinetic energy for temperate and tropical rainfall conditions. Rains were assumed to be erosive if intensities are > 25 mm/h. As a consequence, about 5 % of temperate rain and 40 % of tropical rain were considered to be erosive. The annual rainfall amounts for temperate and tropical areas were assumed to be 750 mm and 1500 mm, respectively. A higher intensity, and thus higher kinetic energy, was estimated for tropical erosive rainfall, 28 J/m²/mm rain, than for temperate erosive rainfall, 24 J/m²/mm rain. With these assumptions the annual erosivity values were:

Temperate rain: $0.05 \times 750 \times 24 = 900$ J/m²

Tropical rain: $0.40 \times 1500 \times 28 = 16,800$ J/m²

3.1.4 Overland flow

Overland flow occurs when the rainfall intensity or meltwater production is greater than the infiltration rate, or when subsurface flow returns to the surface to become overland flow or return flow. Overland flow may be divided into rill flow and interrill flow (Meyer *et al.* 1975, Foster & Meyer 1974, Foster *et al.* 1977a, b). Interrill flow is a very thin film of water flowing on surfaces between the rills.

Rills are not developed under all conditions. Field investigations of vegetated plots sprinkled at an intensity of 200 mm/h showed that the flow rarely occurred as a uniform sheet of water and that the majority of water travelled downslope in several lateral concentrations of flow (Emmett 1978). However, these concentrations were not considered to constitute rill flow. The greater flow depths observed in the field runoff tests, compared to the laboratory tests, were due primarily to vegetation and microtopography. Hence, overland flow may consist of numerous separate unchanneled water courses in an anastomosing pattern (Morgan 1980).

It may be very difficult to observe overland flow in the field. Despite efforts to observe overland flow resulting from thunderstorm activity, during several weeks residence each year over a ten-year period at a project area in a semiarid part of New Mexico, overland flow was never observed in the field (Emmett 1978). Yet during the ten-year period of measurement at permanent installations, surface erosion on unrilled slopes yielded 5251 tons/km² per year.

Transportation of splash-detached particles to rills from interrill areas is primarily accomplished by thin films of water. Direct splash of detached particles through the air to rills appears to be of minor importance compared to soil

transported by thin film flow (Young & Wiersma 1973, Foster & Meyer 1975).

In an experimental study of silt transport by thin film flow, where silt particles were introduced to sheets of flow less than 0.3 cm deep, the transport was by saltation rather than by suspension (Podmore & Merva 1971). Clay may either be transported in aggregates as bed load or as dispersed clay in suspension.

Interrill flow without impacting raindrops is probably able to transport only a small load (Foster & Meyer 1975), cf. also Table 2. When a raindrop hits a thin water film, the water is pressed in all directions from the point of impact. The shear resulting from the velocity acceleration may then be sufficient to move detached particles some distance downslope. The size of the velocity fluctuations depends upon the depth of the water and the drop size.

Meltwater overland flow normally carries less sediment than rainwater overland flow. In an experimental drainage basin in central Utah, 4.6 % of the 6 year average annual surface runoff was caused by summer rainstorms, when there was a 16 % cover of vegetation (Forsling 1931). This summer runoff caused 84.5 % of the erosion, while the remainder of the runoff, which was from melted snow, caused 15.5 % of the erosion, even though the soil in this study

Table 2. Results from simulated rainfall of 46 mm/h for 30–165 minutes with different time lapses between runs, with and without insect screens 5 cm above the soil, on a 1.52 m × 4.52 m laboratory plot with a 9 % slope (from Young & Wiersma 1973)

	Velocity on surfaces be- tween rills		Depth of water between rills mm	Erosion between rills mg/cm ² /min
	cm/s	m/h		
Barnes loam: without screens	4.0	144	<0.15	1.20
with screens				0.11
Crofton silt loam: without screens	3.0	108	0.12	0.60
with screens				0.03
Central sandy loam: without screens	1.5	54	0.08	1.10
with screens				0.11

was not frozen during the period of runoff.

When water is concentrated into rills or when there is a deep sheet of flow on a slope, erosion by flow is important. The water exerts a drag force on the surface particles. The algorithm for the bottom shear stress is (Sundborg 1956):

$$\tau_o = u_*^2 \rho \dots\dots\dots (3)$$

where τ_o = boundary shear stress

u_* = shear velocity

ρ = liquid density

With increasing depth and increasing gradient, the velocity, and thus the shear stress, is increased and the water has a greater ability both to erode and transport material (cf. Table 3).

In an experimental study in non-cohesive material of different grain sizes, it was observed that rainfall during runoff increased the erosion of smaller particles but decreased the erosion of larger particles as compared to the same rate of runoff without rainfall (Meyer & Monke 1965). The causes of increased erosion when small easily transported particles made up the material were suggested to be increased sediment availability due to splash and increased carrying capacity due to raindrop induced turbulence. The

causes of decreased erosion of large particles were suggested to be decreased carrying capacity of the rills due to a retarding of the flow by raindrop impact and smaller depths of the rills as the sides were worn down.

The state of the flow also helps to determine the magnitude of the shear stress. Several researchers have come to the conclusion that overland flow is laminar (Kilinc & Richardson 1973, Woolhiser *et al.* 1972, Emmett 1978); others, that it is turbulent (Podmore & Merva 1971); still others, turbulent or laminar depending on the grain size, depth and velocity (Blom 1962, Phelps 1970). It has been called agitated supercritical laminar flow (Kilinc & Richardson 1973) or disturbed flow (Emmett 1978).

With regard to shallow channel flow, the effect of drop impact depends upon the depth of the flow and the material. Tests on the effect of rainfall on shallow-channel flow with 3–15 cm depths indicate a reduction in the mean velocity. Turbulent mixing is increased (Glass & Smerdon 1967) and suspended sediment transport is generally reduced by rainfall (Smerdon 1964) because of the reduced velocity gradient.

Table 3. Results from simulated rainfall on bare, sandy soil in a 1.5 m × 4.9 m flume. τ_o = boundary shear stress. τ_c = critical tractive force. (Data from Kilinc & Richardson 1973)

Slope %	Intensity of rain mm/h	Mean velocity at flume end		Depth at flume end mm	$\tau_o - \tau_c$ at flume end N/m ²	Sediment discharge tons/h/km ²
		cm/s	m/h			
5.7	32	5.4	193	0.48	0.22	9.75
5.7	57	9.1	327	0.73	0.59	31.90
5.7	93	12.8	461	0.93	0.86	65.37
5.7	117	15.3	553	0.99	1.23	148.29
20.0	32	9.8	352	0.33	1.40	65.46
20.0	57	14.4	518	0.48	2.18	562.66
20.0	93	21.9	788	0.55	2.54	1466.94
20.0	117	25.2	908	0.63	3.06	2647.40
40.0	32	12.3	444	0.28	2.15	106.84
40.0	57	19.0	686	0.37	2.70	1227.76
40.0	93	28.1	1011	0.44	3.07	3508.20
40.0	117	32.1	1155	0.49	3.46	6108.52

3.1.5 Global differences in precipitation intensity

Rainfall of extreme intensity, e.g. storms with a 10 to 100 year recurrence interval, may accomplish a large part of the erosional, transportational and mass movement work in some climatic settings as in the Mediterranean and Saharan zones (Roose 1977). This study will not consider rapid mass movements, but it should be remembered that in some environments rapid mass movements and slide scars are greater sources of water transported material and rapid mass movements are often triggered by high intensity rains.

Rainfall of extreme magnitude affects sediment yield. It is useful to construct graphs of the recurrence or return period of maximum daily or hourly rainfall. As can be seen from Fig. 3, low annual rainfall areas may have comparatively high daily rainfalls for high return periods. Especially on the margins of the arid zone the recorded daily maximum precipitation may be considerably greater than the mean annual values for a long period. In drier parts of monsoon India, daily rainfall sometimes exceeds the mean monthly rainfall. The graphs of

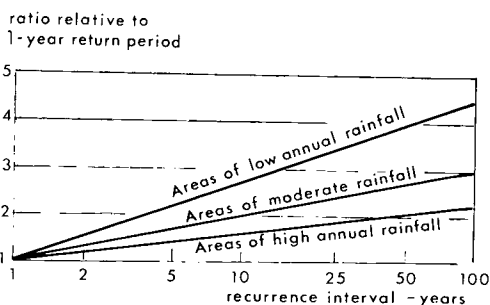


Fig. 3. Ratio of 24-hour maximum rainfall intensity for different return periods, relative to 24-hour maximum intensity with a 1-year return period (from US Dep. Army 1963).

Jodhpur and Poona illustrated in Fig. 4 exemplify this.

A generalized global distribution of high intensity rainfall is illustrated in Fig. 5. The Afro-Asian low rainfall area has low 2-yr, 1-hour maximum intensities, but as has been pointed out above, the variability associated with arid areas is considerable. Detailed maps of 2-yr, 1-hour maximum rainfall for Algeria and Tunisia can be found in a report from FAO (1976a). Tropical areas with a high annual rainfall have high 2-yr, 1-hour maximum intensities. Moreover, a large part of the high annual rainfall in Af and

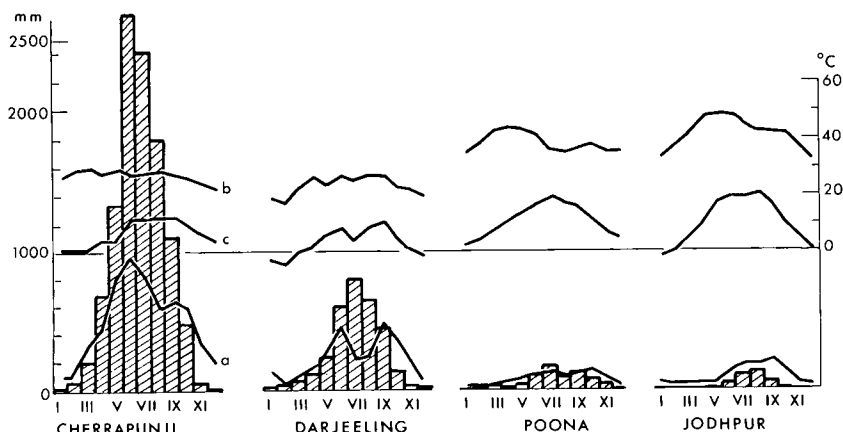


Fig. 4. Monthly rainfall histograms and heaviest recorded rainfall in 24 hours (a). Extremes of maximum (b) and minimum (c) temperatures. Based on data up to 1959 (from Starkel 1972).

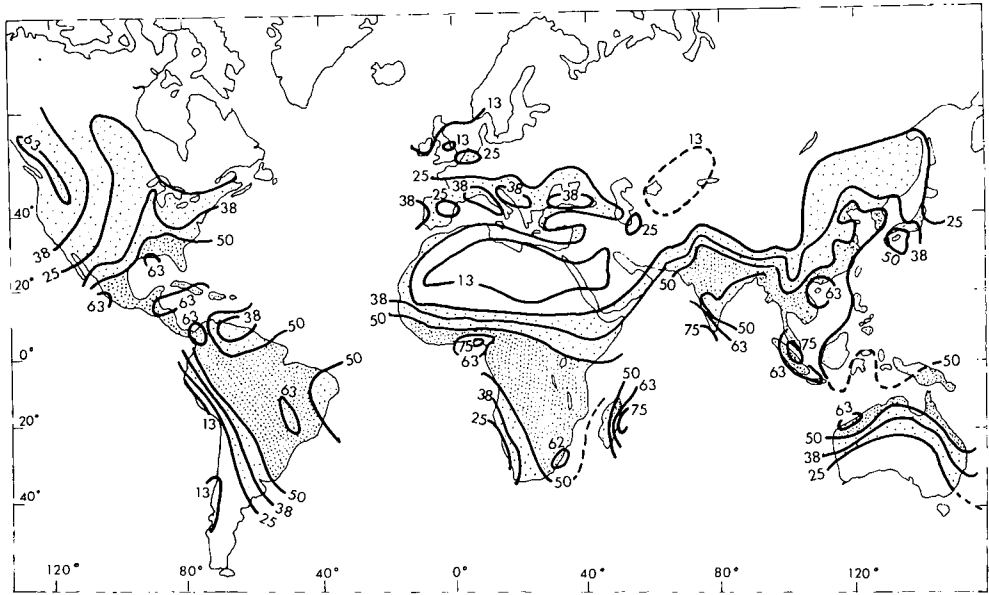


Fig. 5. Isohyetal map of 1-hour maximum precipitation with a 2-year return period, in mm (from Dunne & Leopold 1978, after US Dep. Army 1963).

Aw climates consists of high intensity rains. Near Ibadan, Nigeria, 15 %, 10 % and 30 % of storms had maximum intensities > 75 mm/h during 1972, 1973 and 1974, respectively (Lal 1976).

De Ploey (1972) compared per hour intensities for Kinshasa, Zaire, 1964–1965, and Antwerp, Belgium, in 1968 (Fig. 6). Antwerp has a concentration of intensities under 2 mm/h with a duration varying between 15 and 800 minutes. These intensities occur less frequently in Kinshasa. The maximum intensity recorded in Antwerp amounted to 68 mm/h and the two second highest intensities were about 20 mm/h while Kinshasa had 11 periods of rainfall ≥ 20 mm/h, with a maximum value of 115 mm/h. Mean rainfall intensity was 0.96 mm/h for Antwerp and 4.74 mm/h for Kinshasa. Mean intensity of all rains was calculated by taking into account both intensity and duration.

Starkel (1972) reported high rainfall intensities in the Indian monsoon climate. In the region of Cherrapunji and on the

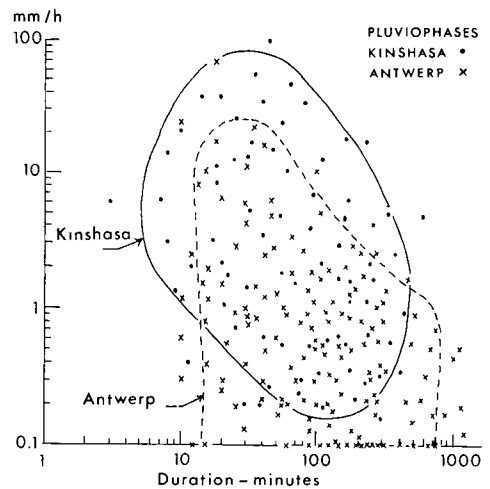


Fig. 6. Duration of rainfall intensities for Antwerp and Kinshasa. Rainfall occasions are divided into phases which correspond to approximately similar intensities (from De Ploey 1972).

scarp of the Ghats, the rainfall in some summer months may be as high as 2500–2800 mm in 26–30 days, i.e. on the average 50–100 mm of precipitation every day. The maximum recorded daily

precipitation at Cherrapunji amounted to 974 mm in June 1956. In the region of Bombay, 983 mm in one day has been recorded. The mean annual maximum daily precipitation for Bombay is as high as 550 mm.

In Af, Aw and Ca climates, infrequent tropical cyclones may achieve a considerable intensity, up to 300 mm/h. Tropical cyclones often give continuous rains of 500–2000 mm, which last for two to three days, with intensities up to 50 mm per hour (Starkel 1975). Daily precipitation totals of ≥ 1000 mm are known from Japan, Taiwan, the Philippines, Assam, Reunion, Angola and Jamaica (Starkel 1975, after Jennings 1950).

25 mm/h is discussed in the literature as a threshold for erosive rains, and the global distribution of this intensity is therefore of interest. Areas with maximum 1-hour precipitation > 25 mm/h during a 2-year period are shaded in the map, Fig. 5. The annual duration of intensities ≥ 24 mm/h, is found in Table 4 but is calculated as a summation of 1-minute or 4-minute durations of intensities ≥ 24 mm/h. The table values are uncertain as they are calculated for a very limited time period.

The map in Fig. 5 shows that in the Cb climate north of the latitude of Paris, maximum intensities corresponding to $>$

25 mm/h do not appear every second year. However, both Paris and Koblenz (Table 4) have a total period of 60 minutes of short duration (≥ 1 minute) high intensity rainfalls per year. The two stations in England only have totals of about half an hour. The two stations with the highest values in Table 4, with about 15 hours of rain > 24 mm/h, lie in Ca climate and have 50–75 mm/h 2-yr, 1-hour maximum precipitation in Fig. 5. A frequency map by Visser (1945) shows that on seven days a year in an average year in Florida there is a rainfall ≥ 25 mm within one hour.

In many parts of the world there are no recording raingauges. Many stations only record 24-hour precipitation, while most storms last for only a few hours or less. Therefore, it is a great problem to obtain information about high-intensity, short-duration rainfalls. A map of annual 24-hour maximum rainfall in Australia, based on data from 330 stations (Jennings 1967), and a 2-year, 24-hour maximum rainfall map of the USA (Atlas of American Agriculture 1936) show that the 24-hour maximum patterns closely resemble the mean annual patterns. Another measure of intensity, based on 24-hour measurements, is amount of rain per rain day. Low rainfall areas in Australia also have low rainfall per rain day.

Table 4. Cumulative annual duration of precipitation ≥ 23.88 mm/h as a summation of 1-minute or 4-minute periods (data from Sims & Jones 1973)

Station	Years of measurement	Minutes per year
Paris, France	1970, 1971	63
Preston, England	1970	38
Reading, England	1970	29
Koblenz, Germany	1970	84
Freiburg, Germany	1964, 1965	48
Seattle, Washington, USA	1965	8
Urbana, Illinois, USA	1969–1972	308
Bismarck, North Dakota, USA	1966	136
Albuquerque, New Mexico, USA	1952	24
Bet Dagan, Israel	1971	168
Naha, Okinawa, Japan	1966, 1967	878
Miami, Florida, USA	Aug. 1957–Aug. 1958	989

But by looking at rainfall per rain day we can infer little about 1-hour or 1/2-hour rainfall intensities.

There is also another problem in the use of precipitation data. Point measurements at a rain gauging station are not representative of a large area. The measurement might be made in the centre of the storm or on the outskirts, or the storm might be very localized as is characteristic of semi-arid regions, where showers are also usually of short duration. The difference between point measured rainfall values and mean values for large areas must be kept in mind when, e.g., correlations between precipitation and runoff are made.

3.1.6 Distribution of overland flow

This section will deal with several aspects of the distribution of overland flow within river basins and also within particular climates. First there is a presentation of the two main types of overland flow, distinguished by way of production, and their climatic appearance. Next their distribution is considered and causes of the differences in distribution within different basins are discussed. Finally, there is a discussion of the amount of overland flow with respect to different climatic and topographic conditions.

The two types of overland flow are (Carson & Kirkby 1972);

1. Horton overland flow, which occurs when the rainfall intensity is higher than the infiltration capacity, and when surface depression storage has been filled.
2. Saturation overland flow which occurs when the soil is saturated.

Meltwater overland flow may be regarded as a type of Horton overland flow or as saturation overland flow depending upon whether the soil is either frozen or saturated.

Horton overland flow is determined by

the rainfall intensity (mm/h) minus the infiltration rate (mm/h) of the soil after subtraction of surface depression storage. High intensity rain seals the surface of bare soil very quickly, and consequently only a shallow depth of soil moisture may be achieved before ponding and surface runoff take place (Rubin 1966). The consequence of this is that in climates with sparse natural vegetation and high intensity rains Horton overland flow should occur. According to Dunne (1978), Horton overland flow is known to be common in arid and semi-arid lands, and in humid areas where the original vegetation and soil structure have been destroyed. In humid regions, the infiltration capacity remains high under natural vegetation (Dunne *et al.* 1975). At least in temperate, humid regions, where most research has been done, saturation overland flow is found to be the most common process of surface runoff under natural vegetation, and Horton overland flow is confined to surfaces such as roads, parking lots, skid trails in forests, ploughed fields, and artificial fills (Dunne *et al.* 1975).

In the past, Horton overland flow was considered to cover extremely wide areas. But as both rainfall intensity and infiltration rate vary spatially, in semi-arid regions as well as in other environments, it is now considered that overland flow only occurs within comparatively small areas. There may be differences in rainfall distribution between ridges and valley bottoms; a rainstorm may only cover part of the basin, and intensities within a storm may also vary.

Wind also plays an important role in the non-uniformity of local rainfall distribution. Results from tilted sprinkler measurements show that at a slope angle of 20°, with a rainfall inclination of 40°, a windward facing slope may receive almost twice as much rainfall as an adjacent leeward-facing one (Yair *et al.* 1978).

The steepness of a slope has an effect on the infiltration rate (Góczán 1972). On sloping ground under conditions of high intensity rainfall, water runs off more quickly than on flat ground, where water has more time to infiltrate. Kazó (1969) found that surface runoff increased with slope angle but to a varying extent for different soil types.

The amount of coarse material and vegetation cover varies within a basin, thus causing differences in infiltration. Infiltrating water may follow the root system if soil permeability is low. In a study by Craddock & Pearse (1938), the ability of wheatgrass to reduce surface runoff was believed to be due to the extensive fibrous root system, which presumably assisted infiltration. Vegetation also slows down the velocity of surface runoff. Thus water has greater time to infiltrate the soil. Sandstone bedrock may yield more homogenous detritus than shale, which may in turn produce more variable soil structures and more easily sealed material. There may also be bedrock revealed within a basin.

Yair *et al.* (1978) suggested that because rainfall duration is usually short in arid areas, overland flow generated on the upper part of slopes exceeding some 50 m in length probably has little or no chance of reaching the slope base at low or medium amounts of rain. Dunne (1978) on the other hand assumed that overland flow has sufficient velocity to reach stream channels during most rain storms.

Saturation overland flow consists of surface runoff from precipitation on saturated areas and return flow which is subsurface flow reentering the surface (Dunne & Leopold 1978). Saturation overland flow occurs in those regions where infiltration capacities are high because vegetation protects the soil from rain impact and creates a favourable soil structure. Saturation overland flow may occur at lower intensities than Horton

overland flow. The duration of the rain affects the degree of saturation. The depth to the water table is smallest near streams, where lateral throughflow water and groundwater from upslope collect. The zones most frequently saturated are therefore those adjacent to stream channels. Throughflow water from upslope areas will also collect in other places where local concavities occur and slope lines converge.

The soil depth is another factor affecting the time that is required to reach the saturation point. Soils are often thinner on steep slopes than on gentler slopes and level ground but at the beginning of a rainstorm the soil moisture on steep slopes is usually lower than on level ground. Thin soil cover and the presence of duricrusts and other less permeable soil layers lead to rapid saturation and surface runoff. Vegetation, depressions and dry soil are large water consumers at the beginning of a rainstorm. The potential evapotranspiration and the interval between storms also affect soil moisture and surface runoff (Temple 1972).

The saturated zones contributing to overland flow have been mapped in a few basins (Dunne *et al.* 1975, Dunne 1978, Dunne & Leopold 1978, Dunne & Black 1970). In a small basin in Vermont, with steep, straight slopes and deep soils, the extension of the saturated zone varied from 3 to 11 % of the basin area in different seasons. In a nearby basin with gentle hillslopes, concave footslopes and shallow soils, the saturated zone varied from 15 to 51 % of the basin area. The saturated zone in a small basin in glacial till and schists in Quebec varied from 36 % in March to 12 % in April. Other workers have also calculated contributing areas as a percentage of the whole basin (Betson 1964, Dickinson & Whiteley 1970).

It is difficult to delimit the probable extent of saturated areas in specific basins. Although the influence of topo-

graphy is clearly defined, it is difficult to develop models of the topographic influence. Kirkby (1978) found the depth of water in the saturated and unsaturated soil zones to be strongly dependent on a topographic parameter. This parameter was used to predict the areal extent of overland flow. Other ways of mapping the probable extent of saturated areas involve comparisons with depth of subsoil, gley morphology, poorly drained soils, presence or absence of individual plant species and the depth to the watertable (Dunne *et al.* 1975, Lundin 1982).

Distribution of amount of overland flow. Analysis of the recession limb of the flood hydrograph is often undertaken to distinguish the overland flow, throughflow and baseflow components of runoff (Jamieson & Amerman 1969, Lambert 1969, Holtan 1970).

The amount of surface runoff depends on many factors. The conditions in Asia will be discussed as an example. The map of surface runoff in Asia (Fig. 7) is based on equations in which the total river runoff is separated into its surface and underground components (Nikolayeva & Chernogayeva 1974). The climatic and topographic influences can be discerned. There are also soil and geologic influences. Surface runoff attains the highest values on the windward slopes of the Caucasus, the Himalayas and the Western Ghats. The proportion of surface runoff in mountain regions increases with elevation, as precipitation increases and evaporation decreases. This is especially apparent above the forest limit. On the leeward side of mountains, precipitation is much lower. High surface runoff is found in the southeastern part of the continent. This is the result of heavy showers. In some mountain areas soils are shallow and quickly saturated. In other parts of the wet monsoon regions, a gleyed soil horizon is formed, with the result that the

permeability of the soils decreases. The same phenomenon is found in rice growing regions. In the lowland region of the temperate belt surface runoff, as a percentage of total runoff, increases from north to south, from 50 % in the taiga zone to 85–90 % in the dry steppe zone. However, the absolute surface runoff values in the latter zone are small.

Geology is also an important factor. In regions where volcanic tuffs and karstic limestones are widespread, as for example in Turkey, the Shan plateau in Burma and the islands of Luzon and Mindanao in the Philippines, underground runoff dominates total river runoff and reaches approximately 70 %. The total runoff fluctuates with the wetness of the region.

Of course this is a very generalized picture of variations in surface runoff. Soil and vegetation differences dramatically affect conditions within the different climatic regions. However, the amount and intensity of precipitation, seasonality of precipitation and evapotranspiration determine the major features of the runoff zones.

3.1.7 Precipitation parameters and soil loss

Wischmeier and Smith (1958) correlated soil loss records from plots with different precipitation parameters. Rainfall energy gave a better correlation than rainfall amount, and maximum 30-minute intensity gave a better correlation than the maximum 15-minute intensity. The best fit was obtained for EI_{30} (the sum of storm energy $\times$ 30 minute intensity of all storms in a year). This parameter explained 71 to 89 % of the storm-to-storm variation in soil loss for different soil types. Energy alone explained 55 to 82 % of the variation in soil loss.

However, a parameter which is highly reliable in one particular area and in one particular soil may not be sufficiently

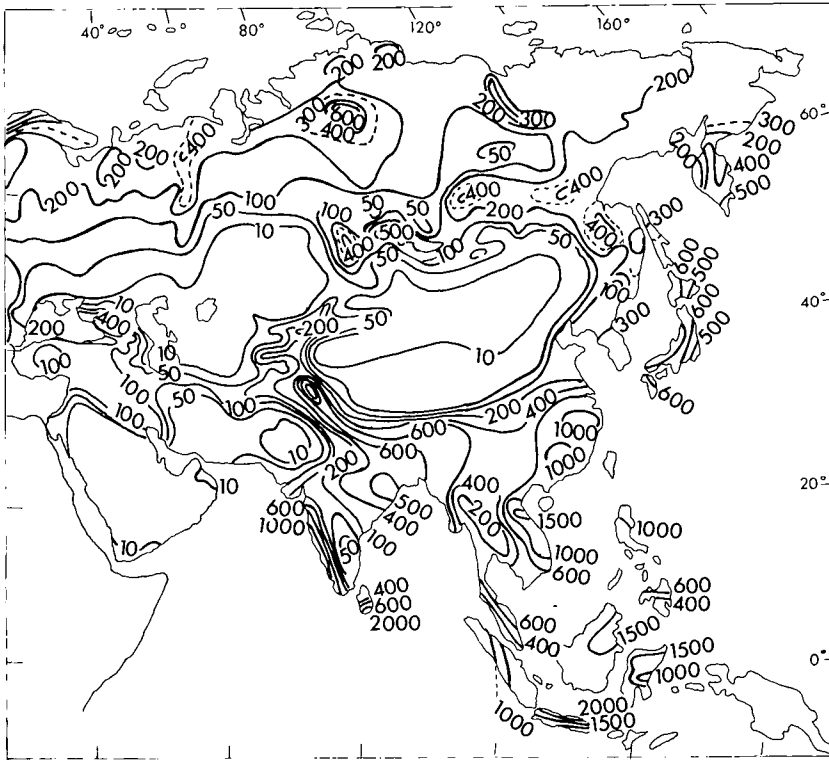


Fig. 7. Map illustrating the surface component of river runoff in mm (from Nicolayeva & Chernogayeva 1974).

representative in another. Lal (1976) investigated soil loss from plots near Ibadan, Nigeria. He found that soil loss was significantly correlated with an index of the product of rainfall amount and short duration intensity AI_m , where A = storm amount of rain, I_m = 7.5-min maximum intensity.

Hudson (1971) gave a threshold intensity for erosive rains, expressed by the index $KE > 25$ (kinetic energy for intensities greater than 25 mm/h). This index showed high correlation with soil loss from plots in Rhodesia.

Large storms also contribute greatly to the soil loss from plots in the USA. Cultivated single-cover plots were studied at seven locations throughout the USA (Piest 1963, after Wischmeier 1962). It was found that 3/4 of the soil loss was

caused by an average of four storms per year. It was also found that about 1/3 of the total soil loss was due to extreme storms with return periods greater than two years and that 1/3 was due to moderate storms with return periods between 1 and 2 years. This indicates that some high intensity rainfall parameters might be used as erosivity parameters in the United States. In fact, 2-yr, maximum 1-hour and 24-hours intensities may be correlated with the EI_{30} index (chapter 11).

As regards drainage basins, the magnitude of rainstorms may be ranked according to the magnitude of the runoff. Peak runoff or peakedness indices are both parameters which have been correlated with sediment yield in single catchments. However, difficulties arise

when several basins are compared. The relative sediment contribution associated with a given frequency of high runoff is very different in different climates and also within the same climate (Piest 1963, Douglas 1975). In one study of the USA, it was found that soil losses resulting from very large storms were proportionally greater in the semi-arid Great Plains than in the humid southeast (Piest 1963).

This is intended as a very short introduction to the topic of the relationships between climatic and hydrologic parameters and soil loss. A more detailed discussion is presented in chapter 11 in connection with the presentation of soil loss equations.

3.2 Temperature

3.2.1 *Effects on evapotranspiration and runoff*

Little attention is normally paid to the effects of temperature upon erosion. The effects of temperature on evapotranspiration and runoff are well known but its effect on erosivity, transport capacity and erodibility are less well documented.

High temperatures imply high evaporation and increased water use by plants. The result is that runoff decreases as temperature increases, assuming constant precipitation. In the eastern USA, streamflow per unit area to the Atlantic Ocean decreases from north to south, in spite of higher precipitation in the southern states (Bue 1970).

Surface runoff is affected by temperature in a number of ways. High evapotranspiration dries the soil between showers, and the infiltration rate at the beginning of a storm is greater than for a wet soil, thus contributing to a reduction in surface runoff. Temperature also affects the viscosity of the water, which in turn slightly affects the rate of infiltration and surface runoff. The higher the temperature, the lower the viscosity and

the greater the infiltration (FAO 1965).

Temperature may, however, sometimes have the opposite effect on runoff. High evaporation may, in some cases, give rise to duricrust formation by capillary action. Duricrusts decrease permeability and may increase lateral throughflow and overland flow and enhance gully formation.

3.2.2 *Effects on water erosivity and transport capacity*

The absolute viscosity of water varies markedly with temperature. It decreases from 1.73 centipoises at 1°C to 0.80 centipoises at 30°C. Viscosity affects the shear stress exerted by the water and the settling velocity of particles. The algorithm of shear stress may be written:

$$\tau = (\mu + \eta) \frac{du}{dy} \dots\dots\dots (4)$$

where τ = shear stress, μ = dynamic viscosity, η = eddy viscosity and $\frac{du}{dy}$ = vertical velocity gradient. In laminar sheet flow, $\eta = 0$. A high temperature will then result in low total viscosity, as μ has a low value.

In rills and concentrated sheet flow, erosion and transport processes are also affected by temperature, but the effects are complex. The main flow is generally turbulent but there is a laminar sublayer. This sublayer tends to become thinner at higher temperatures. The velocity gradient then steepens, inducing higher bottom shear stresses. The tendency for detachment of particles should therefore increase with increasing temperature. Grissinger (1966) found that in flume studies of cohesive soils, with a flow velocity of 55 cm/s and a depth of 2 cm, greater erosion rates were associated with an increase in the temperature of the eroding water.

With increasing temperatures the dynamic viscosity decreases, decreasing

its dampening effect on turbulence. Thus turbulence and the eddy viscosity η increase with temperature. Suspended sediment transport is facilitated with an increase in the eddy viscosity. However, as the settling velocity also increases with decreasing dynamic viscosity at higher temperatures, the net effect is not evident. The greatest temperature effect on settling velocity is found for sediment in the size range 0.1 to 0.4 mm (Colby 1963). Conflicting ideas have been advanced concerning the relation between settling velocity and sediment concentration. Some researchers maintain that settling velocity increases with an increase in sediment concentration (Loyacano 1967, referred to in Coleman 1970), while others maintain that it decreases (Vanoni *et al.* 1961, Colby 1963). The net results of all these temperature influences on rill and channel flow erosion and transport are ambiguous. Flume studies with shallow water depths have resulted in both decreased and increased sediment discharges for given variations in temperature, depending upon grain size distribution and flow regime (Vanoni 1962, Grissinger 1966, Taylor & Vanoni 1972). However, the tendency towards a decreased discharge of coarse particles and an increased discharge of small particle sizes with increasing temperature, seems to apply generally.

3.2.3 Effects on soil erodibility

In a climate with a high evapotranspiration rate, the surface soil dries out quickly after a shower. Any surface crustal layer cracks, and larger aggregates are broken up. This disaggregation makes the soil more susceptible to erosion.

The dryness of a soil affects the resistance of the soil structure to destruction at water entry into the soil. Wetting of a dry cohesive soil induces an additional breakdown effect from slaking by en-

trapped air. Wetting also induces swelling of the surface of a cohesive soil. This swelling due to osmosis is greatest in areas where precipitation is smaller than potential evapotranspiration and the concentration of pore salt is high. The swelling reduces the interparticle bonding and, thus, detachability is increased (cf. section 5.1.3).

In an experimental study, sediment samples with different antecedent water contents were studied in a laboratory flume (Grissinger 1972). Runs with flows of a 2 cm depth of water showed that saturated, cohesive soils were less eroded than non-saturated soils. An exception was very dry soils, i.e. soils with an antecedent moisture content of less than 5 % (cf. Figs 1, 3, 4, 5, in Grissinger 1972). The small amount of erosion associated with the very dry soil samples might be explained by the short test period of only a few minutes.

In arid areas, high potential evaporation promotes intensive capillary water movement, which in turn accentuates the surface salinity of lowlying areas and the sparse nature of the vegetation cover (Schattner 1962). The extremely small organic content also makes a soil more erodible.

Local variations in exposure affect evapotranspiration and cause local variations in dryness and vegetation. In the northern hemisphere southfacing slopes receive more insolation than northfacing slopes. The soil is thus drier than on northfacing slopes, and in regions with limited precipitation there is consequently poorer vegetation. There will also be less organic material in the soil and consequently less aggregated soil. Dryness also affects the soil's erodibility as discussed above.

The availability of material for stream transport is also dependent upon the mass wasting processes. Creep increases as the number of dry-wet cycles increases. Rapid mass movements are in-

fluenced by soil moisture conditions. Oversaturation and dryness also facilitate rapid mass movements. A completely dry soil has a zero (atmospheric) pore pressure. In the Lower Jordan Valley, a high percentage of river load is derived from mass wasting, which comes mainly from thoroughly desiccated slope material, because the steep slopes are prone to collapse with the slightest increase in volume and weight caused by wetting (Schattner 1962).

3.3 Frost and snow

3.3.1 *Freeze-thaw processes and soil characteristics*

Outside the tropical zone, most areas are affected by seasonal or short duration freeze-thaw cycles (Cooke & Doornkamp 1974). High erosion rates in, e.g., the southeastern United States have been attributed to intermittent freezing and thawing throughout the winter of wet soil surfaces (Sediment. Eng. 1975).

At freezing, ice crystals grow and water is drawn to them from the surrounding soil. This growth of ice crystals in soil pores and the simultaneous drying of the adjacent soil causes cracking and loosening of the dry soil. The soil surface is raised by the growth of ice. The resultant disintegration of the soil surface makes it highly permeable. Local variations in exposure and, consequently, in moisture and temperature may cause differences in the effect of the freeze-thaw cycle (Diseker & Richardson 1962).

In forested areas in temperate or polar climates, where sediment yields are normally quite low, the main source of sediment may be bank material that has been loosened by frost action (Sediment. Eng. 1975). The principal effects of frost action on cohesive riverbanks may be to widen lines of weakness in the bank sediment and to disaggregate the surface material (Sundborg & Norrman 1963,

Knighton 1973). Coarse, water stable aggregates, especially those greater than 0.84 mm, tend to break down while fine, water stable particles, especially those less than 0.02 mm, tend to aggregate under repeated freezing and thawing (Chapil 1953).

Besides loosening and breaking up the soil surface, frost heave causes the larger stones in heterogenous sediments to migrate upwards to the surface. Concentrations of stones at the surface make the soil less erodible (section 5.1.2).

3.3.2 *Erosional conditions*

In many D climates most erosion occurs during the spring snowmelt. The soil is usually saturated and completely or partly frozen, thus generating plentiful surface runoff. As meltwater overland flow does not possess the kinetic energy of falling rain, the sediment concentration resulting from snowmelt runoff is usually much less than that resulting from summer rainfall under otherwise similar conditions. However, when only the top layer of the soil is thawed, the erosion and sediment transport may be quite high. In ravines near a hydrologic station in the Ukraine, meltwater runoff caused most erosion at the end of the melting period, because the top layer was not frozen then and the soil was therefore easily eroded (Drozd *et al.* 1974).

Erosion may be extremely severe if rainfall occurs when only the top layer is thawed (Flaxman & Hobba 1955). In an area to the west of the Rocky Mountains in the USA, chinook winds as warm as 16°C and often accompanied by rain can cause rapid melting of the snow and considerable erosion (Mapes 1969).

In mountain areas, sudden melting may generate slush avalanches, debris flows, solifluction and slope wash. Slides are reported to be common during spring

thaws in many mountainous regions, e.g. in a basin in the Rif Mountains of Morocco (Akallal 1966), in Norway (Common 1966), and in Sweden (Rapp & Strömquist 1979).

In periglacial environments many of the smaller streams flow only during a short summer period. A basin in the Canadian high Arctic is an example of the erosional conditions and processes in this climatic environment (Cook 1967). During periods of intense melting, meltwater collects on the valley floor, held back by an ice-jam at the river mouth. Flooding follows the breaking of the ice-jam, transporting material that has accumulated through frost weathering, solifluction and other forms of mass movements. Material that has been deposited as alluvial fans appearing after one to two rains a year is re-eroded and transported.

3.4 Seasonality

3.4.1 *General remarks*

Most climates exhibit seasonal variations in precipitation, temperature, humidity, snow, frost activity and vegetation cover. A rather complete picture of the types of seasonal changes, with an emphasis upon changes in processes is given by Wilson (1973a). According to Wilson, the equatorial and middle latitude equable climatic regimes of Strahler's classification have equable temperature and precipitation the entire year and belong to only one climate process system. Other climatic regimes are strongly seasonal, and involve more than one climate process system. The tropical and middle latitude wet-dry and Mediterranean climates, e.g., show a pronounced seasonality in precipitation.

The tropical wet-dry regimes involve annual oscillations between selva systems during the wet season to arid or semi-arid systems during the dry season.

The Mediterranean regime involves seasonal oscillations between humid temperate to semi-arid or arid climate process systems. A change from semi-arid to more humid conditions always implies increased erosion. The continental climatic regime involves large seasonal temperature variations and produces an annual oscillation from periglacial conditions during the winter to semi-arid and humid temperate conditions in the summer.

Seasonality is also reflected in the river regime in moderately sized basins (Pardé 1961, Beckinsale 1969). Three principal factors affecting a river's regime vary seasonally: precipitation, evaporation and vegetation.

3.4.2 *Water erosion in climates with seasonal aridity*

After a drought, the vegetation ground cover is sparse. If heavy storms occur at the onset of the wet season, splash erosion will be great and compaction of the soil may lead to considerable runoff. During periods when the vegetation is in equilibrium with higher rainfall, there is less erosion. This cyclic development has been reported to occur in the seasonally wet tropical climates of West Africa, Rhodesia and Madagascar (Williams 1969). Runoff increases as the wet season advances. Soil loss decreases largely owing to the growth of a dense groundcover. Similar results are reported from Cyprus, which has a Mediterranean climate (Konteatis 1974). The floods following the dry months contain a considerable amount of suspended material, while higher discharge later in the winter may transport very little sediment.

Monsoon climates and tropical savanna climates have hot, rainy seasons, while Mediterranean climates have cold rainy seasons. There are, therefore,

considerable differences in evapotranspiration conditions, the effects of which are discussed in connection with temperature.

3.4.3 A seasonality index

Fournier (1960) found that there was no good correlation between annual precipitation and sediment yield. The regions which experienced most erosion were those in which precipitation was concentrated in a short period of the year. The index that best fitted his sediment yield data was p_m^2/P , where p_m is the rainfall in the month with the greatest rainfall and P is the mean annual rainfall. This index can also be written $(p_m/P)^2 P$. If all the precipitation occurs during one month, $(p_m/P)^2$ is 1. If the precipitation tends to occur uniformly over the year, the ratio tends towards $1/144$ or 0.007. This ratio of the seasonal asymmetry of precipitation is multiplied by the mean annual precipitation.

The p_m^2/P index has the same value regardless of whether the precipitation falls in the cold season or in the hot season, or if it falls as rain or snow. There may also be two rainy seasons in a year in a single climatic type with dry months and high evaporation in between. A similar index based on runoff instead of precipitation might include the seasonal variation in temperature and vegetation.

3.5 Wind

3.5.1 Influences on erosivity

Wind imparts an extra force to the impact of raindrops, as there is an added

lateral component of velocity, increasing the resultant vector. Wind also influences the distribution and the amount of precipitation. Ahmad & Breckner (1974) attributed the variation in plot erosion in Tobago, West Indies to slope and wind effects on the amount of rain. If rain hits the ground surface at an inclination of 20° , the amount of rain falling on a plot is $\sin 20^\circ = 0.342$ times the amount of rain the plot would receive if the rain falls at a right angle to the plot. Wind also causes snowdrift and an uneven distribution of snow and meltwater within a basin.

3.5.2 Influences on erodibility

Wind is an important erosive agent when the vegetative cover is sparse and the soil is dry. A windstorm in Kansas in 1950 removed about 2 cm of the topsoil (McGinnies *et al.* 1963). Deposited wind-transported material adds to the material that is available for water erosion. Wind also helps to dry out the topsoil between showers. This is true even in frozen soil.

The most serious and significant result of wind erosion is the influence on soil texture. The sorting action not only removes the most important material, from the standpoint of productivity and water retention, but also leaves a more sandy soil than the original (FAO 1960). Eventually the generation of a surface armour of larger particles leads to decreasing water erosion. It is difficult to restore an eroded topsoil in arid lands as the soil-forming processes are slow and as it is difficult to reestablish vegetation.

4 Relief and area

4.1 Introduction

This chapter contains a discussion of the effects of slope steepness and slope length on splash erosion and on overland flow characteristics. The discussion is focused on the amount, depth and velocity of overland flow and the detachment of particles by overland flow. The influence of the areal distribution of slope inclinations on erosion and deposition conditions within drainage basins is also considered. The influence of slope on the extension of Horton overland flow is not the same as the influence on saturation overland flow. As has already been mentioned, the size of a drainage basin has an effect on sediment yield (chapter 1). The reasons for this are discussed, and a short account of some relief and area parameters is included in the chapter.

4.2 Slope inclination and slope length

Slope inclination influences the amount of infiltration and surface runoff. The greater the slope gradient, the greater is the portion of the rain that runs off as overland flow (Kazó 1969, Góczán 1972, Konteatis 1974). As more water runs off, overland flow has a tendency to concentrate, and the velocity is increased both as a consequence of the steepness of the slope and the greater depth of a concentrated flow. As velocity increases, the shear stress increases and thus also the erosion potential of the slope. However, in a vertical rainfall the amount of rain per unit area decreases with slope steepness by a factor $\cos \alpha$ where α is the slope angle. This effect is considerable for very steep slopes.

The precise effect of inclination on splash detachment is not quite clear.

However, the depth of flow is of significance in determining the optimum conditions for maximum splash (Mutchler & Young 1975). The optimum conditions for splash detachment therefore vary with rainfall intensity and soil type. The effect of inclination on interrill soil losses seems to be small (Meyer *et al.* 1975). Splash erosion may double with a steepness change from 2 to 20 %, while erosion by concentrated overland flow may be 20 times greater with the same change of inclination (Meyer *et al.* 1975).

Slope length. The water discharge per unit width perpendicular to the flow direction and thus the water depth as well tend to increase with the length of overland flow. Surface runoff from the upslope segments is added to the water from lower sections, thus increasing the discharge and depth of flow. Velocity increases with increasing depth, an increase that can be approximately estimated by the Manning formula (Simons 1969, after Manning 1889). The exact influence of slope length on erosion depends on the soil, vegetation cover, rain intensity and slope steepness.

4.3 Topography of drainage basins

Drainage basins show a great variation in the distribution of slope steepness. Steep slopes imply greater erosion potential and transport capacity. In a highly dissected region, with steep slopes, there will probably be a high delivery rate of eroded material to the streams. However, if the ground flattens before the sediment loaded water reaches the main waterways, deposition will occur.

Researchers such as Corbel, Fournier and Young have distinguished between rivers with high and low relief. Corbel (1959) divided rivers into plain rivers

having a mean gradient less than 1 ‰ and mountain rivers having a mean gradient greater than 1‰ when he compared sediment yields from different climates. His data show that plain rivers yield 30 tons/km²/yr (12 m³/km²/yr) and mountain rivers 558 tons/km²/yr (223 m³/km²/yr). Fournier (1960) also had to divide the rivers into high and low relief classes to obtain good correlations with sediment yield (Fig. 17). Young (1972) considered two relief classes: "normal relief" which included plains, moderately dissected areas and gentle to moderate slopes; "steep relief" which included mountainous areas and individual steep slopes. He found that the median value of ground lowering was 122 tons/km²/yr (46 m³/km²/yr) for "normal relief" and 1325 tons/km²/yr (500 m³/km²/yr) for "steep relief".

Mountain rivers carry higher sediment yields than plain rivers not only because of steeper slope gradients, but also because of heavier precipitation, mass-movement activity and intense channel erosion.

Topography affects the two types of overland flow differently. In semi-arid areas, where Horton overland flow is a fairly common phenomenon (Dunne & Leopold 1978, Kirkby 1969), topography seems to have slight influence on the overland flow production (Kirkby 1978). However, the distribution of flatland will determine whether deposition of material occurs before it reaches waterways. Most humid regions may experience saturation overland flow (Dunne & Leopold 1978), and topography is of importance in determining where overland flow will occur. Gentle concave footslopes and wide valley bottoms allow a greater areal extent of saturation overland flow than steep straight slopes (Dunne & Leopold 1978). Convergence of contour lines provides greater opportunities for overland flow to develop (Kirkby & Chorley 1967).

4.4 The size of the drainage basin

There are many reasons why the size of a drainage basin influences sediment yield. Some of the main reasons are:

1. Differences in the intensity and extension of individual rain storms.

Intense rainfall, particularly of the convective type, is usually restricted to a comparatively small area. The probability of an intense storm covering the whole of a river basin decreases with increasing basin size. Therefore, the maximum flood discharge per unit area is inversely related to the size of the drainage area (Chorley 1969), and the capacity of the water flow to evacuate material which has been deposited in a channel after local rainstorms is therefore smaller for large basins. Some of the material may be deposited within the river basin but outside the storm area.

2. Differences in slope gradient.

Small drainage basins often exhibit steeper slopes and steeper stream gradients than large river basins, which often have considerable areas with low slope and stream gradients and low channel density in the downstream parts. This means that material may be deposited before it reaches the main waterways. An important factor is also the overbank deposition on floodplains at high water discharges.

Hydrological or erosional investigations often deal with small catchments in the headwater regions of larger river basins. This bias is probably one reason for the over-representation of steep slopes in small basins.

Bogolyubova & Karaushev (1974) found that the ratio of sediment yield from a small river basin (<2000 km²) to that of zonal river basins (2000-50 000 km²) within a region was proportional to an exponential function of the ratio of their respective stream gradients. The equation was:

$$M_s/M_z = k (I_s/I_z)^a \dots\dots\dots (5)$$

where

M_s = sediment discharge in small rivers, tons/km²/yr

M_z = sediment discharge in zonal rivers

I_s = average gradient of a small river

I_z = average gradient of zonal rivers

The coefficient k varied between the two regions investigated. The exponent a was found to be 1.6 for both regions.

3. Differences in the influence of environmental factors

Within a large drainage basin, there are often great local variations in climate, soil, vegetation, cultivation, settled areas, constructions such as roads, etc. Some parts of the basin may contribute much sediment and others only very little. If the basin is small, one single factor can dominate and give rise to a very high sediment yield per unit area. On the contrary, small basins sometimes yield less sediment per unit area than bigger catchments. The range of sediment yield diminishes with an increase in catchment size (Schwartz & Pullen 1966). However, most graphs show a decreasing sediment yield per unit area with an increase in basin size.

4. Water losses in ephemeral streams. Solution losses.

In sandy ephemeral channels water losses through infiltration may cause a continual increase in sediment concentration. This will ultimately reduce the transport capacity of the flow and promote deposition. Sediment movement out of large basins is partially prevented (Schumm & Hadley 1961).

To my knowledge there have been no investigations concerning how much material is dissolved when sediment is stored in channels or when it is carried by the water flow over long distances. Research should be directed towards studies within this field.

4.5 Relief and area parameters

A large number of investigations have been performed with the purpose of studying soil loss from plots and from basins of different sizes. Several different parameters of relief and area have been related to soil loss or the delivery of eroded material to streams.

In *plot* studies, power functions of the slope gradient have been used in several cases, e.g. $E \propto S^a$ where E = soil loss per unit area, S = gradient in %. However, as could be expected, different authors have obtained different values of the exponent 'a'. Musgrave (1947) obtained $a = 1.35$; Zingg (1940) $a = 1.49$; Hudson & Jackson (Hudson 1971), $a = 2$ for tropical conditions; and Kilinc & Richardson (1973), $a = 1.66$ in a laboratory plot.

Another proposed power function is $E \propto b + cS^a$ where b and c are coefficients. Smith & Whitt (1947, referred to in Smith & Wischmeier 1962) found a value of $a = 4/3$. A third type is $E \propto (S - S_c)^a$ where S_c = critical steepness below which no appreciable erosion occurs and 'a' varies between 2 and 2.5 (Meyer & Monke 1965). Wischmeier *et al.* (1958) used data from plots with 3–22 % slope gradients and obtained the best results from $E \propto (0.43 + 0.30S + 0.043S^2)$. This function is used in the Universal Soil Loss Equation. In a study on 900 cm² pans, Bryan (1979) also obtained the best results with polynomial relationships because soil loss increased at a lower rate at very high angles.

Horton's equation for the rate of erosion (Horton 1945) is in this study simplified to:

$$e = k (X^{0.6} - X_c^{0.6}) \sin \alpha / \tan^{0.3} \alpha \dots\dots (6)$$

where

e = depth of soil removed at a given point per unit of time

X = distance from the crest

X_c = "critical distance", where no erosion occurs

- k = a compound parameter dependent upon the runoff/h, soil erodibility and surface roughness
 α = slope angle in degrees

According to the formula, the rate of erosion varies with slope gradient as $\sin \alpha / \tan^{0.3} \alpha$. The value of this expression increases with the slope angle up to 56.8° , when it reaches a maximum of 0.7368. Thereafter, it decreases again for still greater slope angles. It should be pointed out that this maximum is not in agreement with the curve published in Horton's article where probably some mistake has been made (Horton 1945, p 321). The rate of erosion also varies with the 0.6 power of the distance from the crest minus the 0.6 power of the "critical distance". The 0.6 power of the length is used by Zingg (1940) and the 0.5 power is used in the Universal Soil Loss Equation in most cases (Wischmeier & Smith 1965).

With most *drainage basins*, it is difficult to define a parameter which takes the probability of deposition into account. One parameter that might reflect such processes is the mean gradient of one selected high stream order (Stall & Bartelli 1959). Another possibility might be to estimate the proportion of land that is covered with water during floods.

Mean basin slope has also been correlated with sediment yield. The calculation technique is found in Wood & Snell (1960) and Stocking (1972). Stocking (1972) determined which of eight relief parameters provided the best correlation with erosion in three tribal lands in Rhodesia. The parameters were correlated with gully erosion, as previous experience in Rhodesia indicated that there is a high correlation between the degree of gully erosion and the degree of sheet and rill erosion. In all three areas, one parameter, average slope, was found to be significantly correlated with the amount of erosion. One problem with this parameter is that the same value is

obtained regardless of whether the headwaters or the areas further downstream, nearer the main channel, are steep. Therefore, it should not be used in a study where basins with different types of topography, e.g. lowland with isolated mountains and high plateaus with deep valleys, are compared.

The mean gradient of the main river, i.e. the ratio of the difference in elevation between the highest and the lowest point on the river and the length of the main river, is easy to calculate and is often used in multiple regressions in large basins (cf. Jansen & Painter 1974, Lisitsyna & Alexandrova 1972).

The relief/length ratio is a commonly used index. The relief/length ratio is defined as the relative relief height between the highest and lowest point of the drainage basin divided by basin length, measured in a straight line approximately parallel to the major drainage channel. In basins where isolated high points on the divide tend to give undue weight to basin height, the relief is calculated with reference to the average elevation of the divide. The same criticism applies to this index as to the mean basin slope.

The relief/length ratio has been used to estimate both the sediment delivery ratio and the sediment yield. The sediment delivery ratio is the percentage of gross erosion that is delivered to streams and transported to a gauging station. The relief/length ratio has explained much of the variation both in the sediment delivery ratio (Maner 1958, Roehl 1962) and in sediment yield (Schumm & Hadley 1961, Douglas 1973). However, in other studies, sediment yield has not been found to be significantly correlated with the relief/length ratio (Stall & Bartelli 1959). In basins which contain two distinct types of topography, the relief/length ratio is not a satisfactory variable (Hadley & Schumm 1961).

The relief/length ratio, the mean basin slope and the main stream gradient are

Table 5. Relations between sediment yield and catchment area. SAY=sediment yield in m³/km²/yr, Y=sediment yield in tons/km²/yr, X=area in km².

Source	Position	Basin sizes km ²	No. of basins	Equation
Scott <i>et al.</i> 1968	Southern California	8-1036	8	SAY=1801 X ^{-0.215} (7)
Fleming 1969	Mostly American, African and UK basins		235	Y= 140 X ^{-0.0424} (8)
Sediment. Eng. 1975	Upper Mississippi River. 'a' varies with 'land resource areas'			SAY=a X ^{-0.125} (9)
Strand 1975	Southwestern USA		8	SAY=1421 X ^{-0.229} . . . (10)
Equations for the upper limits of the silting rate in reservoirs				
Khosla 1953	Different parts of the world	<2590	89	SAY=3225 X ^{-0.28} (11)
Joglekar 1960	95 American, 24 Indian, 10 European, 5 Australian and 5 African basins		139	SAY=5982 X ^{-0.24} (12)
Varshney 1970	Northern Indian mountain basins (north of Vindhya)	<130		SAY=3950 X ^{-0.311} . . . (13)
"	Northern Indian plain	<130		SAY=3920 X ^{-0.292} . . . (14)
"	Northern India	>130		SAY=15340 X ^{-0.264} . . (15)
"	Southern India	<130		SAY=4600 X ^{-0.468} . . . (16)
"	Southern India	>130		SAY=2770 X ^{-0.194} . . .

the most widely used topographic parameters in sediment yield studies. Other topographic or basin shape parameters are discussed in works by Melton (1957), Fournier (1960) and Wood & Snell (1960).

Area, like the relief/length ratio, is commonly used in correlations with the sediment delivery ratio. In some studies, the size of a drainage basin explained a great part of the variation in sediment delivery ratio (Roehl 1962, Renfro 1975, after Maner 1962, William & Berndt 1972), but in other studies, basin size was poorly correlated with the sediment delivery ratio (Glymph 1954a, Maner 1958, Piest *et al.* 1975, after Beer *et al.* 1966). Many other variables, especially land use may conceal an area-delivery correlation. The same holds true for correlations with sediment yield as the dependent variable.

The scatter in the correlations between delivery ratio and area, when many dif-

ferent regions are considered, is derived from differences in the influence of other variables such as soils, climate, land use and topography. In the USA, the exponent in equations which relate sediment delivery ratios to drainage area generally varies between -0.14 and -0.34.

The results of some studies on correlations between sediment yield and basin size are shown in Table 5. As can be seen from Table 5, the sediment yield varies with the -0.04 to -0.47 power of the drainage area. The Indian equations describe the upper limits of the silting rate in reservoirs. Fleming's exponent is very close to zero. One reason for this is that his records were obtained from larger catchments. Fleming (1969) also excludes 18 Indian basins because they strongly depart from his UK, African and American data. Two of the three Chinese basins and the few Swedish basins lie outside the 95 % confidence limit.

5 Soil and rock type

5.1 Soil characteristics

5.1.1 Introductory remarks

The susceptibility of a soil to erosion depends on many factors: infiltration characteristics, detachability by raindrop impact and overland flow, transportability of material, and susceptibility to rilling, gully and channel erosion. In addition susceptibility to creep and rapid mass movements has a bearing on the availability of material for water erosion. An article by Musgrave (1947) implied that sheet and rill erosion in the USA varies by a factor of 32 because of variations in soil erodibility. The effects of different soil properties on erodibility will be discussed in the sections which follow.

5.1.2 Texture

Erosion by raindrop impact is affected by texture. Ekern (1953) obtained the relations presented in Table 6 between particle size and relative amount of splashed material. Ekern's results are in agreement with Ellison's (1947) conclusions that sandy soils splash more readily than do clay soils, and that they do not tend to compact and consolidate as readily as clay soils.

Coarse material such as gravel is not

easily detached by splash or overland flow. Consequently, it protects the finer particles from raindrop impact (Young 1976). This protective effect is especially important in dry areas where vegetative protection is low and on cultivated land (Epstein *et al.* 1966).

Erosion by overland flow is also dependent on soil texture. A graphical relationship between the particle size and the critical erosion velocity of uniform material in a channel shows that particles in the 0.2–0.8 mm range (Hjulström 1935, 1939), or the 0.06–0.5 mm range (Sundborg 1956) are most easily eroded. Critical erosion velocities may not be exactly the same in thin overland flow and rill flow as in channel flow. Particle size distribution is seldom uniform. As particle size distribution affects flow conditions, critical erosion velocities in non-uniform material are less well-defined than in uniform material.

Two main forces resist detachment, namely friction and cohesion (Fig. 8). The resistance to erosion afforded by small grain sizes is due primarily to the effects of cohesion. Basically, attraction between particles is due to negatively charged clay particles attracting positive solute ions and water on their surfaces to form a strongly bonded solid-fluid system (Stratham 1977). Some cohesion may result from negative pore-water pressures, which tend to draw the grains together by a suction effect (Stratham 1977).

In Sweden, gullies are often found where silt borders on another soil because silt is more prone to erosion (Kihlbom 1970). Many secondary, unpaved roads in the loess areas in the USA have eroded into gullies and waterways (Sediment. Eng. 1975). This phenomenon is also observed in similar soils in Poland and in China.

Table 6. Relation between particle size and the relative amount of material splashed from targets (from Ekern 1953)

Grain size interval, mm	Relative amount of splashed material during 5 minutes, %
0.84–0.59	30.0
0.42–0.25	77.2
0.25–0.175	100.0
0.10–0.05	61.0
0.05–0.002	21.0

With non-cohesive material, laboratory tests show that erosion increases with decreasing particle size (Meyer & Monke 1965); cf. Fig. 8.

When *transport* is considered, cohesion is no longer a resisting force and is effective only in cases of aggregate transportation and flocculation. The smaller the particles, the greater is their transportability by overland and channel flow. Even the smallest overland flows may be capable of transporting silt and clay for long distances in suspension (Walker 1980). In cohesive soils, erosion is normally limited by detachment and in friction soils it is limited by transport. If a soil possesses a low transportability, an increase in the slope may cause a considerable increase in the transportation processes (Ellison 1947).

Infiltration rate is a function of texture (Table 7). Therefore, clay soils often generate overland flow, while sandy soils do not. There are, however, other

Table 7. Variation in infiltration capacity with texture (from Morgan 1969, after Konke & Bertrand)

Soil texture	Infiltration capacity mm/h
clay loam	2.5–5.1
silt loam	7.6–15.2
loam	12.7–25.4
loamy sand	25.4–50.8

factors that compensate for the influence of texture on permeability and infiltration (cf. 5.1.4). In conclusion it can be stated that a high clay content induces overland flow and increases transportability but decreases detachability.

5.1.3 Structure

The soil structure is the arrangement of the soil particles and of the pore spaces between them. Soil particles are held together in aggregates. There are many investigations concerning the stability of aggregates (Browning 1937, Chepil 1955a, b, Kemper & Koch 1966, Anderson & Wenhardt 1966, Emerson 1967, Bryan 1968, 1979, Sorozan & Lawrence 1974, Henin 1976, Greenland 1977). Other researchers have studied how different chemical characteristics of soils affect erosion (Grissinger 1966, 1972, Arulanandan & Heinzen 1977).

The content of clay, organic matter, Ca, Mg and free iron oxide contributes to the aggregation of particles. Exchangeable sodium, Na, is a dominant factor in reducing aggregate stability.

Aggregation increases the number of large pores and thus increases the infiltration rate. Dispersion causes conditions which help to clog pores and to form a surface seal which facilitates overland flow. Organic matter in the soil is of importance because of its ability to produce a soil with an open, porous but water-retentive structure (FAO 1976a).

Disintegration of the soil structure may be caused by both mechanical and

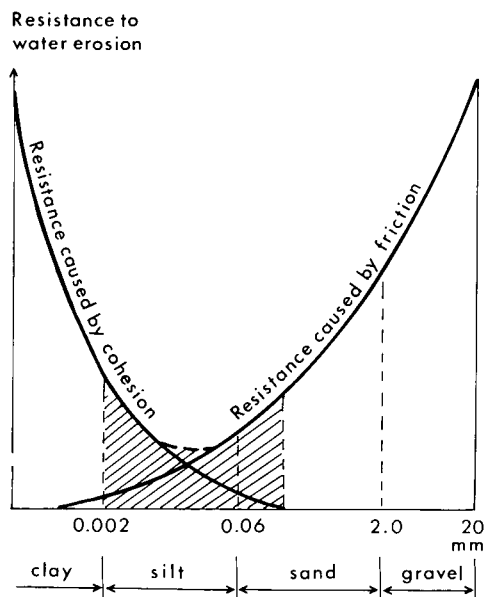


Fig. 8. A schematic graphic presentation of the relative resistance of different grain sizes to erosion by water. The graph cannot be used for quantitative estimations (from Kihlbom 1970).

natural agencies. On cultivated land, soil disintegration often occurs through mechanical agencies. On both cultivated and non-cultivated land, drying and wetting and freeze-thaw are important processes in disintegration. In periods between rains, the thin surface crust and large aggregates dry and crack, disintegrating into smaller aggregates. These may be stable but more easily detached. This is a most important process in the preparation of the surface soil for erosion.

As has already been pointed out, rain induces mechanical destruction of the soil aggregates. Two other stresses may also be involved in the disintegration of the aggregates at water entry into a dry soil: slaking and differential swelling. Aggregates may slake due to pressures exerted by entrapped air (Emerson 1967, Henin 1976). Differential swelling may be caused by rapid wetting occurring in one direction or in the outer parts of the aggregates or by osmotic phenomena (Emerson 1967, Henin 1976, Arulanadan & Heinzen 1977). Such swelling would reduce the interparticle bonding forces. Bowie *et al.* (1975) concluded that the susceptibility of most impermeable material to erosion is less in continuously wet soils than in soils which are alternately wet and dry.

For common agricultural soils, soil particles detached by raindrops are largely aggregates with densities close to 2.0. Their diameters range from approximately 0.002 to 2.0 mm with the majority being larger than 0.5 mm (Foster & Meyer 1975).

5.1.4 Permeability

The effect of texture on infiltration rates is shown in Table 7. Large pore sizes are decisive in determining permeability differences within a soil profile (Arnett 1976). Aggregation, vegetation and soil fauna as well as grain size distribution,

are therefore of importance in determining permeability. In cohesive soils, the large pores are normally interaggregate pores. If the aggregates are stable, the interaggregate pores will also persist. Permeability is often highest in the A horizon due to the organic content.

Some clay soils are subject to cracking when they are dry and this encourages initial infiltration, often to great depths. It may also lead to piping and gullyng of adjacent soils, where subsurface channeling of runoff water through the cracking network emerges (Virgo & Munro 1978). As wetting proceeds and the cracks close, infiltration is reduced to a very low rate in these soils.

5.1.5 Erodibility indices

There are many erodibility indices presented in the literature. Some are based on texture; others on structure; while a third group is based on a combination of water transmission and aggregate stability or dispersion. An extensive presentation and discussion of erodibility indices can be found in an article by Bryan (1968). Other works include Dumas 1965, Barnett & Rogers 1966, Fournier 1967, Lugo-López 1969, Yamamoto & Anderson 1973, Bollinne & Hanotiaux 1975, Bryan 1977, Tienseemuang & Ponsana 1977. Only three indices are treated below.

It is difficult to derive an erodibility index that is valid for both sandy and clayey soils, for soils developed on calcareous as well as non-calcareous parent material, and for soils in humid tropical, humid temperate and dry climates.

The dispersion ratio is the ratio of dispersed silt and clay to all silt and clay. However, if the silt and clay fraction in the soil is small, this ratio does not accurately reflect the erodibility of soils (Bryan 1968).

The surface-aggregation ratio is defined as the ratio of the total surface area

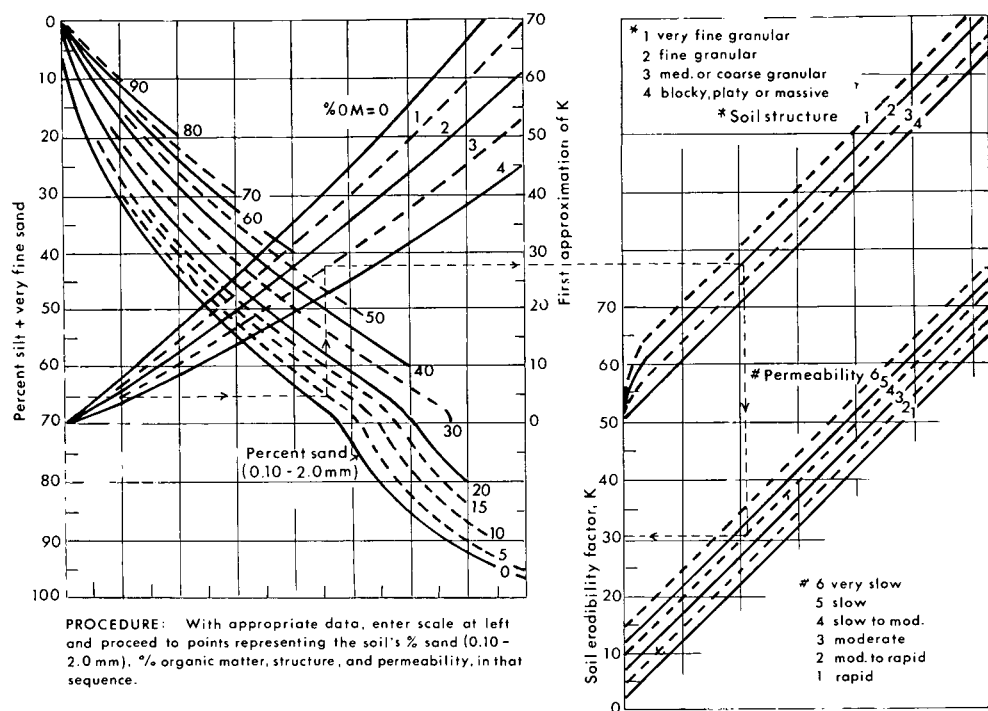


Fig. 9. The soil erodibility nomograph (from Wischmeier et al. 1971).

of particles > 0.05 mm in cm^2/g to aggregated silt and clay in % (Anderson 1954, André & Anderson 1961). This index was intended to show how effectively the soil was bound together. The binding efficiency is represented by the ratio between the amount of soil surface which requires binding and the fraction of the soil which tends to bind the soil together. However, silt is included in the binding fraction, but is usually considered to be highly erodible (Ayres 1936, Wischmeier et al. 1971, Young 1976) and easily transported.

An index which has been much used lately is the soil factor K in the so-called USLE equation (cf. the presentation in chapter 11). Soil loss is measured from a 22.2 m long plot on cultivated continuously fallow land on a 9 % slope, with either simulated or natural rainfall. K is then calculated as soil loss per EI_{30} -unit, i.e. per rainfall energy intensity unit

($\text{tm}/\text{ha} \times \text{cm}/\text{h}/10^2$). K is the mean soil erodibility for the annual rainfall pattern of the site and it consequently includes an appreciable range of storm sizes and antecedent moisture conditions (Wischmeier et al. 1971).

Attempts have been made to correlate other soil parameters with the plot-derived soil factor K . Wischmeier et al. (1971) succeeded in constructing a soil erodibility nomograph based on five parameters: Percentage of silt plus very fine sand, percentage of sand > 0.10 mm, organic matter content, structure and permeability (Fig. 9). Silt and very fine sand accounted for 85 % of the variance in the observed K values for the 55 tested soils. A high silt content gives a high K value (Fig. 9). Sand percentage is inversely related to clay content and increases erodibility. Organic material in the soil decreases erodibility. Permeability refers to the soil profile as a whole.

Soil layers of low permeability within the profile are therefore of importance. In a later study (Arnoldus 1977a, after US Dep. Agric., SCS 1973), K values were adjusted for the gravel content of the top soil.

The nomograph was established for USA conditions, but has also been tested in tropical climates. Preliminary tests of the nomograph on 10 selected Hawaii soils indicated serious discrepancies from plot-measured values, especially for soils with either a low or a high susceptibility to erosion. El-Swaify (1977) thought that a measure of amorphous or oxidic constituents should be added to the index in highly weathered tropical soils. Roose (1977), on the other hand, found that measured K values and those obtained by using the nomograph compared reasonably well for nine soils in the Ivory Coast (with the exception of gravelly soils). In a CaCO_3 rich soil in northern Greece, the K value of the nomograph before the removal of CaCO_3 corresponded best to the plot-measured K value (Silleos 1981).

Dumas (1965) determined the K factor in the USLE in Tunisia with the help of simulated rains. The empirical relationship between the K factor and characteristics of the soil was found to be:

$$\log 1000K = 3.4623 - 0.0282X_1 - 0.1695X_2 - 0.0214X_3 \dots\dots\dots (18)$$

where X_1 = % gravel

X_2 = % organic material

X_3 = moisture equivalent in %

Moisture equivalent is a measure of water content at field capacity. The above equation explained 88 % of the variance in K. The accuracy of the equation was found to be good when the gravel content was greater than 10 %. In a simple correlation, the percentage of the total weight accounted for by gravel in soil samples explained 69 % of the variance.

5.2 Relation between rock type and soil characteristics

5.2.1 Effects on soil texture

Comparatively little research has been undertaken on the general effect of bed-rock on soil characteristics such as texture, aggregation and permeability. It is therefore difficult to draw conclusions as to general differences on a global scale (cf. Leopold *et al.* 1964, chapter 4). Some comments will, however, be presented below.

Eyre (1968) made a number of general observations concerning the relation between rock type and soil. Acid igneous rocks, such as granite, contain a high percentage of quartz, and thus weather into predominantly sandy material. Basic igneous rocks, such as basalt, produce a residue which is predominantly clay. Generally speaking, sedimentary rocks produce a weathered residue, which is more homogeneous than that from igneous rocks. Shales thus weather into a predominantly clayey residue, whereas sandstones and grit produce a sandy residue.

Douglas (1968) has studied the weathering and soil conditions in the hot, humid climate of the Sungey Gombak catchment, Malaysia. His study may be cited as an example of the significance of rock type for soil texture. Gentle granite slopes exhibited extremely deep weathering profiles. The granite weathered into a friable sandy, clay loam with a bimodal grain size distribution. There was a tendency for more sandy residual soils to develop on the top of a slope because colloidal particles were washed out, and for more clayey soils to develop at the base of the slope. Quartzitic rocks yielded thin soils with a high proportion of sand, while shales yielded a clay regolith (Douglas 1968).

5.2.2 Effects on soil aggregation

Surface-aggregation and dispersion ratios are indices of dispersion (cf. 5.1.5). Özyuvaci (1971) reported from the Marmara region, Turkey, that the magnitude of the dispersion ratio decreased in the following manner: crystalline schist soils > arkose soils > granitic soils > neogen soils. André & Anderson (1961) analyzed soils at 168 places in northern California. Lithology was found to be most significant in explaining the variation in surface aggregation and dispersion ratios. A table containing values in surface aggregation and dispersion ratios from André & Anderson's study and from three other investigations (Anderson & Wallis 1963) showed that recent volcanic and acid igneous rocks were highly erodible, while Cenozoic marine sediments and older sedimentary rocks had low erodibility indices. Cenozoic non-marine sediments, schists and metamorphic rocks exhibited medium values.

5.2.3 Effects on infiltration rate

Differences in permeability in soils derived from different parent materials, and differences associated with the different depths at which the soil samples were taken are presented in Table 8. It can also be noted that overland flow was generated spatially uniformly on sandstone soil plots but not on shale soil plots in a semi-arid region of Alberta, Canada. This was attributed to differences in the

composition and structure of the surface material (Bryan *et al.*) 1978).

5.3 Relation between climate and soil characteristics

5.3.1 Introductory remarks

Soil characteristics are controlled by both rock type and climate. Correlations between climate and soil distribution have formed the basis for zonal soil classifications. Many types of zonal soils are associated with specific types of vegetation as well as particular types of climate. Some soil types can be characterized as intra-zonal soils and reflect differences in parent material, relief, biological activity and age of the soils.

In subsequent sections, there is a discussion of variations with climate in several soil properties: clay eluviation, type of clay, crust formation and aggregation. Climatic variation in soil erodibility is then demonstrated by comparing values of the K factor in different climates. There is also some other information on the erodibility of several soil types.

5.3.2 Clay eluviation

The clay content of surface soils affects runoff and detachability (cf. section 5.1.2). Clay is colloidal and de-flocculates readily in strongly acid conditions. Consequently, in acid podzols clay is transported downwards through the profile by percolating water (Eyre 1968).

Table 8. Permeability of soils at different depths and from different parent rocks near Istanbul, Turkey (from Özyuvaci 1971)

Depth cm	Quartzite	Granite	Permeability in cm/h		Clay schist	Crystalline schist
			Arkose	Neogen		
0- 15	116.06	55.81	34.17	15.84	6.39	6.02
15- 40	15.06	6.50	3.35	3.55	1.84	0.72
40-120	4.31	2.01	0.60	0.16	0.40	0.21

The clay and sesquioxides may be redeposited in the illuviated horizon. Because of eluviation, acid soils normally have high surface permeabilities. When humid soils are cultivated, fertilizers and mulch may prevent the soil from becoming acid. In drier areas with $\text{pH} > 7$, the humus may remain stable. Chernozems also have $\text{pH} > 7$, which means that both clay and humus remain stable.

5.3.3 Type of clay

Different types of clay have different capacities for absorbing water within the structure and different cation exchange capacities. The type of clay may therefore influence the erodibility through swelling and slaking (cf. section 5.1.3). Soil types are often associated with certain kinds of clay. Although somewhat of an oversimplification one can say that in tropical regions highly weathered, non-swelling kaolinite minerals, with low exchange capacities, are often found on slopes, and montmorillonite with a highly expanding, water-absorptive lattice and high exchange capacity is found in valleys. Kaolinite is also found in the soils of the subtropical, eastcoast climates of the southeastern USA, central China, eastern Australia and eastern Brazil (Bridges 1970). Montmorillonite predominates in tropical and subtropical black soils, chernozems, prairie soils, "humic gleys" and renzina soils, and illite in podzol, brown forest soils and tundra soils (Ollier 1969). Illite is also present in renzina and desert soils (Leopold *et al.* 1964, p. 105). Illite has a low water absorption capacity.

5.3.4 Crust formation

The presence of a crust in the soil profile restricts water transmission and saturation overland flow is more quickly attained than in soils without a crust. Water movement along lateral percolines

can start gully and slide formation in crust areas (cf. Conserv. Found. & FAO 1954, Temple & Rapp 1972).

In tropical savanna climates, there may be concretions of iron oxide (ferricrete) or duricrusts of silica (silcrete) in the soil profile. Irregular calcium carbonate concretions normally occur in chernozem soils, some 70–100 cm below the surface. Chestnut soils, and both brown and grey semi-desert soils may have concretion horizons nearer the surface. Locally, where rainfall is very low this horizon in light brown and grey semi-desert soils may actually emerge at the surface as a rock-like lime (Ca CO_3) or gypsum (Ca SO_4) crust which may attain considerable thickness (Eyre 1968).

5.3.5 Soil structure

Certain tropical soils are found to have a resistant structure, while others may have a much weaker structure, as they contain iron oxides which do not affect the structure (Greenland 1977). Tropical soils developed on highly basic rocks appear to contain structurally active oxides, probably amorphous iron and aluminium hydroxides. These soils have an exceptional structural stability (Greenland 1977). Ives (1951) found that the high percentage of rather large aggregates was responsible for the high permeability of a soil in Turrialba, Costa Rica. The reason for such a favourable aggregation was, in part, explained by the sesquioxide content. Analysis of deep 'coffee' soils in the Kenyan highlands, developed from basic rocks, revealed very stable microaggregation, which made them feel like loams and sandy loams, even though they contained 50–80 % clay (Ahn 1977).

Bridges (1970) stated that there is little silt in humid tropical soils, but that they possess a variable quartz sand content. Whether this can be regarded as a uni-

versal fact or not is hard to say. If it is true, humid tropical soils would be less erodible than soils in other climates. Douglas (1968) noted that there was a bimodal grain size distribution in samples from soils on granite in the Sungei Gombak catchment, Malaysia. However, Roose (1977) found that ferruginous soils had a relatively high percentage of silt and fine sand, and Fournier (1967) noted that tropical ferruginous soils had an unstable structure once they were cultivated.

Aggregated stability of some tropical soils is in part due to their high organic content (Greenland 1977). In some desert soils, the amount of organic material is low and the cation exchange may be dominated by sodium (Bridges 1970). This means that a soil with easily dispersed aggregates will develop. Aggregation of the soil surface is also dependent on climatic wetting-drying and freeze-thaw processes (cf. sections 3.2.3 and 3.3.1).

5.3.6 Climatic differences in soil erodibility

Values from different sources of the erodibility index K in the Universal Soil Loss Equation for tropical and non-tropical soils have been tabulated in Table 9. This makes possible a comparison between soils from different climates. The higher K values of ferruginous soils on granite compared to ferrallitic soils in the Ivory Coast were explained by their low permeability, their tendency to develop crusts and the relatively high percentage of silt and fine sand (Roose 1977). Ferruginous soils at Sefa have lower K values. Tropical soils usually exhibit low values of the factor K. However, in the East Central state of Nigeria, K values were as high as 0.33 in the major soils. The K values of Ibadan were found to vary in relation to the elapsed time after forest removal and the tillage techniques.

Even outside the tropical zone, erodibility is dependent on the organic matter

Table 9. Values of the factor K (erodibility index) in the Universal Soil Loss Equation

Source	Location	Climate	K value	Comments
Wischmeier & Smith 1965	USA, east of the Rocky mountains	Ca, Da Db	0.03–0.69	3 soils have 0.03–0.10, 10 " " 0.22–0.31, 7 " " 0.32–0.41, 2 " " 0.48–0.69
Hadley (stencil)	Big Horn county, Mont.	B	0.32–0.39	
Lal 1977	Turrialba, Costa Rica	Af	0.103–0.155	
Barnett <i>et al.</i> 1972	Puerto Rico	Af(Am)	0.004, 0.013 0.017, 0.113	
Roose 1977	Ivory Coast		0.05–0.10 0.10–0.15 0.15–0.18 0.20–0.30 0.04–0.17	ferrallitic soil on tertiary sand ferrallitic soil on granite ferrallitic soil on schists ferruginous soil on granite ferruginous soils
Lal 1977, after Charreau 1972	Sefa, Senegal	Aw		
Lal 1977	Ibadan, Nigeria	Af	0.06–0.36	
Lal 1977, after Niger Techno Ltd 1975	East Central State of Nigeria	Af	0.12–0.48	
Lal 1977, after Hasselo & Sikurajapathy 1965	Sri Lanka	Am	0.01, 0.05 0.31, 0.31	

content. In Hungary, erosion is lowest in chernozem soils (Stefanovits 1964, cf. Bobrovitskaya *et al.* 1977 for USSR). When the humus-rich top layer is reduced, erosion increases abruptly. It is greatest on loess regolith. On brown forest soils, erosion is normally low. When the humus-rich top layer is stripped off, erosion is still lower, as the

clay rich illuviation layer, which contains clay and iron colloids, is exposed at the surface. On renzina soils, the thin top soil is easily washed away, and a dense layer then checks erosion. Therefore, in renzina regions, soil loss is small, but surface runoff is high. This has a negative, secondary effect upon neighbouring areas.

6 Vegetation

6.1 Introduction

Vegetation cover strongly reduces erosion. The influences of vegetation on erosional and transportational processes are numerous and varied. Any man-induced changes in vegetation cover will affect these processes. Knowledge of the effects of plant cover is essential for understanding the rate of erosion. Some aspects of the influence of vegetation are briefly considered below.

Following an account of general vegetational effects on erosion, the more specific protective qualities of natural vegetation in different climates are discussed and values for soil loss from plots under forest, grass and crop cover are tabulated. Finally the opinions of some researchers about the degree of the vegetational effects on erosion are presented.

6.2 Effects on soil detachment and transport by rainfall and runoff

The systematic approach applied by Wischmeier (1975) is followed in this section. Wischmeier separated the effects of vegetation on erosion and transport into three distinct types: the

effects of canopy cover, the effects of organic material on the ground and the effects of organic matter in the soil.

Effects of canopy cover: Canopies intercept rain and snow. A relatively large proportion of the precipitation in light storms is prevented from reaching the soil by trees, bushes, herbs and grasses, whereas in heavy storms the proportion is much smaller (For. Serv. and Soil Conserv. Serv. 1940, Kittredge 1948). Water drops re-form and fall from the canopy with a reduced velocity. The reduction in the energy of the drop impact depends on the density of the canopy and its height above the ground.

Effects of organic material on the soil surface: Mulch, leaf litter and close growing vegetation intercept raindrops. There is no remaining fall height to the ground for re-formed water drops. Thus, the area of the soil surface exposed to splash erosion is reduced. Material on the ground also protects the pores from sealing. This results in higher infiltration rates and less surface runoff than on bare soil.

A third effect of this type is the reduction of flow velocity by surface friction. This implies a reduction in shear stress and in the potential of the runoff to detach and transport soil material. This effect is exaggerated, since part of the

shear stress by surface flow is applied to the vegetation instead of to the soil.

The effects of organic matter on surface runoff and erosion have been evaluated by using insect screens or mulch in plot studies (cf. Hudson 1958, 1971 and Tables 2, 10). Meyer *et al.* found that screens reduced interrill erosion on plots by as much as 90 %, and rill erosion, by approximately 60 %.

Effects of organic matter in the soil: Soil aggregation is increased by organic matter, but the magnitude of the effect is related to texture (Young 1976). Thus, when the clay content is high, the effect of organic matter on erodibility is low, but when the clay content is low, the effect of organic matter on erodibility is high. An organic matter content which exceeds 4 % seems to have little additional effect on erodibility (Young 1976, cf Kemper & Koch 1966).

Channels left by decayed roots promote infiltration (Ayres 1936). Biological activities which increase with increased organic matter content have a positive influence on the number of large pores, and thus the infiltration rate (Smith & Wischmeier 1962). Organic matter in the soil affects infiltration and retards soil erosion for a period of several years after clearing and cultivation. Enquiries from farmers in Western Australia showed that "the first signs of soil erosion (=gullies) appeared within five to ten years after clearing" (Teakle 1939). For the first five years after clearance, the physical properties of the soil apparently remained intact, but the conditions deteriorated.

The root system also has a binding effect on the soil (Ayres 1936). Tree roots stabilize the soil on steep slopes, partly because the infiltrated water is carried further down into the profile and partly because the roots keep the soil stable. As a rule, trees help prevent minor landslides.

Table 10. Effect of ground cover on runoff and soil loss from an alfisol near Ibadan, Nigeria, with a 10 % slope. 6 tons/ha of straw mulch was applied (from Lal 1977)

Mean annual values	Bare ground	Mulched
Soil loss (tons/ha/yr)	232.6	0.2
Surface runoff (mm)	504.1	29.3
Runoff (% of rainfall)	42.1	2.4

6.3 The influence of vegetation on water erosion in different climates

Under natural conditions, the broad character of the ground cover is related to precipitation and temperature (Douglas 1975). Young (1972) provided the following list, ranking vegetation communities in order of decreasing protection against rainsplash and erosion by overland flow:

- Temperate deciduous woodland
- Temperate coniferous forest
- Humid temperate grassland
- Sub-tropical rainforest
- Tropical semi-deciduous rainforest
- Tropical evergreen rainforest
- Mediterranean communities
- Savanna woodland
- Tree and shrub savanna
- Steppe
- Thorn scrub and other semi-arid communities
- Tundra and montane communities
- Extreme polar communities
- Desert

In many cases, the list assigns vegetation to climatic zones. Tropical rainforests provide less protection against erosion than temperate forests. Surface runoff and bare soils often occur in tropical rainforests, whereas they are unusual in temperate forests (cf. Young 1972). Some forms of sub-tropical and tropical, semi-deciduous forests as well as tropical montane forests also give almost complete protection to the soil.

In tropical rainforests, the shrub layer is often sparse and the leaf litter thin as a result of the more rapid vegetation decay in the prevailing, higher temperatures, and in the absence of seasonal leaf drop (Ruxton 1967). In the mature primary rainforests of Papua, leaf litter has an average thickness of 20 mm under closed canopy forest on gentle slopes. On hill and mountain slopes, the thickness averages less than 3 mm, and on many steep slopes the ground has only a scattering of recently fallen leaves (Ruxton 1967).

Many of the water drops which are re-formed in the canopy of rainforests experience a free fall of over 8 m and are close to their terminal velocities when they strike the ground (Ruxton 1967). Splash erosion is thus more important under the canopy of rainforests than within temperate forests because of the higher canopy, the sparse shrub layer, the less abundant leaf litter and the greater rainfall. The impact of large water drops from leaves which fall on approximately the same point results in considerable quantities of detached soil which are carried downslope by surface runoff (Douglas 1968). Rain water streaming down tree trunks removes leaf litter, and sometimes has sufficient energy to scour and undermine surface roots on the downslope side of trees.

On the margin of rainforests in eastern Australia, the denser ground cover be-

neath the main tree layer of wet sclerophyll forests, protects the soil more effectively than in adjacent rainforest regions, where the forest floor is subject to surface runoff and soil loss (Douglas 1975, after Plummer 1963).

As can be seen from Young's list and from Table 11, the grass cover in C and D climates provides good protection against soil loss. The conditions are different in savanna, tall grass and short grass prairie communities, where most grasses grow as compact clumps, leaving bare soil between. There is no wholly organic soil horizon except at high altitudes. The drier the climate, the higher the proportion of bare soil. In the savanna, the ground cover is at a minimum at the onset of the rains. Under savanna, in northern Australia, soil movement by surface runoff, at comparable rainfalls, varied seasonally between periods with sparse and dense plant cover, by a factor of 20. However, Table 13 shows that in West Africa, plot erosion from savanna and ungrazed grassland is only 1 % of that from bare ground.

In semi-arid climates, plant growth is severely restricted and there is correspondingly little plant litter. This results in small ground protection. On unstable soils in semi-arid rangeland with sparse vegetative cover there may be substantial sediment transport, even from un-

Table 11. Annual soil losses from forested and grassy plots in the USA (from Bennett 1939)

Soil and location	Slope, %	Rainfall, mm	Period, inclusive	Cover	Soil loss, tons/km ²
Kirvin fine sandy loam, Tyler, Tex.	12.5	1064	1933-35	woods	17.7
	8.7	1029	1933-36	grass	8.1
	16.5	1092	1933-36	grass	1.8
Vernon fine sandy loam, Guthrie, Okla.	5.17	849	1933-36	woods	3.8
	7.7	838	1931-35	grass	6.5
Cecil sandy clay loam, Statesville, N. C.	10.0	1150	1932-35	woods	0.4
	10.0	1150	1932-35	grass	2.7
Muskingum silt loam, Zanesville, Ohio	12.0	926	1934-36	woods	trace
	12.0	926	1934-36	grass	6.9

grazed areas. Hadley & Lusby (1967) reported a 50 % runoff from an ungrazed 0.04 km² basin and a sediment yield of 3362 tons/km² from a single one-inch rainstorm at the medium maximum intensity value (51 mm/h). As can be seen from Table 12, grasses and scrubs can be effective in controlling sheet and rill erosion, even in more arid environments (van Rensburg 1955, Temple 1972). In semi-arid Tanzania, low bushes give better ground protection during high intensity rainstorms than trees (Strömquist & Johansson 1978).

Dry grasses are very difficult to protect from fire, which can cause serious erosion. Grasses do not offer as good protection against rapid mass movements as tree cover, partly because of the lesser binding effects of the roots and partly because of the less deep infiltration.

Bare ground is also exposed where growth is limited by cold and by snow cover, as in tundra regions and high mountain environments above the tree-line. Overland flow can, therefore, occur relatively unhindered following snow-melt. Figure 10 shows the relation between ground cover and erosion in a sub-alpine rangeland, Utah. Once the groundcover has been reduced to below 60 % to 50 %, overland flow and soil losses increase at an extremely rapid rate.

6.4 Some reflections on the importance of vegetation on erosion

Bennett (1939, p. 126) states that "cover affects rates of soil removal and runoff more than any other single factor". The great difference in the rate of erosion is caused by the removal of natural vegetation. Brune (1948) estimated that the rate of sediment production in 1950 in the Ohio and Great Lakes drainage basins

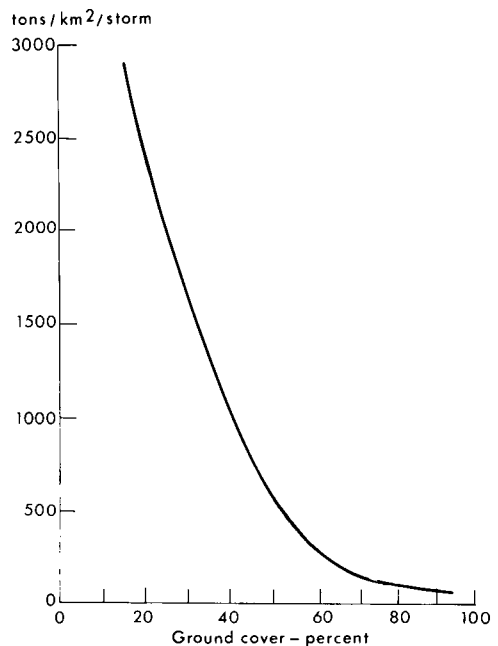


Fig. 10. Relation of eroded soil to ground cover density in sub-alpine-herbaceous rangeland in Utah. Based on simulated rainfall of 38 mm/h for 20 minutes (from Noble 1963).

with about 30 % cultivated or idle land was roughly 50 times the geologic norm. In the upper Mississippi River drainage basin, where about 42 % of the land was cultivated or idle, the rate of sediment production was approximately 75 times the geologic norm. Kittredge (1948, p. 300) states that "fires or other agencies that destroy most of the vegetation and the forest floor increase the discharge of eroded material by 10 to 6000 times".

A compilation has been made of data on soil loss from investigations in different climates and with different covers. These have been analysed and compared in the last column in Table 12. This comparison illustrates the order of magnitude of the difference in soil erosion which results from the removal of natural vegetation. Further discussions and comparisons can be found in chapter 7.

Table 12. Annual soil losses from plots with different vegetation cover

Source	Location	Soil type	Period of record	Annual rainfall mm	Slope %	Cover	Soil loss tons/km ²	Loss ratio, grass crop %
Singh et al. (1967)	Kota	Kota clay 45.5 % clay, 21.0 % silt and 33.5 % sand	1957-60 " " " " " " " " " "	646 " " " " " " " " "	0.5 " " " " 1 " " " "	natural grass cultivated fallow grass groundnut black gram jowar (sorghum) natural grass cultivated fallow grass groundnut black gram jowar	26 349 29 111 206 286 39 372 37 194 208 242	 23.42 12.62 9.09 10.48 20.10 18.75 16.12
Starkel (1972)	Nurpur, Punjab			1092 "	25 "	grass, 80 % without vegetation	545 1026	 53.14
Starkel (1972)	Dehra Dun		1967-68 "	1928 "	8 "	grass maize + wheat	10 130	 7.69
Gorrie (1938)	Sholapur near Bombay		- 1937 -	711 - 711	1.25 1.25 1.25	jowar " clean fallow	28,872 33,391 6277	
van Rensburg (1955)	Mpwapwa, Tanzania		1946-54 "	690 "	6 6	grass cultivated sorghum	65.0 7814	 0.83
Temple (1972)	Mpwapwa, Tanzania	coarse, red sandy friable loam	1933-35 " " "	675;564 " " "	6 8 6 6	grass ungrazed thicket millet bare, weeds removed	0 0 7353 21,923	
Temple (1972)	Tengeru, Tanzania	deep red volcanic soil	1958-60 " " "	1070 " " "	32 32 32 32	grass, cut for hay maize banana, trash mulch coffee, clean weeded	0 1200 50 2240	m ³ /km ² " " "
Hudson (1958)	Henderson research station, Rhodesia	Tatagura yellowish brown clay loam with high silt content	1953-56 3plot yrs 3plot yrs 5plot yrs 5plot yrs 5plot yrs 1953,55/56 3plot yrs 3plot yrs	988 fr. Oct. to March 1138 1151 1279 993 1241	6 4.5 3 6 4.5 3 6 4.5 3	Napier fodder " " Maize " " Sunhemp " "	172 209 52 793 305 322 5313 1344 1196	 21.69 68.52 16.15 3.24 15.55 4.35
Roose (1967)	Sefa, Senegal	grey ferruginous soils	1955-63 1956,1962 1963 1959,1963 1954-63	1138 1151 1279 993 1241	2 1.25-2 1.5 1.5-2 1.25-2	Forest Cotton Maize Millet Groundnuts	7.6 775 1022 1034 689	
Lal (1977 after Marques et al.)	State of Sao Paolo	Sandy bauru, 45 % clay Terra roxa, 15 % clay		1123 " 1292 "	6 6 10 10	corn, downhill corn, contour corn, downhill corn, contour	5280 1980 2140 410	

Source	Location	Soil type	Period of record	Annual rainfall mm	Slope %	Cover	Soil loss tons/km ²	Loss ratio, grass crop %
Fournier (1967)	Adiopo-doumé, Ivory Coast		1956	1939	7	secondary forest	240	
			"	"	7	bare soil	11,770	
			"	"	7	manioc	9280	
			1957	2383	7	secondary forest	3	
			"	"	7	bare soil	10,450	
			"	"	7	manioc	2870	
			1961	2289	7	secondary forest	2	
			"	"	7	bare soil	14,320	
			"	"	7	naturally fallow	46	
			1959-60	978	20	grass, 20 % cover	1200	
			"	"	36	grass, 100 % cover	2.6	
			"	"	7	crops	5900	0.04
			1956-58		7	citrus on bare soil	1890	
			"		7	citrus plus Pueraria	270	
Glymph (1954b)	Lake Alaotra, Madagascar. Kindia, Guinea		1932-42	528	30	perennial grasses bare, untilled	17.9	0.36
			"	"	"		4950	
			1931-41	832	4	Bermuda grass	4.5	0.10
			"	"	"	continuous corn	4613	
			1931-40	1033	8.75	bluegrass	17.9	0.33
			"	"	"	continuous cotton	5396	
			1930-38	778	7.7	Bermuda grass	4.5	0.11
			"	"	"	continuous cotton	4237	
			1933-42	733	9	bluegrass	6.7	0.08
			"	"	"	continuous corn	8586	
			1931-40	749	8	bluegrass	35.9	0.31
			"	"	"	continuous corn	11,417	
			1933-38	859	16	bluegrass	22.4	0.09
			"	"	"	continuous corn	25,039	
Bennett (1939)	Zanesville, Ohio		1934-42	988	12	bluegrass	5.2	0.02
			"	"	"	continuous corn	22,260	
			1931-38	1214	10	mixed grass	69.5	0.99
			"	"	"	continuous cotton	6998	
			1930-35	515	5	native grass	6.5	0.25
			"	"	"	kafir	2632	
			1926-37	527	2	buffalo grass	289.2	18.35
			"	"	"	cotton	1576	
			1947-50	286	1.7	cultivated land	605	
			1947-50	240	1.5	cultivted land	491	
			1947-50	165	1.7	cultivated land	1459	
Basu & Puranik (1951)	Akolner, India		1947-50	286	1.7	cultivated land	605	
			1947-50	240	1.5	cultivted land	491	
			1947-50	165	1.7	cultivated land	1459	

7 Man's activities

7.1 Types of effect

Population increase has brought about and continues to bring about extensive changes in land use. All operations that reduce the vegetation cover may induce accelerated erosion. Reduction in vegetation cover may be caused by cultivation, grazing and burning, lumbering, road construction, urbanization and mining.

A change in the vegetation cover may also increase the chances of rapid mass movements. Mudflows are thought to increase in the mountainous and piedmont regions of the Soviet Union because of tree felling on steep slopes, destruction of the sod cover in the alpine and sub-alpine zones as a result of cattle grazing, and mining and construction of roads, canals and terraces on steep slopes (Gagoshidze & Natishvili 1974). Another example of the effect of a change in the vegetation cover on rapid mass movements is the Mgeta valley, Tanzania. The original vegetation was woodland. More than 1000 landslides were triggered during one thunderstorm, of which 47 % originated in cultivated fields, and 46 % in grassland but less than 1 % in woodland covered areas. Almost all debris flow material was removed by channel flow (Temple & Rapp 1972).

Where man practices erosion control, the supply of material to the waterways is reduced. If the water discharge is not decreased in proportion to decreased load, channel erosion will compensate for the change in conditions.

Dam construction also brings about changes in the relation between water and sediment discharge. The channel immediately downstream from a reservoir adapts to clear water conditions through lateral or deep erosion, producing a flatter gradient which is in equilib-

rium with the smaller sediment transport (Joglekar 1960, Tinney 1962, Kinawy *et al.* 1973). Aggradation some distance upstream of reservoirs is a common phenomenon (Aust. Water Resour. Counc. 1970).

Such accumulations in channels are generally not included in reservoir sedimentation calculations. Reservoirs also modify the flow regime. The elimination of high peak flows reduces the clearance of accumulated sediment from the stream bed and affects the sediment yield.

Withdrawal of water for industrial and domestic uses may have an effect on groundwater level, with increased gully-ing along channels. Boreholes in semi-arid regions may lower the water table, thus reducing the vegetation potential.

7.2 Man's influence on erosion magnitude

Soil loss following the removal of vegetative cover, building of reservoirs and withdrawal of water is also affected by climate, because of climate-dependent differences in erosivity, flow regimes, and water supply to the ground water.

Cultivation. Different crops provide different degrees of vegetative cover. The relative erosion from crop plots compared to that from bare soil, *C*, is shown in Table 13, for West African conditions. In the USA, cotton has *C* values which range from 20 % to 80 %, depending on the type of rotation and crop stage period (Wischmeier & Smith 1965). Maize has *C* values which range from 6 % to 80 %. The relative erosion in American and West African conditions is comparable.

Erosion values from cultivated plots are given in Table 12. Normal values for

Table 13. Erosion from vegetation plots as a percentage of the erosion from bare plots under West African conditions—annual C factors in the Universal Soil Loss Equation (from Roose 1977)

	C, relative erosion, %
Bare soil	100
Dense forest or culture with a thick straw mulch	0.1
Savannah and grassland, ungrazed	1
Cover crops: late planted or with slow development: First year	30–80
Second year	10
Cover crops with rapid development	10
Maize, sorghum, millet	30–90
Rice (intensive culture, second cycle)	10–20
Cotton, tobacco (second cycle)	50
Groundnuts	40–80
Cassava (first year)	20–80
Palms, coffee, cocoa, with cover crops	10–30

cultivated land with a slope greater than 5 % range between 1000 and 5000 tons/km². There are exceptional values of around 25 000 tons/km² recorded in the USA as well as in India and Tanzania. Compared to cultivated land, grassland and forests seem to provide good protection against erosion. Only five values out of 30 have soil loss values greater than 100 tons/km². No climatic differences in soil loss values can be discerned from Table 12, as slopes and soils are not constant. The West African erosion values in Table 12 are not greater than the American soil loss values. The long period of mechanized agriculture in the USA might be responsible for the high soil loss figures. At Sefa, in southern Senegal, soil loss from traditionally cultivated ground-nut plots was 723 tons/km², while loss from mechanized cultivation was 1190 tons/km². Values for rice were 1844 tons/km² and 3211 tons/km², respectively. Agricultural practices are of great significance in erosion. Contour ploughing, strip cropping and terracing reduce erosion significantly (Stall 1962, Wischmeier & Smith 1965).

Grazing and burning are land use practices that leave the soil unprotected. In humid areas, where there is a contin-

uous sward of grass, overgrazing is no great problem. In semi-arid, marginal lands, where recovery of vegetation is slow, overgrazing may prove to be disastrous. Nomads may hold large numbers of cattle for prestige rather than for profit (UNESCO 1972). On soils lacking in humus, overcrowding on wet ground causes puddling and packing, increased runoff and sediment yield (Lusby 1965).

Trampling and grazing around corrals and watering places may result in localized destruction of vegetation (Thorntwaite *et al.* 1942). Since the 1950's, some 2300 boreholes have been drilled in the Kalahari. Livestock tends to concentrate in the vicinity of the permanent boreholes. Surrounding the boreholes is an area of heavily trampled, desiccated ground (Zumer-Linder 1975). In the Kisongo catchment, Tanzania, gully development starts where stock routes run downslope, as runoff is channeled down the bare strips (Murray-Rust 1972).

Four centuries of overgrazing in low, strongly seasonal rainfall is responsible for the high sediment yield of 2560 tons/km²/yr from a 2500 km² catchment in the mountainous area in the Bermejo headwaters, Argentina (UNESCO 1972). This high figure is comparable to severe cultivated plot erosion from crops. Sartz (1970) ranks heavily grazed pasture as equal in erosion to cultivated land in southwestern Wisconsin. In northern Mississippi, sediment yield is lower from pasture than from cultivated land (Ursic & Dendy 1963).

Erosion is also increased by burning before the onset of intense rains. Burning of grass, bushes and trees is a general practice in Colombia, South America and in Africa, south of the Sahara. According to an investigation using ERTS-satellite pictures, the coverage of burnt areas during the drought season of 1972–73 was 40 to 60 % of all land in a zone between 5 and 12° N lat. stretching from Nigeria to the Red Sea (SIES Rep.

1975, after Deshler 1974). The order of magnitude of the effect of fire on dry chaparral rangeland was measured in four catchments in Arizona. In the three years before the fire, the cumulative total rainfall was 1702 mm and sediment yield was 19 tons/km². In the 21-month period after the fire, 1303 mm of precipitation yielded 8308 tons/km² (UNESCO 1972). In two nearby basins with steep slopes and immature granitic soils, the sediment yields after the fire were even greater, 19,300 and 24,700 tons/km² (UNESCO 1972).

Lumbering. In a mountainous area of Idaho, USA, sediment yield in logged areas without road building was only 1.6 times that from undisturbed areas. Road construction in connection with tree felling increased the sediment production, and the sediment yield of the entire catchment was 155 times the undisturbed sediment yield (Megahan 1975). Roads to logging areas may also be subject to landslide hazards (Table 14). The effect of logging on landslide frequency generally increases in drier climates as a result of greater moisture differences (Rice 1977).

A consequence of clear-cutting in many humid areas is a raised water table. In a stream in Alaska, rising ground water increased the capacity of the flow to move bedload by a factor of 1000 (Anderson 1975, after Clayton *et al.* 1966). In effect, the groundwater inflow "lifted"

Table 14. A general survey of 725 slides in western Oregon after the 1964-65 floods (from Anderson 1975, after Dyrness 1967) and landslides/km² in an experimental forest on the same occasion (from Anderson 1975, after US For. Serv. 1965)

Land use	Slides, %	Land-slides/km ²
In undisturbed areas	22	0.1
Associated with logging areas	24	1.0
Associated with roads	54	31.5

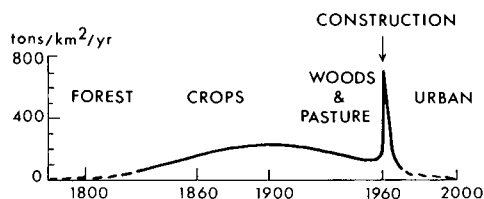


Fig. 11. Changes in land use and sediment yield in a fixed area near Baltimore, USA (from Wolman 1967).

the whole streambed, starting rapid sediment transport.

Mining. Spoil banks resulting from strip mining, particularly in coal fields, are often steep-sided and devoid of vegetation. A study of the effects of strip mining on the Cane Branch Watershed in Southern Kentucky revealed that an average annual sediment yield of about 10,425 tons/km² was recorded in that portion of the catchment with unreclaimed spoil banks. In contrast, the annual sediment yield from unmined catchments in the area was about 10 tons/km² (Sediment. Eng. 1975, p. 457, after Collier *et al.* 1964, and 1971).

Urbanization. Urban expansion implies construction of roads, pipes and buildings and paved ground. During the transition from rural to urban land, the sediment yield accelerates to a high peak during landscaping or construction, then declines as the land "heals", and finally levels off at some low stable rate representative of business or residential lands. Fig. 11 shows the order of magnitude of change in sediment yield with such a change in land use. In another study of eight small basins (0.91-25.2 km²) in Rock Creek and the Anacostia River, also in Maryland, the values were much higher than those shown in Fig. 11 (Proc. Sediment. Conf. 1976). Annual sediment yields attributable to cultivated and residential areas averaged 615 and 840 tons/km², respectively. Construction-site sediment yields ranged from 1614 to 22,640 tons/km² and averaged

7330 tons/km². The high post-construction yield was thought to be due to intense channel erosion as a consequence of increased runoff.

As has been described in this chapter, man's activities may accelerate erosion by as much as 10,000 times. Man's

activities are a threat to the equilibrium in nature, since more soil is eroded than is formed. In Parts II and III of this study, the recorded erosion values often reflect man's activities. This influence is not always mentioned explicitly. Nevertheless, it should always be kept in mind.

PART II THE CLIMATIC CONTROL ANALYSED IN SOIL LOSS MODELS

8 Introduction

It is apparent from Part I that erosion and transportation processes are mainly affected by the five major factors of climate, relief, soil, vegetation and man's activities. The interrelation between these factors was also pointed out. The manner in which the elements of the major factors are structurally related in a system may constitute the basis of a systems classification. In principle, one can distinguish between morphological, cascading and process-response systems.

Morphological systems consist exclusively of the network of structural relationships between the constituent parts of the systems. Cascading systems are defined by the path followed by the throughflow of energy or mass. Process-response systems represent the linkage of the former two systems so that the process-response system demonstrates the manner in which form is related to process (Chorley & Kennedy 1971, p. 3).

Anyone of the three systems can be chosen and applied when erosion and the factors which contribute to erosion are analysed. The models of erosion reflect the systems approach that has been employed in the analysis. A morphological systems approach leads to parametric models and a cascading systems approach leads to deterministic models. A parametric model is based on the identification of statistically significant relationships between variables which are

assumed to be important and for which there is a reasonable data base. A deterministic model is based on mathematical equations which describe the processes involved in the model, taking into account physical laws, e.g. the conservation of mass and energy (cf. Morgan 1979).

The main objectives of Part II are to investigate climatic differences in soil loss and in the importance of the five major factors and their elements in causing land erosion by water. In the following chapters, soil loss models by various authors are discussed and analysed under four headings:

1. Bivariate regressions on sediment yield (chapter 9).
2. Multiple regressions on sediment yield and delivery ratio (chapter 10).
3. Empirical models based on results from plot and slope studies (chapter 11).
4. Deterministic models (chapter 12).

Yevjevich (1972 p. 233) has presented the following definition of the regression concept: "Regression represents a mathematical equation expressing one random variable as being correlatively related to another random variable or to several random variables. All variables on the right side of this equation do not need to be random variables. The re-

gression equation may be any function that can be fitted to a set of points of observed variables”.

The first two groups of models include regressions from both small and large basins. The first group contains bivariate regressions of sediment yield on one climatic or hydrologic variable. The second group of regressions contains several variables. The last two groups constitute models of sheet and rill erosion that are based on data from plots or slopes but which have also been modified for use in small drainage basins.

Before presenting the models, it is necessary to define some frequently used terms, viz. gross erosion, sediment yield, sediment concentration and delivery ratio. *Gross erosion* is the total erosion (total detached and entrained material)

which includes sheet, rill, gully and channel erosion. Gross erosion may be expressed by different units, e.g., weight per unit area and time or mm of ground lowering per unit time. *Sediment yield* is the sediment flow past a given cross section in a stream channel or a river. Sediment yield is usually expressed in weight per unit of time and includes suspended load as well as bed load. However, because of difficulties in measuring the bed load, sediment yield often includes only suspended material. *Sediment concentration* is the ratio of the weight of dry solids to the volume of the water-sediment sample. It is usually expressed in parts per million (PPM) or mg/l. *Sediment delivery ratio* is the ratio of sediment yield to gross erosion and is often expressed as a percentage.

9 Bivariate regressions on sediment yield

9.1 Introduction

In the analysis of climatic differences in soil loss, simple bivariate regressions are first examined. A regression model, linear or curvilinear, may be approximated by fitting a line to the scatter of points to give the best-fit relationship between the independent and dependent variable. One important assumption underlying the notion of the best-fit regression and the calculation of the regression equation is that the values of the dependent variable are normally distributed in their scatter about the regression line (cf. Chorley & Kennedy 1971, p 27).

It should be stressed that regressions are only representative of those areas and those environmental conditions from

which the data are drawn. Furthermore, the validity of the regression is controlled by the density of data in the region investigated. One main task in the present study has therefore been to examine the data base used in the regressions to see from where the data are derived. The aim has been to find out if there is any bias in the selection of catchments or if any environmental factors such as soil, land use or any neglected climatic elements may dominate the data base and thus the resulting regression between the variables. Some general conclusions have then been drawn, based on the assumption that sediment yield in a particular region with a given climate varies mainly in accordance with the regression lines.

9.2 Precipitation as an independent variable

Several workers have correlated sediment yield and mean annual precipitation. The most well-known correlations are those by Langbein & Schumm (1958), which are illustrated in Fig. 12. One curve was obtained from sediment station data from 94 stations, which were situated mostly in the central USA (with no stations in the Pacific Coast Region, New England or in the Gulf Region). The other curve was based on data from 163 reservoirs in the USA. The position of the basins was only indicated in a few instances.

Langbein & Schumm (1958) have derived equations representing the two regression curves which are based on theoretical considerations. The equations of sediment station data and reservoir data are converted to metric units in equations 19 and 20, respectively.

$$S = \frac{20.57 \cdot 10^{-4} P^{2.3}}{1 + 1.47 \cdot 10^{-8} P^{3.33}} \quad \dots\dots\dots (19)$$

$$S = \frac{41.14 \cdot 10^{-4} P^{2.3}}{1 + 1.47 \cdot 10^{-8} P^{3.33}} \quad \dots\dots\dots (20)$$

where

S = sediment yield in tons/km²/yr

P = effective precipitation in mm

The first factor $20.57 \cdot 10^{-4} P^{2.3}$ or $41.14 \cdot 10^{-4} P^{2.3}$ was thought to describe the erosive action of rainfall in the absence of vegetation, while $1/(1 + 1.47 \cdot 10^{-8} P^{3.33})$ was thought to represent the protective action of vegetation (Langbein & Schumm 1958).

The curves for sediment stations and reservoirs do not coincide. Langbein & Schumm (1958) attributed the discrepancy between the curves to three factors. One was the difference in the size of the drainage basins which were used in the analysis. The reservoir catchments averaged about 80 km² in area and the sediment station basins averaged about 3900 km². The second was that reservoirs are

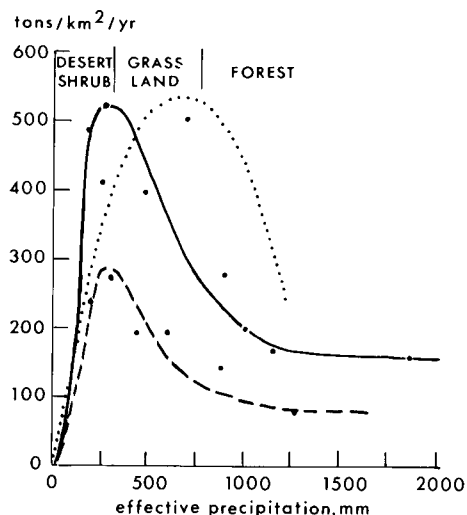


Fig. 12. Variation of sediment yield with precipitation determined from reservoir surveys and sediment station records (dashed line) (from Langbein & Schumm 1958) and as a comparison a similar curve based on both reservoir and sediment station data (dotted line) (from Rango 1970).

generally constructed in steep terrain, whereas hydrological stations are situated in various topographical environments. The third was that bedload is trapped in reservoirs, but is normally not measured at sediment stations. The last factor may, however, be balanced by low trap efficiencies in small reservoirs.

There are some weaknesses in the data base used in the construction of the Langbein & Schumm curves which make them inappropriate for use on a regional or on a world-wide basis. Several values are based on data collected over a period of a few months or less (Wilson 1973). This period is too short, as the annual sediment yield of a river can vary by as much as a factor of five or more from one year to the next (Douglas 1967). Thus, the representativity of much of the data is uncertain. The exact location of the peaks on the curves may be explained by factors other than precipitation. Scrutiny of the original data reveals that some points represent tributaries on a few riv-

ers. The first point on the reservoir curve (at 216 mm) is the mean sediment yield from tributary reservoirs on two rivers, the San Rafael Swell in Utah and the Badger Wash in Colorado. The second point is the mean of 26 reservoirs in Twenty-mile Creek basin, Wyo.; 7, in Cornfield Wash, N. Mex., and 5 others. The Twenty-mile Creek values influence the mean value considerably. There are only five sets of data making up a single mean value on the reservoir curve from 1400 to 2540 mm. The curve is therefore uncertain within that precipitation range.

Effective precipitation is estimated from mean annual runoff. A graph relating precipitation to runoff was used to convert known values of annual runoff to the effective precipitation at a temperature of 10°C. This means that their curves are only representative of basins within a limited range of climates, viz. basins in the USA which have a semi-arid or humid continental climate, with a mean annual temperature of 10°C. However, Schumm (1965) has also constructed curves which are displaced laterally and which represent different mean temperatures (Fig. 13). The curves are based on the relationships established in the USA between precipitation and runoff for different temperatures. No new measurements are included. The curves in Fig. 13 must therefore be regarded as illustrations of principles rather than quantitative relationships. This applies also to Fig. 12 and should be kept in mind, as the curves in Figs 12 and 13 are widely referred to both in textbooks and scientific articles.

Kirkby (1975) simulated the effect of precipitation and temperature in a runoff and sediment model, employing a 27° slope with a length of 100 m. He considered a simulated rainfall intensity of 15 mm/rainday and applied the model for annual rainfalls of up to 3000 mm (Fig. 14). Different curves were produced for different temperatures (potential evapo-

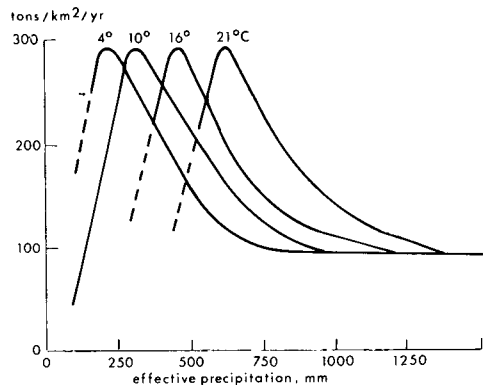


Fig. 13. The effect of temperature on the relation between sediment yield and precipitation according to Schumm (1965).

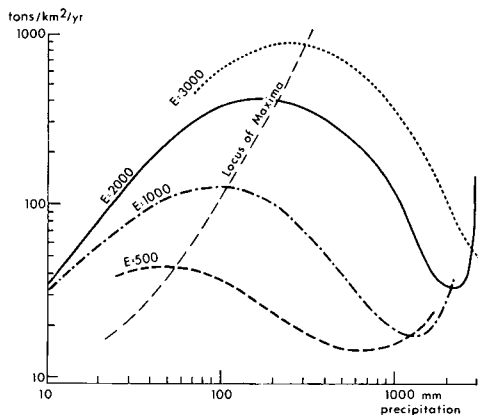


Fig. 14. The effect of temperature on simulated ground lowering. E = potential evaporation. Soil creep was estimated by $S_{\text{creep}} = 10 \tan \beta$ ($\text{cm}^3/\text{cm}/\text{yr}$) and soil wash by $S_w = 170 Q^2 \tan \beta$ ($\text{cm}^3/\text{cm}/\text{yr}$) where β = slope inclination in degrees, and Q = annual overland flow in m^2/yr (from Kirkby 1975).

ration). The temperature dependence of sediment yield is more pronounced in Kirkby's than in Schumm's curves. There is also a second rise in Kirkby's curves, produced by high rainfall values, which does not appear in the Langbein-Schumm curves. This can be compared with other curves discussed below.

Langbein & Schumm's curves (Fig 12) can be compared to another curve regressed with data from 673 catchments in the USA less than 26 km² (Rango 1970). All data were adjusted to a drainage area of 26 km². Most of the reservoirs and sediment stations were located in the Midwest, but other areas of the USA were well represented (Rango 1970, p. 1079). This curve (Fig. 12) has a peak sediment yield of 567 tons/km² at a mean annual precipitation of 700 mm. A comparison with Langbein & Schumm's curves shows that qualitatively the shape of the curves is similar but nevertheless the position of the peaks differs. One relationship which is still more interesting, since it includes data from outside the USA, is illustrated in Fig. 15. This curve, derived by Wilson (1973b), is based on data from 1500 drainage basins, 1250 of which represent continental climates; 200, Mediterranean; and 50, tropical wet-dry climates. Short term records and those markedly affected by upstream regulation were not included, and only data based on three or more years of records were accepted. Records from both large and small basins were used, but all values were adjusted to a basin size of 259 km², in accordance with a relationship established by Brune (1948). The data were grouped according to the amount of precipitation, using 76 mm (3 inch) class intervals. Moving averages were then calculated.

Adjustment to a single basin size is highly desirable, but involves a considerable uncertainty, as there is no well-defined relationship which is generally applicable to all climates. Different climates are better represented in Wilson's basic data than in Schumm's data. Wilson's data include a wider range in annual temperatures. This is probably the main reason for the wide scatter encountered in the raw data (Wilson 1977, p. 85).

Wilson's curve exhibits one peak at

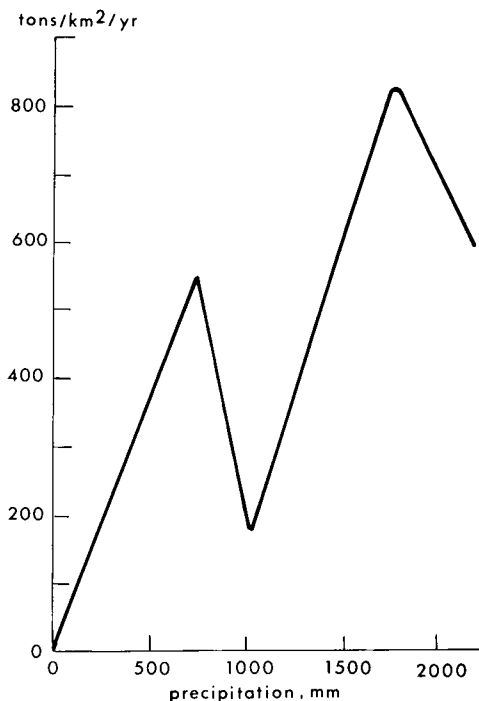


Fig. 15. Relationship between mean annual sediment yield and mean annual precipitation (from Wilson 1973).

about 750 mm annual precipitation and a second peak at about 1750 mm. Wilson found that much of the basic data contributing to the first peak was obtained from small basins in loessial material in regions subject to intense agricultural usage. When these basins were eliminated, the resulting curve exhibited a peak in the sediment yield curve at 380 mm precipitation. Land use and soil conditions may thus significantly affect the position of the peak. The second peak represents tropical wet-dry and Mediterranean climates. For the 200 Mediterranean basins, peak sediment yield occurred within the 1270–1650 mm range. The 50 sets of data from tropical wet-dry basins indicate peak sediment yield in the 1780–1900 mm mean annual precipitation range. For large basins, $\geq 25,000$ km², the peak at 1750 mm precipitation was absent. Consequently, the size of

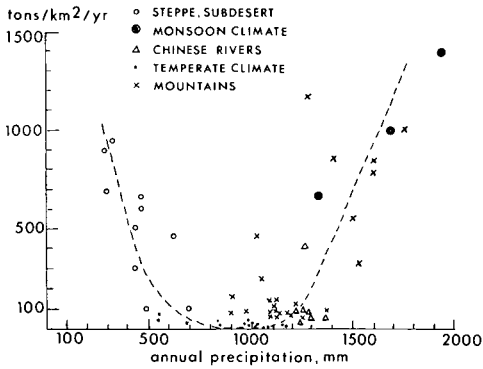


Fig. 16. Mean annual sediment yield related to mean annual precipitation (from Fournier 1949).

drainage basins may also affect the height of the peak.

Fournier (1949) also obtained a second rise in sediment yield with high precipitation values (Fig. 16). Several mountain basins and four monsoon climate basins, viz. the Yangtze-kiang, the Sikiang, the Red river and the Ganges basins, were responsible for this rise in the curve. These river basins are intensely cultivated. The first limb of the curve represents rivers such as the Indus, Rio Grande, Huangho, Tigris and Colorado.

The aforementioned graphs cannot be used to estimate or predict sediment yield in a particular river basin. They only give the general trend of the effects of precipitation. Mean annual precipitation data mask the importance of temperature, seasonality and rainfall intensity. The discrepancy between the curves (Figs 12, 15, 16) is mainly a result of the different climatic regimes from which the raw data are drawn, and how much the natural environment is affected by human activity. However, all the studies indicate that semi-arid regions have a large sediment production. If precipitation data from low temperature areas of northern Canada and the northern Soviet Union had been included, there would certainly be a much-reduced low precipitation peak (cf. Fig. 14). Fournier's data

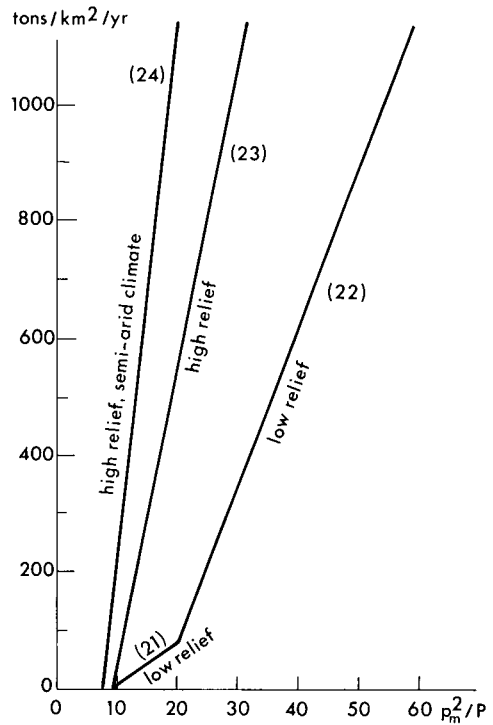


Fig. 17. Relationships between the parameter p_m^2/P and sediment yield. Equation numbers refer to the text (from Fournier 1960).

points (Fig. 16) indicate a low sediment production in humid temperate regions. None of the curves include data from Af climates. When data from wet-dry Mediterranean, tropical wet-dry (Wilson) and monsoonal climates (Fournier) are included, a second rise in sediment yield is observed.

In order to make allowance for seasonality, Fournier (1960) introduced the parameter p_m^2/P , where p_m is the mean precipitation in the month with the greatest rainfall, and P is the mean annual rainfall. The suspended sediment yield values from drainage basins greater than 2000 km² could be represented by four lines, two associated with high relief and two associated with low relief (Fig. 17).

Fournier found the following relationships:

$$Y = -49.78 + 6.14 p_m^2/P \dots\dots (21)$$

$$Y = -475.4 + 27.12 p_m^2/P \dots\dots (22)$$

$$Y = -513.21 + 52.49 p_m^2/P \dots\dots (23)$$

$$Y = -737.62 + 91.78 p_m^2/P \dots\dots (24)$$

where Y = suspended sediment yield in tons/km²/yr

(21) refers to regions with low relief and a precipitation regime such that

$$8.1 < p_m^2/P < 20. \quad r = 0.80$$

(22) refers to regions with low relief and a precipitation regime such that $p_m^2/P > 20. \quad r = 0.96$

(23) refers to regions with pronounced relief in all climates except semi-arid climates. $r = 0.95$

(24) refers to regions with pronounced relief in semi-arid climates. $r = 0.98$

Fournier used basic data from 96 drainage basins greater than 2000 km². There was no river from a tropical rainy climate with no dry season. Only one river basin, the Irrawady, was classed as having an exclusively Aw climate. Two others, the Ganges and the Sikiang, were classed as having a climate somewhere between Ca and Aw. These river basins are subject to intense agriculture.

Equation 23 was tested by Robert (1973, referred to in Arnoldus 1974) against measured values of suspended sediment in Morocco (Table 15). The

measured values in Table 15 are as a rule much lower than the erosion rates calculated by Fournier's equation.

Corrections for land use differences or any factors other than temperature and area have not been made in the data for the regressions under discussion. This means that the regressions may be quite unreliable and inaccurate when applied to a particular river basin because land-use is such a major factor.

9.3 Runoff as an independent variable

In several studies, sediment yield has been correlated with runoff and has been represented in the form of regression curves or data points in diagrams. In order to compare the results, a composite diagram drawn from different sources has been constructed and reproduced as Fig. 18.

Regressions of runoff on sediment yield from Asia and East Australia (Douglas 1967) are compared to regressions and data from the United States (Langbein & Schumm 1958, cf. Douglas 1967, Judson & Ritter 1964, cf. Stoddart

Table 15. Comparison between values computed by Fournier's equation 23 and measured sediment yield values in Morocco (from Arnoldus 1974, after Robert 1973)

Catchment and sub-catchment	Stations	Area km ²	p_m^2/P	Sediment yield tons/km ² /yr	
				Equ. (23)	Measured
<i>Sebou</i>					
Sebou	Pont du Mdez	3 474	39	1 533	320
Sebou	Ain Timedrine	4 429	39	1 533	590
Sebou	Azzaba	4 736	39	1 533	530
Querrha	M'Jara	6 183	66	2 951	2 500
Inaouène	Touaba	3 324	48	2 006	1 100
Beth	El Kansera	4 542	43	1 743	450
<i>Oum er Rbia</i>					
Ht Oumher Rbia	Decza el Qued	3 320	60	2 636	570
El Abid	Bin el Quidane	6 400	43	1 743	365
Lakhdar	Sidi Driss	2 850	43	1 743	370
<i>Bou Regreg</i>					
Bou Regreg	Barrage	9 620	47	1 953	165
Grou	Barrage	5 540	47	1 953	300

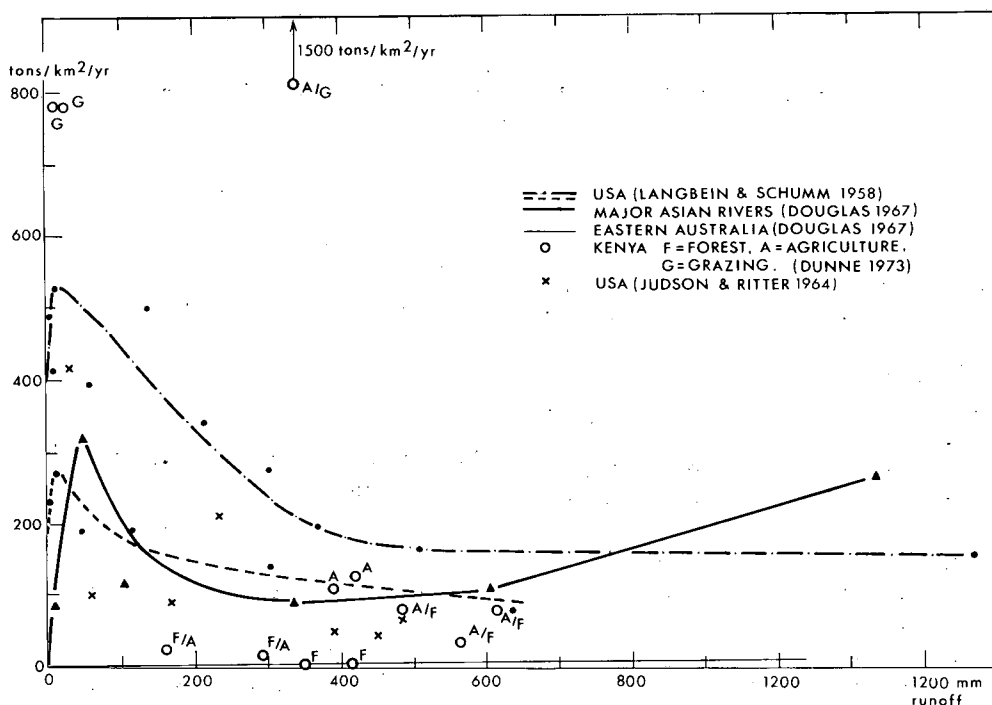


Fig. 18. Relationships between sediment yield and runoff from different investigations. The US curves are converted from effective precipitation back to runoff by using the graphs included in the appropriate article. The upper US curve represents reservoirs. The US values by Judson & Ritter represent large single basins or groups of basins. The curve by Douglas is based on data from various sources including some major Asian rivers. The east Australian curve is based on data from small catchments with natural vegetation.

1969) and Kenya (Dunne 1973). Douglas' curve for a "selection of rivers including some major Asian rivers" is converted from m^3/km^2 per year to tons/km^2 per year by assuming a density of $2.65 \text{ tons}/\text{m}^3$ (cf. Douglas 1973). Langbein & Schumm's sediment station curve and Douglas' curve exhibit considerable agreement in spite of the fact that Douglas includes some major rivers. There is a second rise in Douglas' curve for high runoff values, which is in agreement with the precipitation regressions of Figs 15 and 16.

The curves from the United States and Asia in Fig. 18 and the plotted values from Judson & Ritter (1964, cf. Stoddart 1969) for very large basins and from Dunne (1973) for grazing and agriculture

basins in Kenya, show sediment yields from areas where human activity is considerable. More natural vegetation is represented by the east Australian curve (Douglas 1967) and the forest catchments in Kenya (Dunne 1973).

Mean annual runoff masks different runoff regimes and differences in sediment yield within an average year. Wilson (1977) tried to overcome the problem by using monthly values (Fig. 19). The diagram cannot be used to compare many rivers, but it can be used to assess seasonal variations in concentration and sediment yield in a few rivers in different climates. It clearly represents the varying water-sediment relations during an average year.

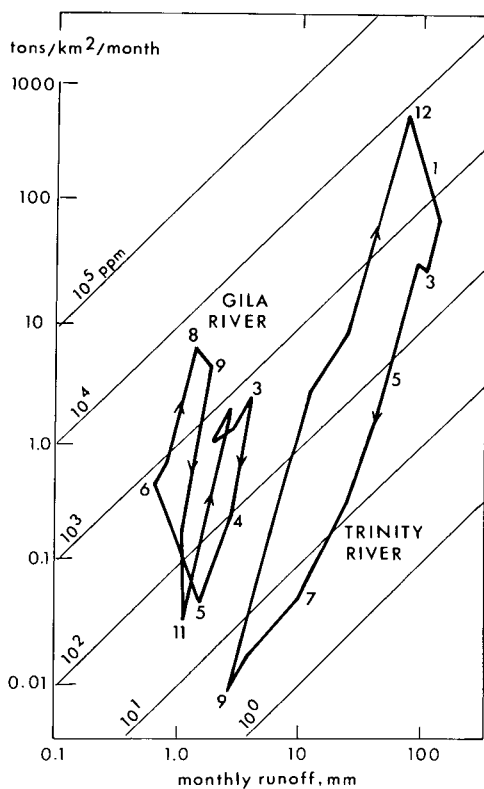


Fig. 19. Mean monthly suspended sediment yield related to monthly runoff for the Trinity river, located in a Cs climate, and the Gila river, located in a BS climate (from Wilson 1973).

9.4 Water discharge as an independent variable

Variations in sediment yield or sediment concentration with water discharge are treated in many papers (Piest 1963, Nilsson, 1971, Negev 1972, Temple & Sundborg 1972, Flaxman 1975, Wood 1977, Jacobi 1977, Kung & Chiang 1977), in general surveys (Sediment. Eng. 1975, Walling 1977) and in a theoretical discussion by Allen (1974).

A sediment rating curve is a curve representing the relationship between water discharge and suspended sediment load in tons/s or tons/day. There are also suspended sediment concentration ratings (Walling 1977). Sediment rating curves

are often constructed to calculate sediment yield. Many sediment rating curves are assumed to be linear when plotted on logarithmic paper, even though the general shape of the curve often is concave downward (Sediment. Eng. 1975). Assuming a linear log relationship, the equations of the suspended sediment concentration curve and the sediment yield rating curve respectively, are (Nilsson 1971):

$$\log S_c = \log a + m \log Q \quad \text{or} \\ S_c = aQ^m \dots \dots \dots (25 \text{ a})$$

$$\log S = \log a + (m+1) \log Q \quad \text{or} \\ S = S_c \cdot Q = aQ^{m+1} \dots \dots \dots (25 \text{ b})$$

where S_c = sediment concentration in mg/l, kg/m³, or PPM, S = sediment load in tons/s, Q = water discharge in m³/s

The log-transformation is often undertaken since a linear regression is easier to compute than a non-linear regression. However, the regression technique implies that greater weight is given to smaller log-transformed values than to larger log-transformed values (Yevjevich 1972). This is a particularly weak point in the construction of a sediment rating curve since more samples of low concentration which occurs at low discharge are often taken than of high concentration, which occurs at high discharge. This adds to the uncertainty of the curve at higher discharges.

The coefficient a in equations 25 a and b indicates the general availability of sediment, while m shows how sediment concentration varies with water discharge. If $m < 1$ the concentration increases at a rate which is less than the proportionate increase in Q . If $m > 1$, concentration increases at a rate that is greater than the proportionate rate of increase in discharge. Consequently, when $m > 1$ much of the annual sediment yield might be contributed by the peak discharges during a few days.

Other methods of constructing sediment rating curves have been proposed

by Piest (1963) and Jacobi (1977). There is often a very wide scatter in plots of sediment concentration or load against water discharge. The variation of sediment concentration at low discharges can be one hundred fold, as a result of, e.g., seasonal variations in erodibility, and differences in concentration between rising and falling water stages. Sediment load is controlled by the supply of sediment to the rivers rather than by the ability of the rivers to transport the available material. Therefore, the use of stage-adjusted rating curves and seasonal relationships considerably reduces the errors in estimating sediment loads.

Fleming (1969) plotted annual suspended sediment load against mean annual water discharge on log-log paper. The resulting graph showed a considerable scatter. When the 253 data points were classified and divided into four main vegetational zones, correlations were improved (Fig. 20). As can be seen from Fig. 20, one value of mean annual

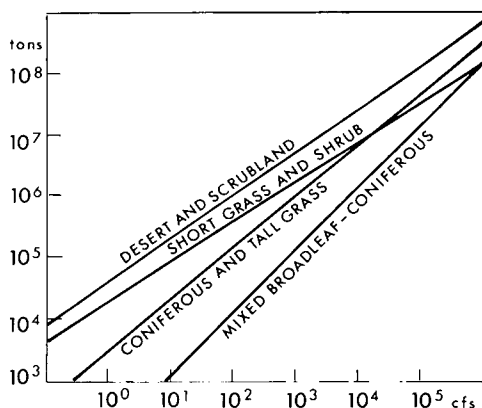


Fig. 20. Regression lines of sediment yield against mean annual discharge for different vegetation zones (modified from Fleming 1969).

discharge is associated with different sediment yields in each vegetational zone. However, the differences between the zones seem to decrease with increasing basin sizes as the lines converge towards higher discharges. This fact is relevant to the discussion in the next chapter.

10 Multiple regressions on sediment yield and delivery ratio

10.1 Sediment yield equations

Sediment yield is influenced by several factors as is apparent from the discussion in Part I. Therefore, the yield can be expected to be predicted more adequately by multiple regression equations than by bivariate regression equations. Multiple regressions may be linear or non-linear. The true form of the functional relationship between sediment yield and the 'independent' variables is not known and can never be expected to be fully clarified. Simple graphical correlations between dependent and inde-

pendent variables are often used to test what type or relationships are the most appropriate. Because linear equations are easier to treat than non-linear multiple equations, non-linear functions are often transformed to linear relations (Yevjevich 1972). Linearization by logarithmic transformation is often undertaken. However, a logarithmic transformation of variables introduces a bias in itself, as the use of least squares of transformed variables in multiple regression techniques gives greater weight to the smaller values than to the larger values (Yevjevich 1972).

In multivariate regression analysis, those variables among the test variables which account for most of the variance are the ones selected for the final regression equation. The regression technique implies that the sum of squared residuals of the dependent variable is minimized. However, intercorrelation between the independent variables often creates problems. It is possible to accidentally omit from the analysis that variable which is genuinely 'causing' the observed variation in the dependent variable, and to obtain a satisfactory explanation of the variation in terms of other 'independent' variables which are similarly under the control of the excluded variable. On the other hand, if one variable is found to be non-significant, the study areas may be similar with respect to that variable. Another possibility is that the tested variable may not be a good indicator of differences in the factor that the variable was intended to describe.

Variables covering different characteristics of all the five major factors of climate, relief, soil, vegetation and land use should be included as test variables in order to make the resulting regression the best possible equation. Principal component analysis is used by several authors to select variables (Flaxman 1966, Imeson 1971, Douglas 1973). The usefulness of principal component analysis is illuminated in Rice (1970), and Blake *et al.* (1970). Element analysis is explained in Newton & Spurrell (1967) and Douglas (1969) and has been used in one study to identify those variables which are most important in controlling sediment yield (Douglas 1973). It is not within the scope of this study to discuss these types of analysis.

The following pages contain a discussion and an examination of regressions from various sources. This examination has been undertaken in spite of several difficulties in obtaining comparable data from the different investigations. Some

authors give complete information on which variables have been tested; others do not give any information at all. Only a few variables may have been tested, or the variables may have been chosen by chance or on the basis of the availability of data. Different techniques of sediment yield measurements have been used. Nevertheless, the conclusions that are drawn from the collected regression equations have been based on the assumption that the regressions in every case contain those variables of all possible variables which best explain the variation in sediment yield within the region covered by the equation. The regressions are arranged according to climatic types. Each region for which a regression equation is developed is then looked upon as a sample of the population consisting of all regions of the same climatic type. Conclusions have then been drawn about the significance of different variables in different types of climate.

Because of the intercorrelation between the independent variables, it is difficult to find a statistic that can correctly assess the importance of the individual variables in causing variation in sediment yield. Most articles give the coefficient of multiple determination R^2 which is the percentage of the variation in the dependent variable that is accounted for by the entire regression equation. Only a few authors give the change of R^2 at each step by inclusion of a new variable (Flaxman & Hobba 1955, Lisitsyna & Alexandrova 1972). In order to get an idea of the importance of the individual variables, it has been necessary to re-calculate the equations. This has been possible in those cases in which the articles contain raw data (cf. Table 18). Because there is no statistic that unambiguously indicates the importance of individual variables, several parameters are presented in Table 18.

Very few equations have been recalculated.

lated for each climatic type. Therefore, the conclusions drawn from the statistics concerning climatic differences are necessarily based on a very limited amount of data and are therefore extremely uncertain.

The multiple regressions which have been studied are tabulated in full and discussed climate by climate in the Appendix. Table 16 gives the equations in a simplified form. The first column contains the type of climate using the symbols explained in Table 1. The number of basins and the size range of the catchments used in the regression analyses are presented in columns 2 and 3. The equation numbers in Table 16 are the same as those in the complete versions given in the Appendix. The last column indicates how much of the variation in sediment yield that is accounted for by each equation.

The main objective of this examination of multiple regression equations is to attempt to answer the following four questions about the influence of different variables on sediment yield:

1. Does the importance of the variables vary with catchment size?
2. Are some variables of unequal importance in different climates?
3. What is the relative importance of the five major groups of factors: climate, relief, soil, vegetation and land use in different climates?
4. Which climatic variables should be used in an equation intended to be valid for many climates?

The first question: Table 17 has been constructed as a basis for the discussion of the first question. The variables have been grouped into the major factors of climate, relief, soil, vegetation and land use. In some cases, additional variables are used, listed under the heading 'Others'. The equations are divided into two groups representing small and large basins, respectively. More detailed information about the variables is given in

the Appendix. Some general conclusions about the importance of variables in relation to basin size will be drawn in the following paragraphs.

In equations for small basins, gross erosion (E) is often used as a variable together with variables such as relief and area. Gross erosion is a compound variable but is closely related to climatic conditions. Therefore, it has been listed under this heading. However, in the assessment of gross erosion, equations are often used that take into account all the five main factors (cf. chapter 11). Equations for small catchments include precipitation variables (included in E), while equations for large drainage basins generally include discharge and temperature variables. In two studies of small basins, when discharge or peak discharge is used, alternative equations with precipitation variables give better or equally good results (41-42, 55-56). In equations of sediment concentration, discharge is the most suitable variable for all basin sizes.

Relief parameters are frequently used for all basin sizes. Equations for small basins often include relief/length as a parameter while those for large basins often include gradient of the main river. Area is often included in equations for small basins and is also common for large basins.

In equations for small basins, soil variables are common but for large basins they appear in only one case. A very rough estimate of geological conditions is included in two cases. This lack of soil variables in equations for large basins may be due to the difficulty in generalizing over large areas, the lack of an adequate soil variable, or a lack of sufficiently detailed soil studies. Anderson (1954) has shown one way of overcoming the problem (cf. equation 78 in the Appendix). Arnoldus (1977b) has also described a way of estimating the distribution of a soil variable. (cf. section 16.3).

Table 16. Simplified functional equations. Complete equations are found in the Appendix

List of symbols used in this Table and in Tables 17–19

S = sediment yield	q = peak discharge	B = bifurcation ratio
S ₂ = sediment load/storm	M = month	K = soil
SA = sediment yield/area	T = temperature	G = geology
SAY = sediment yield/area/year	E = gross erosion	C = land use or vegetation
SAD = sediment load/area/day	R = relief or gradient	EB = eroding banks
SAM = sediment load/area/month	H = altitude	QS = stage of discharge
Sc = sediment concentration	L = channel length	TR = trap efficiency
P = precipitation	A = area	AGE = age in years
I = intensity of precipitation	F = shape factor	
Q = discharge or runoff	D = drainage density	

A. Small drainage basins

Climate	No of basins	Approximate size range km ²	Functional equation	R ²
Da, Ca (grassland)	30	0.1–110	S = f(E,A,AGE, TR) (26)	93 %
	36	0.1–7250	SAY = f(E,A,P) (27)	88 %
	7	1–20	SAY = f(E,A) (28)	93 %
	20	0.2–670	SA = f(E,R,AGE,D,TR,A) (29)	88 %
	12	20–2350	SAY = f(E,TR,TR,R,F) (30)	97 %
		"	SAY = f(E,TR,TR,R) (31)	95 %
	22	0.2–10	SAY/E = f(R,TR,A,TR) (32)	76 %
		"	SAY/E = f(R,R,TR,A,TR) (33)	79 %
BS (Middle latitude steppe)	14	0.01–60	SAY = f(P,Q,C,R,A) (34)	
	17	0.05–1	SAY = f(R,C) (35)	86 %
	84	<10	SAY = f(TR,E,A) (36)	71 %
	22	<10	SAY = f(TR,R,A) (37)	77 %
	13	<10	SAY = f(TR,R,A) (38)	85 %
Cs	38	0.04–20	SAY = f(TR,AGE,A,P,C) (39)	80 %
	8	9–1100	SAY = f(P,I,K) (40)	86 %
	8	12–520	SAY = f(q,D,C) (41)	90 %
Ca,Cb,Cc,Da (forest land)		"	SAY = f(A,I,P,D,C) (42)	
	1	7	S ₂ = f(AGE,P,P,T,q) (44)	73 %
	1	20	Sc = f(Q,q) (45)	79 %
	1	15	Sc = f(Q,P) (46)	92 %
	20	3–350	Sc = f(Q,EB,K,K,K,C,QS) (47)	52 %
	"	"	Sc = f(Q,EB,QS,K) (48)	49 %
	"	"	SAD = f(Q,EB,K,K,K,QS) (49)	76 %
	"	"	SAD = f(Q,EB,QS,K) (50)	76 %
	"	"	SAD = f(Q,EB,QS,K) (51)	76 %
Different climates (Western USA) (Western USA) (N.E. Queensland) (N.E. Queensland) (N.E. Queens- land + N S Wales)	13	1–230	SAY = f(P,B,R,D,R,C,G) (52)	
	30	0.05–620	SAY = f(I,R,K,K) (53)	81 %
	28	0.01–>130	SAY = f(P/T,R,K,K) (54)	92 %
	11	6–590	SAY = f(Q,R,D) (55)	83 %
	"	"	SAY = f(P,B,D) (56)	82 %
	24	1–590	SAY = f(P,B) (57)	55 %

Equations for small basins often include land use variables, which describe density of vegetation cover, the percentage of bare soil, or the percentage of

cultivated land. In other instances the land use variable is included in the gross erosion calculation. Only two equations (78, 79) for large basins include indices

B. Large drainage basins

Climate	No of basins	Size range km ²	Equation	R ²
A	14	>5000	SAY = f(Q,A,R,T) (58)	94 %
B	14	>5000	SAY = f(Q,H,P,T,C) (59)	86 %
"	16	500–25000	SAY = f(Q) (74)	42 %
"	14	1000–31600	SAY = f(Q,H) (76)	66 %
D+B	14	1500–35000	SAY = f(Q,R) (72)	83 %
D+B	26	500–35000	SAY = f(Q) (73)	47 %
C	25	>5000	SAY = f(Q,R,T,C) (60)	63 %
C, R≤3		>5000	SAY = f(Q,A,H,R,T) (61)	76 %
C, R>3		>5000	SAY = f(A,H,R,P,C) (62)	95 %
C	29	150–18850	SAY = f(A,q,Q,R,K,K,C,C,C,C) ... (78)	
"	17		SAY = f(C,EB) (79)	
"	1	350	SAM = f(M,P,P,M) (87)	80 %
"	1	800	Sc = f(Q,M,T,Q,q,P) (81)	
"	1	750	Sc = f(Q,M,T,Q,P,I) (82)	
"	1	1600	Sc = f(Q,M,T,I) (83)	
"	1	1200	Sc = f(Q,T,q,P) (84)	
"	1	5400	Sc = f(Q,M,T,Q,q) (85)	
"	1	11840	Sc = f(Q,M,T,Q,q) (86)	
"	5	1860–57000	SAY = f(TR,Q) (89)	
"	"	"	SAY = f(TR,Q,A) (90)	
"	"	"	SAY = f(TR,Q,R) (91)	
D	1	250	Sc = f(Q,Q,q,P) (80)	
"	26	>5000	SAY = f(H,P,C,G) (63)	64 %
"	8	100–50000	SAY = f(Q) (65)	74 %
"	76	100–50000	SAY = f(Q,C) (66)	53 %
"	21	500–20000	SAY = f(Q,R,D) (67)	67 %
"	11	100–10000	SAY = f(Q) (68)	52 %
"	14	1500–35000	SAY = f(Q,R) (69)	59 %
"	7	<1500	SAY = f(Q,R) (70)	88 %
"	11	500–25000	SAY = f(Q,R) (71)	94 %
"	36	75–1400	SAY = f(R,A,L) (88)	25 %
H(B?)	26	140–4650	SAY = f(Q,H,A) (75)	77 %
H(B?)	9	2600–13500	SAY = f(Q,A) (77)	90 %
All climates A,B,C,D	79	5430–6130500	SAY = f(Q,A,H,P,T,C,G) (64)	58 %

of man-made changes in natural vegetation. Five equations (59–60, 62–64) include a very generalized natural vegetation variable.

Re-calculations of original data: In order to attempt to answer the three remaining questions, the equations from articles containing raw data were re-calculated in a forward stepwise regression program (Nie *et al.* 1975). With the exception of equation 78, the regression coefficients of the new equations do not differ much from those in the original articles. However, in equation 29 a

negative constant of –4.2 was obtained instead of +3.9 in the article. In the case of equations 52 and 57, higher R² values were obtained than in the article (Table 19). One reason for this may be that both original values and log values were listed in the article and in some cases these values were not comparable. The non-logged data were used in the re-calculation. The R² values of the re-calculated equations 53 and 88 also differ from those in the original articles (cf. Table 19).

Table 18 contains statistics from the

Table 17. Types of variables used in multiple regressions on sediment yield. Regressions for single basins are not included. A list of symbols precedes Table 16.

Climates covered by the equation	Equation number	Variables grouped into major types of factors					
		Climatic, Hydrologic	Relief, Area	Soil	Vegetation, Land use	Others	
<i>Small basins</i>							
Da, Ca (grassland)	26	E		A		AGE,TR	
	27	E, P		A			
	28	E		A			
	29	E		R, D, A		AGE,TR	
	30	E		R, F		TR,TR	
	31	E		R		TR,TR	
	32	E		R, A		TR,TR	
	33	E		R, R, A		TR,TR	
34	P,	Q	R,	A	C		
BS (middle latitude)	35			R		C	
	36	E			A		TR
	37			R,	A		TR
	38			R,	A		TR
	39	P			A	C	AGE,TR
Cs	40	P, I			K		
	41		Q	D		C	
	42	P, I		D, A		C	
Ca, Cb, Cc, Da (forest land)	49		Q		K,K,K		EB,QS
	50		Q		K		EB,QS
	51		Q		K		EB,QS
	52	P		R, R, B,D	G	C	
Different climates							
BS,BW,Cs (W. USA)	53	I		R		K,K	
BS,BW,Cs (W. USA)	54	P/T		R		K,K	
Am,(Aw),Ca(N.E. Queensland)	55		Q	R,	D		
Am,(Aw),Ca(N.E. Queensland)	56	P			B, D		
Am(Aw),Ca,Cb,Cc (N.E. Queensland + N S Wales)	57	P		B			

re-calculated equations. The simple correlation coefficient is interesting in itself and also in comparison with the statistic R^2 change. The standardized partial regression coefficient, beta or beta weight, includes intercorrelation effects. The greater the degree of intercorrelation between the 'independent' variables the less is the reliability of the beta weight in indicating the importance of the variables.

The next statistic shown in Table 18 is the change in R^2 on the addition of a new variable. Since this value is affected by the order of inclusion of each variable, this order is indicated in the table and the squared part correlation coefficient is

included for comparison. Part correlation (also called semi-partial correlation) is a simple correlation between the original Y and the residual of the independent variable X, from which the effects of all other independent variables used in the equation are removed. The interpretation of the square of the part correlation of Y and X_1 is the absolute increment of R^2 due to the addition of X_1 to the equation already containing all the other variables X_2 to X_k (Nie *et al.* 1975).

In three equations (38, 52, 57) the sums of the squared part correlation coefficients exceed 1. This can occur since, under certain conditions of negative correlation between the independent vari-

Climates covered by the equation	Equation number	Variables grouped into major types of factors						
		Climatic, Hydrologic		Relief, Area		Soil	Vegetation, Land use	Others
<i>Large basins</i>								
A	58	Q,	T	R,	A			
B	59	P, Q,	T		H		C	
	74	Q						
	76	Q			H			
B + D	72	Q		R				
B + D	73	Q						
C	60	Q,	T	R				
	61	Q,	T	R,	H, A			
	62	P		R,	H, A			
	78	Q,	q	R,	A	K,K	C,C,C,C	
	79						C	EB
	89	Q						TR
	90	Q			A			TR
	91	Q		R				TR
D	63	P			H	G	C	
	65	Q						
	66	Q					C	
	67	Q		R, D				
	68	Q						
	69	Q		R				
	70	Q		R				
	71	Q		R				
	88			R, L, A				
H (B?)	75	Q			H, A			
H (B?)	77	Q			A			
All climates	64	P, Q,	T		H, A	G	C	

ables, the increase in R^2 can be greater, when a variable enters a regression after another variable than when it enters in the first step (Ågren 1981, cf. Newton & Spurrell 1967). In these equations, intercorrelation between the independent variables makes the determination of the importance of individual variables especially difficult. Equation 52 exemplifies the intercorrelation problem. In equation 52, simple correlation between C_9 and $R_{11} = -0.84$ and simple correlation between C_9 and $G_2 = +0.68$. R_{11} has a negative simple correlation coefficient but a positive beta weight of 4.36. B , D_4 and G_2 have negative betas but positive simple correlation coefficients (Table

18). Intercorrelation makes it impossible to find out the significance of individual variables in this case.

The second question: The second question raised above was whether some variables are of unequal importance in different climates. This question is hard to answer satisfactorily, as Table 18 contains few equations with similar variables in different climates. Nevertheless, some inference can be made. Fournier's index p_m^2/P seems to be of about equal importance in the C climates of New South Wales as in the Am (Aw, Ca) climates of Queensland.

Discharge is another variable that can be considered. In equation 78 represent-

Table 18. Statistics to show the significance of variables in predicting sediment yield. A list of the symbols precedes Table 16. The precise meaning of each variable is given in the Appendix.
(1) = step at which included; 0.27 = not significant at 95 % level

Small basins									
Climate	Equ. Source		Variables				Soil	Veg.	Trap eff.
	No.	Statistics	Climate	Relief				Age	
Ca, Da (grass-lands)	(29)	Stall & Bartelli 1959	E ₃	R ₁	D ₁	A ₄		AGE ₁	TR ₃
		Simple corr. coeff.	0.45	0.71	0.18	-0.59		0.38	0.48
		Beta	0.75	0.27	-0.36	-0.17		0.40	0.32
		R ² change %	7(4)	50(1)	4(6)	10(2)		10(5)	7(3)
		Sq. part corr. coeff. %	16	4	4	2		10	9
BS	(38)	Hadley & Schumm 1961		R ₆	A ₁			TR ₆	
		Simple corr. coeff.		0.51	-0.58			0.16	
		Beta		-0.13	-1.37			1.03	
		R ² change %		1(3)	34(1)			55(2)	
		Sq. part corr. coeff. %		1	50			55	
Cs	(39)	Flaxman & Hobba 1955	P ₃	A ₂				C ₄	TR ₇ AGE ₁
		R ² change, fr. article %	15	17				10	31 6
	(40)	Scott et al. 1968	P ₂	I ₁			K ₁		
		Simple corr. coeff.	0.71	0.43			0.55		
		Beta	0.61	0.23			0.62		
		R ² change %	50(1)	4(3)			33(2)		
		Sq. part corr. coeff. %	26	4			37		
	(41)	Anderson 1949	q ₁	D ₂				C ₅	
		Simple corr. coeff.	0.94	0.31				0.03	
		Beta	0.93	0.11				-0.10	
Ca, Cb, Cc, Da (forest land)	(52)	Douglas 1973	P ₁₁	B	R ₁₀	D ₁	R ₁₁	G ₂	C ₉
		Simple corr. coeff.	0.32	0.29	-0.11	0.25	-0.16	0.04	0.32
		Beta	0.88	-0.23	-1.31	-1.91	4.36	-0.93	5.37
		R ² change %	15(2)	3(7)	24(6)	9(5)	8(4)	16(3)	10(1)
		Sq. part corr. coeff. %	45	3	25	33	35	28	48
Many climates (Western USA)	(53)	Flaxman 1966	I ₃	R ₇			K ₇	K ₈	
		Simple corr. coeff.	0.45	0.58			-0.33	0.44	
		Beta	0.32	0.50			-0.22	0.44	
		R ² change %	10(3)	33(1)			5(4)	22(2)	
		Sq. part corr. coeff. %	10	24			5	19	
	(54)	Flaxman 1972	P ₃ /T ₂	R ₇			K ₇	K ₉	
		Simple corr. coeff.	-0.23	0.49			-0.30	0.58	
		Beta	-0.44	0.82			-0.30	0.56	
		R ² change %	10(3)	40(2)			9(4)	33(1)	
		Sq. part corr. coeff. %	10	47			9	23	
(Queens-land)	(56)	Douglas 1973	P ₁₁	B	D ₁				
		Simple corr. coeff.	0.62	0.28	0.04				
		Beta	0.94	0.78	0.48				
		R ² change %	38(1)	26(2)	18(3)				
		Sq. part corr. coeff. %	71	42	18				
(Queens-land + N S Wales)	(57)	Douglas 1973	P ₁₁	B					
		Simple corr. coeff.	0.68	0.65					
		Beta	0.54	0.50					
		R ² change %	47(1)	23(2)					
		Sq. part corr. coeff. %	26	23					

Large basins			Variables										
Climate	Equ. No.	Source; Statistics	Climate			Relief		Soil	Veg. or Land use				
All climates	(64)	Jansen & Painter 1974	Q_4	P_8	T_3	A_3	H_1	G_1	C_7				
		Simple corr. coeff.	0.33	0.34	0.49	-0.24	0.36	0.30	-0.32				
		Beta	0.05	0.34	0.19	-0.32	0.32	0.18	-0.36				
		R ² change, %	0.1(7)	12(3)	24(1)	4(4)	7(5)	3(6)	9(2)				
		Sq. part corr. coeff. %	0.01	4	2	8	7	3	7				
C	(78)	Anderson 1954	q_4	Q_7		A_1	R_{13}	K_{10}	K_1	C_{10}	C_{11}	C_{12}	C_{13}
		Simple corr. coeff.	0.43	0.76		0.02	0.11	0.55	-0.55	0.04	0.01	0.08	0.44
		Beta	0.96	0.80		0.19	0.02	0.36	0.91	0.74	-0.30	0.43	0.11
		R ² change, forward %	3(3)	58(1)		3(5)		9(2)	1(6)	2(7)	3(8)	4(4)	1(9)
	R ² change, blockwise	22(2)	42(3)		0(1)	1(9)	10(4)	0(5)	2(6)	1(7)	6(10)	0(8)	
	Squared part corr. coeff., blockwise %	10	24		1	0	6	5	4	2	6	1	
	Climate												
	(82)	Guy 1964	I_4	Q_8	T_4	P_5	Q_9	M_1					
		Simple corr. coeff.	0.47	0.32	0.38	0.33	-0.08	0.05					
		Beta	0.38	0.83	0.45	-0.44	-0.25	-0.11					
R ² change %		22(1)	9(2)	12(3)	3(4)	2(5)	1(6)						
		Sq. part corr. coeff. %	10	16	14	5	2	1					
			Climate			Relief		Soil	Veg. or Land use				
D	(88)	Mc Pherson 1975				R_2	A_6	L_1					
		Simple corr. coeff.				0.46	-0.08	0.00					
		Beta				0.61	-0.43	0.45					
		R ² change %				21(1)	9(3)	2(2)					
			Sq. part corr. coeff. %			30	9	9					
	(65-74)	Lisitsyna & Alexandrova 1972	Q_5			R_8	D_3	C_8					
		(65) R ² change % from	74										
		(66) article	19					34					
		(67)	17		21	29							
		(68)	52										
(69)		43		16									
	(71)	56		38									
D+B	(72)		15		68								
	(73)		47										
B	(74)		42										

ing a C climate, discharge and peak discharge explain 61 % (R²) or 34 % (sq. part. corr. coeff.) of the variation in sediment yield. In equations 65-74 from the Soviet Union, mean annual discharge explains 17-74 % (R²) of the sediment yield variations in D climates; 15-47 %, in D+B climates; and 42 %, in an example of B climates (Table 18). Thus, there are great differences between regions with similar climates. An example of an inverse relationship between sediment

yield and discharge is found in equation 68 for the Carpathians (cf. the Appendix). This was attributed to the increase in woodland associated with elevation, and the fact that discharge increases with elevation - more precipitation, less evapotranspiration.

Equation 59 from B climates is of special interest in arriving at an answer to question two. The original equation from Jansen & Painter (1974) is given since it has not been re-calculated.

Table 19. Variation in sediment yield accounted for by the major factors.

Small basins

Climate	Equ. Source; No. Statistics	Whole Ar- equ.	Cli- ticle	mate	Relief	Soil Geol- ogy	Land use	Trap eff. Age
Ca, Da (grass- land)	(29) Stall & Bartelli 1959 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	87.8 44.4	88 16	(E ₃) 7 10	(R ₁ D ₁ A ₄) 64 10			(TR ₃ .AGE ₁) 17 18
B	(38) Hadley & Schumm 1961 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	89.8 105.1			(R ₆ A ₁) 35 50			(TR ₆) 55 55
	(39) Flaxman & Hobba 1955 ΣR^2 from article %		80	(P ₃) 15	(A ₂) 17		(C ₄) 10	(TR ₇ .AGE ₁) 37
Cs	(40) Scott et al. 1968 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	87.1 65.9	86 29	(P ₂ I ₁) 54 29		(K ₁) 33 37		
	(41) Anderson 1949 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	90.8		(q ₁) 88 79	(D ₂) 2 1		(C ₅) 1 1	
Ca, Cb, Cc, Da (forest land)	(52) Douglas 1973 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	84.3 182.0	79.5	(P ₁₁) 15 45	(B,R ₁₀ D ₄ R ₁₁)(G ₂) 44 96	(G ₂) 16 28	(C ₉) 10 48	
Many climates (Western USA)	(53) Flaxman 1966 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	70.5 58.1	81	(I ₃) 10 10	(R ₇) 33 24	(K,K ₈) 27 24		
	(54) Flaxman 1972 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	91.8 88.9	92	(P ₃ /T ₂) 10 10	(R ₇) 40 47	(K,K ₈) 42 32		
(Queens- land)	(56) Douglas 1973 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	82.2 131.6	82.3	(P ₁₁) 38 71	(B,D ₄) 44 60			
(Queens- land+NS Wales)	(57) Douglas 1973 ΣR^2 change % $\Sigma Sq.$ part corr. coeff. %	69.6 49.1	54.7	(P ₁₁) 47 26	(B) 23 23			

$$\log SAY_7 = 12.133 - 0.340 \log Q_4 + 1.590 \log H_1 + 3.704 \log P_8 + 0.936 \log T_3 - 3.495 \log C_7 \dots\dots\dots (59)$$

where

SAY₇ = sediment yield in tons/km²/yr

Q₄ = annual discharge in 10³ m³/km²
(= mm)

H₁ = altitude in meters above sea level

P₈ = annual rainfall, mm

T₃ = average annual temperature, °C

C₇ = natural vegetation (1-4)

As the rainfall variable and the runoff variable have comparable units, the re-

gression coefficients are comparable. Precipitation is a more important variable than runoff, as the coefficient is larger. What might appear strange is the negative sign of log Q₄. The reason for a negative sign in a B climate may be that the equation also contains a rainfall variable and that Q₄ is also a measure of vegetation cover. C₇ is a very rough estimation, and the effect of vegetation does not enter the regression to any great extent. The equation from A and C climates, by the same author, have postive signs of log Q₄ (cf. the Appendix). One conclusion that can be drawn from this

Large basins

Climate	Equ. Source; No. Statistics	Whole Ar- equ.	Cli- mate	Relief	Soil Geol- ogy	Land use	Trap eff. Age
All climates	(64) Jansen & Painter 1974			(Q₄P₈T₃)(A₃H₁)	(G₁)	(C₇)	
	ΣR ² change %	58.5	57.9	36	11	3	9
	ΣSq. part corr. coeff. %	32.0		7	15	3	7
C	(78) Anderson 1954			(q₄Q₇)	(R₁₃A₁)	(K₁₀K₁)	(C₁₀C₁₁C₁₂C₁₃)
	ΣR ² forward %	83.9		61	3	10	9
	ΣR ² blockwise %	83.9		63	1	10	9
	ΣSq. part corr. coeff. %	58.9		33	1	12	13
D	(88) Mc Pherson 1975			(R₂A₆L₁)			
	ΣR ² change %	31.2	25	31			
	ΣSq. part corr. coeff. %	48.0		48			
	(65-74) Lisitsyna & Alexandrova			(Q₅)			
	(65) 1972 ΣR ² change from		74	74			
	(66) article, %		53	19			
	(67)		67	17			
	(68)		52	52			
	(69)		59	43			
	(71)		94	56			
					50 (R ₈ D ₃)		34 (C ₈)
D+B	(72)		83	15			
	(73)		47	47			
B	(74)		42	42			
					68 (R ₈)		

example is that the variable "mean annual runoff" may have an additional meaning in B climates.

The third question: This question deals with the strength of the major groups of factors in different climates. The discussion is based on the rather meagre items of information contained in Table 19. However, certain general conclusions concerning small basins may be drawn.

In equations containing soil variables (one from a Cs climate and two from the western United States), the soil factor explains a high percentage of the variation in sediment yield. In equations from Cs climates (40, 41) and the Am (Aw, Ca) climates (56), the climatic factor explains a large part of the variation in sediment yield. In C climates, except for the Cs climates, (29, 52), in the entire western United States (53, 54) and in B climates (38, 39), the relief and area factor is important. In equations which refer to large basins, discharge is important in explaining the variation in

sediment yield in C, D and B climates. However, in some cases the relief factor is of great importance.

The fourth question: The problem of which climatic variables should be used in an equation which would be valid for many different climates, is discussed with reference to four of the equations (53, 54, 57, 64) in Table 18 and with reference to the answer given to the second question. In two equations (53, 54), intensity and the ratio of precipitation/temperature both explained 10 % of the variation in sediment yield. Fournier's index p_m^2/P explained 47 % of the variation in equation 57. Catchments in B climates were not, however, covered by the latter equation. In fact, Fournier (1960) got different relationships for arid and humid areas (Fig. 17). Therefore, in a mixture of basins from both arid and humid climates, this index might not give as good a result.

In equation 64, developed for large basins, precipitation and temperature

explain more of the variation in sediment yield than does discharge. The coefficient of discharge is not significant at the 95 % probability level, and this indicates that discharge is not a suitable variable. The elimination of discharge in equations containing both humid and arid areas is in agreement with the observations made in the discussion of question two.

Concluding remarks: This section has to some extent been an attempt to test the validity of Jansen and Painter's (1974) and Sundborg's (1978) notion that processes are controlled by a multitude of different factors which are of varying significance in different climatic environments. In conclusion, let us consider equations 58–64 (cf. the Appendix). As can be seen, the variables are not the same in all climates. The coefficients for the variables vary for equations in the respective zones. To some extent, this is due to intercorrelation between accepted variables or because some variable may be missing. Equation 64 for all climates explains less of the sediment yield variation than those for individual zones, although it includes the most variables. This may be explained by the unequal importance of different factors in different climates.

10.2 Delivery ratio equations

One method of calculating sediment yield from small drainage basins is to estimate the gross erosion and then use a delivery ratio equation. Gross erosion includes material from sheet, rill, gully and channel erosion. The detached material may originate from fields, construction sites, spoil heaps, roads and slide scars. Delivery ratio is the percentage of the eroded material that is delivered from the erosion source to a downslope or downstream location. It is affected by such factors as land use and vegetation, amount of overland flow,

size of the transported material, channel density (which affects proximity to streams), relief, depositional areas and catchment size. So variables describing all or most of these factors should be included in a universally applicable delivery equation. There are, however, no delivery relationship available that can be applied under all circumstances, although a number of sediment delivery ratio equations for specific areas do exist.

Roehl (1962) developed the following equation for the southeast Piedmont region in the USA:

$$\log DR = 4.50047 - 0.23043 \log 10 A - 0.51022 \log R - 2.78594 \log B \quad (92)$$

- where
- DR = delivery ratio, in percent
- A = catchment area, in square miles
- R = relief/length ratio, in feet/feet.
The average stream length is the sum of the averages for each stream order
- B = weighted mean bifurcation ratio.
Mean of the ratio of the number of streams in successively higher stream orders

All these variables affect the peak flow of high waters. In addition, catchment area affects length of transport, and thus the chances of deposition. The multiple correlation coefficient R was found to be 0.961.

Williams (1977) developed a delivery ratio equation for 35 sub-basins in Little Elm Creek, Texas:

$$DR = 1.366 \times 10^{-11} A^{-0.0998} R^{0.3629} CN^{5.444} \dots\dots\dots (93)$$

- where
- DR = delivery ratio of sheet and rill eroded material
- A = drainage area in km²
- R = relief/length ratio, in m/km
- CN = the long-term average curve number, used to calculate runoff (described in chapter 11)

This model only considers material originating from sheet and rill erosion. Drainage area and relief/length ratio affect time lapse and the intensity of peak flows. Infiltration, expressed by CN, is important in determining the runoff volume. The importance of the variables can be seen from their standard partial regression coefficients which are 0.71 for curve number, 0.54 for relief/length ratio and 0.22 for drainage area. The coefficient of multiple determination R^2 was 0.93.

A delivery ratio regression equation can be used to estimate the delivery of material to the basin outlet. Another way of estimating the delivery of material is to use a routing equation. The technique of sediment routing means that the flow of sediment is more or less followed from different parts of the catchment to the basin outlet. Routing equations are not regressions, but a brief description of two routing equations will be presented in order to make a comparison with delivery regressions. Onstad & Bowie (1977) divided the basin into rectangular incremental units and calculated annual gross erosion. Sediment was routed from the center of each increment to the outlet of the basin by the function:

$$\Sigma S_i = \sum_{i=1}^n S_{oi} e^{-\beta T_i} \dots \dots \dots (94)$$

where

S_i = sediment contribution at the basin outlet from increment i

S_{oi} = total erosion from increment i

β = routing coefficient

$T_i = L_i^{0.77}/R_i^{0.385}$ where

L_i = "channel" length = distance from increment centre to basin outlet

R_i = "channel" slope = ratio of the relief from increment centre to basin outlet to the "channel" length

n = number of incremental areas

In a study by Williams (1977), sediment yield from small sub-basins was estimat-

ed, and the sediment material was routed through the channel system to the outlet of the large basin. William's sediment routing equation is most interesting since it refers to the transportation processes which are closely related to the physical behaviour of individual particles. Deposition within the drainage ways is assumed to be a function of the amount of sediment from the sub-basin, the travel time to the basin outlet and the settling velocity, expressed by the median particle diameter.

$$S = \sum_{i=1}^n S_i e^{-\beta T_i} \sqrt{d_i} \dots \dots \dots (95)$$

where

S = sediment yield from an individual storm for the entire basin

S_i = sediment yield from subbasin i , predicted by the Modified Universal Soil Loss Equation (cf. chapter 11)

β = routing coefficient

T_i = travel time from sub-basin i to the basin outlet computed by flood-routing

d_i = the median particle diameter of sediment for subbasin i

n = number of sub-basins

Both the routing equations 94 and 95 stress distance as one of the main variables, represented by T_i in both equations.

Sediment delivery ratios in two loessial catchments near Treynor, Iowa, were studied for 7 years by Piest *et al.* (1975). The annual sediment delivery ratio ranged from 4 to 72 % with a 7-yr mean value of 46 %. Piest *et al.* tried to explain the variation in sediment delivery by looking at sediment delivery on the basis of individual storms in a contour-planted, continuous corn catchment. During 52 of the best-sampled runoff events in one of the catchments, antecedent soil moisture content, season and the rainfall factor of the USLE ac-

counted for 83 % of the variation in the sediment delivery ratio. During the same season (31 storms in June), peak runoff rate, runoff volume and antecedent moisture content accounted for 96 % of the variation in the sediment delivery ratio.

The delivery ratio of about 50 % in this loessial area can be compared to the estimated delivery ratio of nearly 100 % in the hilly loess areas of the Middle Yellow river basin (Gong & Xiong 1980, Jiang & Wu 1980). The exceptionally high delivery ratio in the Yellow river basin is probably due to steep relief, scanty vegetation, easily transported material and heavy rainstorms.

In conclusion it can be said that the sediment delivery ratio within a region is principally a function of the runoff variables, particularly the volume and peakedness of individual flows. Parameters that affect these runoff variables can be used in a sediment delivery equation. Because delivery ratio equations are so few, and were developed in similar climates, it is impossible to distinguish climatic differences. However, peak flow differences are dependent on surface runoff, which might imply that climates with a high surface runoff experience a high rate of sediment delivery.

11 Empirical models based on results from plot and slope studies

In the preceding two chapters, erosion and transport has been analysed in statistically developed models for drainage basins. In order to find out more about the significance of different factors, especially the climatic factor, to erosion processes it is necessary to study the processes under controlled conditions. This can only be done on a plot scale. Measurements of soil loss from plots under different conditions have provided the basic data that are required for the construction of equations of gross erosion and of sediment yield from very small basins.

As early as the 1940's, empirical equations were developed to describe the relation between different factors and soil loss from cultivated and bare fields. At present, the equation most commonly used in the USA to predict areal erosion is the so-called Universal Soil Loss Equation (USLE). This equation is based on data obtained from more than

10,000 plot years in the eastern USA. The general form of the equation is (Wischmeier & Smith 1965, Arnoldus 1977a):

$$A = R K L S C P \dots\dots\dots (96)$$

where

A = soil loss in tons per unit area and time interval

R = rainfall factor

K = soil erodibility factor

L = slope length factor

S = slope gradient factor

C = cropping-management factor

P = erosion-control practice factor

Detailed presentations of the USLE are given in the Agriculture Handbook 282 by Wischmeier & Smith (1965) and in Arnoldus (1977a). A brief description of the equation is given in the following paragraphs.

The calculation of rainfall erosivity is

essential to any prediction of erosion. The parameter that was found to give the best results under different precipitation and soil conditions was EI_{30} (Wischmeier & Smith 1958). Therefore, the EI_{30} index was accepted as the R factor in the USLE. The EI_{30} for a rainstorm is calculated as follows: The precipitation record is divided into periods with approximately the same rainfall intensity. The kinetic energy per cm of rain for each intensity class is calculated according to equation 2 (cf. section 3.1.3) or taken from a table (cf. Wischmeier & Smith 1958) and is then multiplied by the amount of rain in cm at that intensity. Energy terms for each intensity class are summed for the whole storm. This total energy value, E, is multiplied by the maximum 30-minute intensity multiplied by two, I_{30} , and divided by 100. In the metric system EI_{30} is expressed by the unit $\frac{\text{ton m}}{\text{ha}} \cdot \frac{\text{cm}}{\text{h}} \cdot 10^{-2}$. Joules/m² is sometimes used in the expression instead of ton m/ha. The numerical difference between these units is small as $1 \text{ tm/ha} = 0.9807 \text{ J/m}^2$.

In order to estimate an annual rainfall factor, the EI_{30} values for individual storms >12.7 mm (0.5 inch) during the year are summed. In the USLE equation, the rainfall factor was originally calculated as an average annual value i.e. the number of EI-units in a normal year's rain (Wischmeier 1962). The equation was also originally meant to estimate long term average annual sheet and rill erosion.

The soil erodibility factor K for a given soil is obtained by measuring the soil loss from a plot under specified standard conditions. These standard conditions include a 22.1 m (72.6 ft) long, 9 % steep plot which has been fallow for at least two years and a normal annual rainfall pattern. The soil loss is then divided by the rainfall erosivity value, EI_{30} , measured during the test period. K is defined

as soil loss per EI_{30} unit (cf. section 5.1.5).

The other factors are calculated as ratios of soil loss under the conditions in question to the soil loss from standard plot conditions. The numerical values of these ratios can be taken from manuals when slope length, slope gradient, cropping-management conditions and erosion control practices are known (Wischmeier & Smith 1965, Arnoldus 1977a). C factors for natural vegetation are found in articles by Wischmeier (1974, and 1975).

Continuous precipitation records are needed to compute the EI_{30} index. Such records are not available on a world wide basis. Even in the eastern USA, where Wischmeier (1962) intended to draw a map of the mean annual erosivity index, there were too few values to plot isolines. Wischmeier (1962) obtained a high correlation between the average annual EI_{30} index and the product of the mean annual rainfall, the 2-year 1-hour maximum rainfall, and the 2-year 24-hour maximum rainfall for 77 meteorological stations east of the Mississippi river. One regression equation of the general form of equation 97 was derived for each of three regions in the eastern United States (cf. Bergsma 1981)

$$EI_{30} = a \cdot P \cdot I_{1,2\text{yr}} \cdot I_{24,2\text{yr}} + b \dots\dots\dots (97)$$

where

P = mean annual rainfall

$I_{1,2\text{yr}}$ = maximum 1-hour rainfall with a return period of 2 years

$I_{24,2\text{yr}}$ = maximum 24-hour rainfall with a return period of 2 years

a, b = constants

In the western USA, long duration recording raingauges did not cover the area sufficiently to provide data for the calculation of EI_{30} values for a map. Ateshian (1976) found the 2-year 6-hour maximum rainfall to be the critical rainfall for estimating the mean annual erosivity index, EI_{30} . He prepared the equations:

$$EI_{30} = 16.55P_6^{2.2} \dots\dots\dots (98)$$

$$EI_{30} = 27.00P_6^{2.2} \dots\dots\dots (99)$$

where

P_6 = maximum 6-hour rainfall with a return period of two years

(98) is valid for the coastal side of the Sierra Nevada and the Cascade mountains of California, Oregon and Washington, the Hawaiian Islands and Alaska. (99) refers to the remaining United States.

In semi-arid and arid climates, an index which includes all occurrences of rainfall, e.g., rains of short duration or of low intensity, can be misleading. The 10-year 6-hour maximum rainfall might be an even better indicator than the 2-year 6-hour rainfall. Roose (1977) mentioned that observations from the Mediterranean and Saharan zones indicated that the decennary and centenary exceptional rains were of decisive importance for the transformation of the landscape.

In West Africa, analysis of the relation between daily rainfall amounts and the EI_{30} index for eight recording raingauges resulted in two regressions (Roose 1977). In the Ivory Coast, within 40 km of the sea, a linear regression was obtained for the "monsoon" rains from June to September, and a curvilinear regression for the rains during the rest of the year. The same curvilinear regression was also obtained for the inland stations in the Ivory Coast and Upper Volta. A simple relation was found between the mean annual EI_{30} over a 5 to 10 year period and the mean annual rainfall in mm, P_{annual} , during the same period.

$$EI_{30} = 0.5 P_{\text{annual}} \pm 0.05 \dots\dots\dots (100)$$

This relation was found to hold for a number of locations in the Ivory Coast, Upper Volta, Senegal, Niger, Tchad, the Cameroon and Malagasy, with the exception of stations situated near mountains and on the coastal plain (Roose 1977).

Another index that can be regressed to the EI_{30} index is a modification of Fournier's index: $\sum_{i=1}^{12} p_i^2/P$ where p_i is monthly precipitation and P is annual precipitation. This index has been successfully used in Belgium (Bollinne *et al.* 1979) as well as in the USA and Morocco (Arnoldus 1977b), but the constants in the regression, differ from one region to another.

The USLE was originally developed for the eastern USA, and EI_{30} can certainly not be expected to be the best climatic variable in predicting soil loss in all other climates of the world. The EI_{30} index has been tested in different parts of the world with varying results. A modification of the EI_{30} index for Nigeria takes into account the threshold EI_{30} value required to initiate runoff, which, in turn, is related to the time it takes to reach maximum storm intensity (Wilkinson 1975). A second index used in Nigeria is AI_m , which gave as high (Lal 1976) or higher (Lal 1977) correlations with single storm soil losses as did EI_{30} . The AI_m of a storm is the storm amount of rain, in cm, times the maximum intensity over a 7.5 min. interval, in cm/h.

In Tobago, Lesser Antilles, the correlation between soil loss and the EI_{30} index was low, since storms lasting for 30 minutes are the exception rather than the rule, and since wind plays a great part. EI_{15} was found to be better, but explained only 34 to 69 % of the soil loss from plots 1.5×5 m with an inclination of 10° and still less for steeper plots (Ahmad & Breckner 1974). One source of error is the EI_{30} calculation itself, since the computation of energy (E) may involve considerable uncertainties. With the exception of a few studies (Elwell & Stocking 1973, Moore 1979), equation 2 by Wischmeier & Smith (1958) is used to calculate energy (Ahmad & Breckner 1974, Ramaiah & Sreenivas 1975, Wilkinson 1975, Babu *et al.* 1978), despite

the fact that there may be different types of drop size distribution in tropical rains compared to non-orographic rains of temperate zones for the same intensity (cf. chapter 3). Especially for typhoons, the drop sizes and impact velocities differ from ordinary rain (Lam & Cheng 1977).

Hudson (1971) found that for one research station in Rhodesia, the kinetic energy for precipitation > 25 mm/h ($KE > 25$) was a slightly better index of erosivity than EI_{30} . Stocking & Elwell (1973) found EI_{30} to be the most reliable parameter for storm soil losses from plots at two research stations in Rhodesia. Instead of $KE > 25$ Joshua (1977) suggested the kinetic energy for precipitation greater than the infiltration rate of the soil as an erosivity parameter in Sri Lanka. In Indonesia and Java, however, the EI_{30} index yielded a high correlation with soil loss (Arnoldus 1977a).

One of the limitations of the EI_{30} index as a general climatic factor in temperate climates is that it fails to reflect the erosive potential of runoff from thaw and snowmelt runoff which was also pointed out by Wischmeier (1976). In Sweden, for example, amount of snow, frost conditions and temperature conditions during the snowmelt are climatic elements that affect the erosion processes, and are most important factors for the total soil loss.

Even if EI_{30} is not accepted as an ideal indicator of the erosivity of rain in all climates, it may be of interest to study its regional and climatic variations. This is illustrated in six maps (Figs 21–26). The maps show mean annual values of EI_{30} and are all converted to comparable units in the metric system. The erosivity map of the eastern USA (Fig. 21) is based on 181 EI_{30} values calculated from 22-year rainfall records, and about 1700 additional values estimated by means of equation 97. For the western states, the

location of the isolines is based on values calculated by equations 98 and 99. The western USA has a much lower erosivity than the eastern USA which does not imply that the soil loss must be lower, since other factors may counteract this tendency.

The erosivity map of the USA may be compared with an erosivity map of West Africa (Fig. 22). EI_{30} values were estimated by inserting daily rainfall records from a 20 to 50 year period in Roose's equations mentioned above.

Two other erosivity maps from tropical climates are included in this presentation. The map of Rhodesia (Fig. 23) is based on direct calculations of EI_{30} from automatic raingauge charts from 33 meteorological stations. Regression equations relating the EI_{30} index to daily rainfall were developed for different physiographic regions. The equations were used to calculate EI_{30} for another 400 stations with more than 10 years of rainfall data (Stocking & Elwell 1973). The erosivity map of India (Fig. 24) is based on records from 44 stations, and on calculations by a regression equation relating the EI_{30} index to the mean annual rainfall for 180 other stations (Babu *et al.* 1978).

The maps from A climates in Figs 22, 23, 24 can be compared with an erosivity map of Belgium from a Cb climate in Fig. 25. The map prepared by Bollinne *et al.* (1979) is based on EI_{30} calculations of rainfalls > 1.27 mm at four stations. A regression equation relating the four EI_{30} values to mean annual precipitation was obtained. With the help of this equation, erosivity indices were then calculated for 480 other rain-gauge stations. The erosivity in Belgium as indicated by EI_{30} is closely related to topography, since the precipitation distribution is mostly dependent on topography. A provisional rain erosivity map of the Netherlands was compiled using a version of equation 97 (Bergsma 1980). Data from 21 meteo-

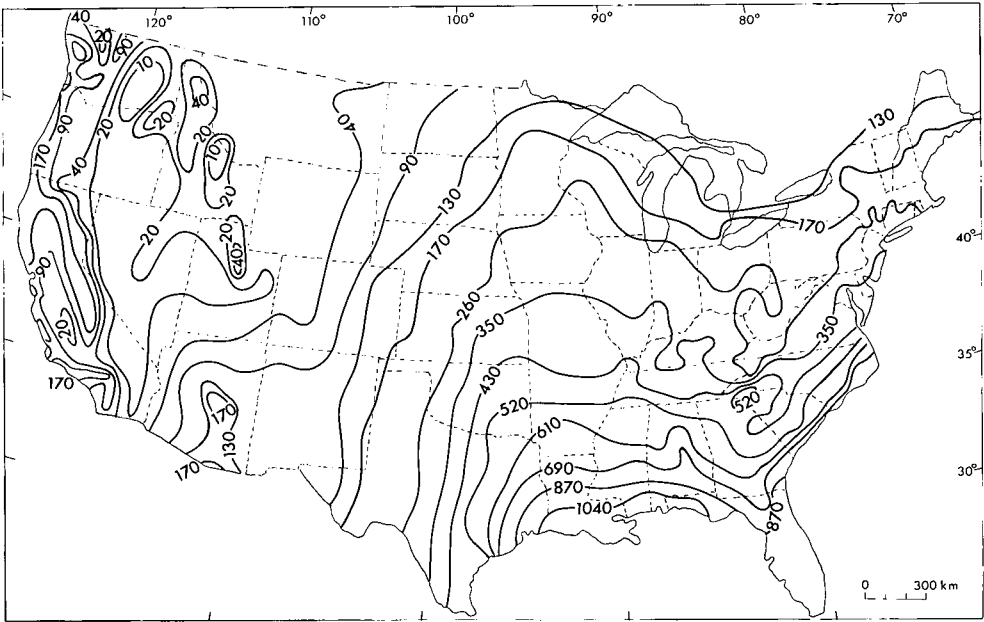


Fig. 21. Mean annual erosivity in the United States estimated by the EI_{30} index. Conversion from US units to nearest tens of metric units through multiplication by 1.735 (from Wischmeier & Smith 1965, USDA, SCS 1975 and Ateshian 1976). USDA, SCS (1975) recommended that 610 should be the maximum EI_{30} value used in the USLE in the United States.

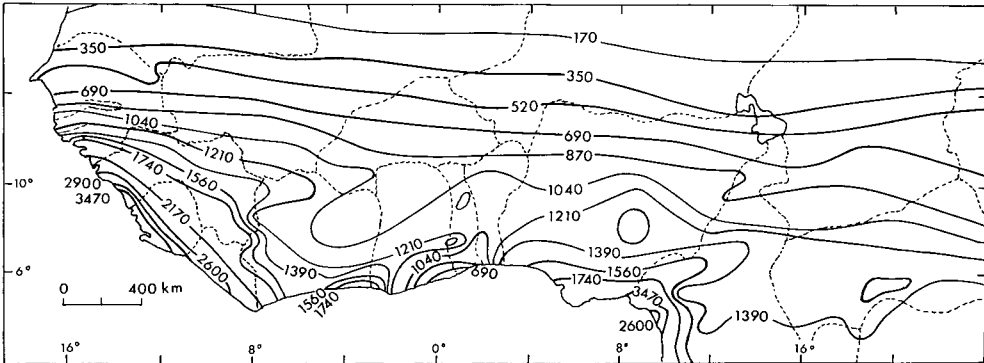


Fig. 22. Mean annual erosivity, EI_{30} in West Africa. Conversion from US units to nearest tens of metric units through multiplication by 1.735 (from Roose 1977, after Roose 1975).

rologic stations were used. The map is not reproduced in this study because of its small variations in EI_{30} . The mean EI_{30} was found to be approximately 74 in metric units and to vary between 57 and 85.

The map of Morocco, prepared by Arnoldus (1977b), refers to Cs and B climates (Fig. 26). It is based on three EI_{30} values, which were regressed to the modified Fournier index. The regression equation was used to estimate EI_{30} for

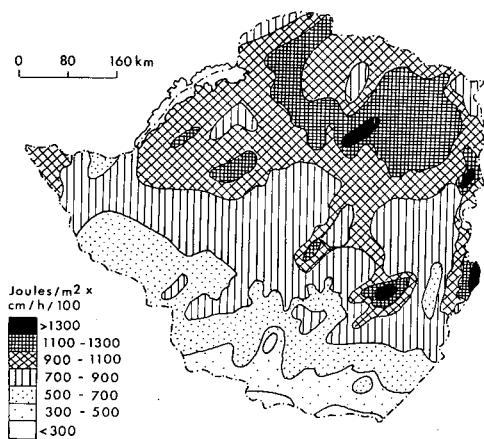


Fig. 23. Mean annual erosivity in Rhodesia. EI_{30} was originally expressed in $J/m^2 \cdot mm/h$ (from Stocking & Elwell 1973).

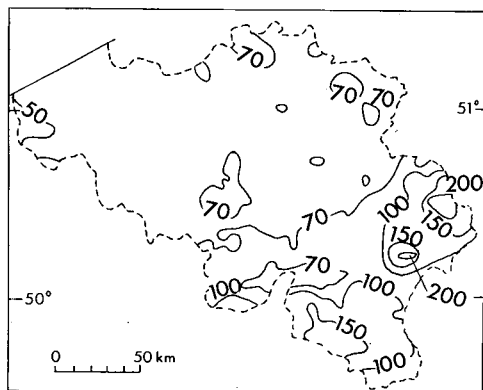


Fig. 25. Mean annual erosivity in Belgium. EI_{30} in metric units (from Bollinne *et al.* 1979).

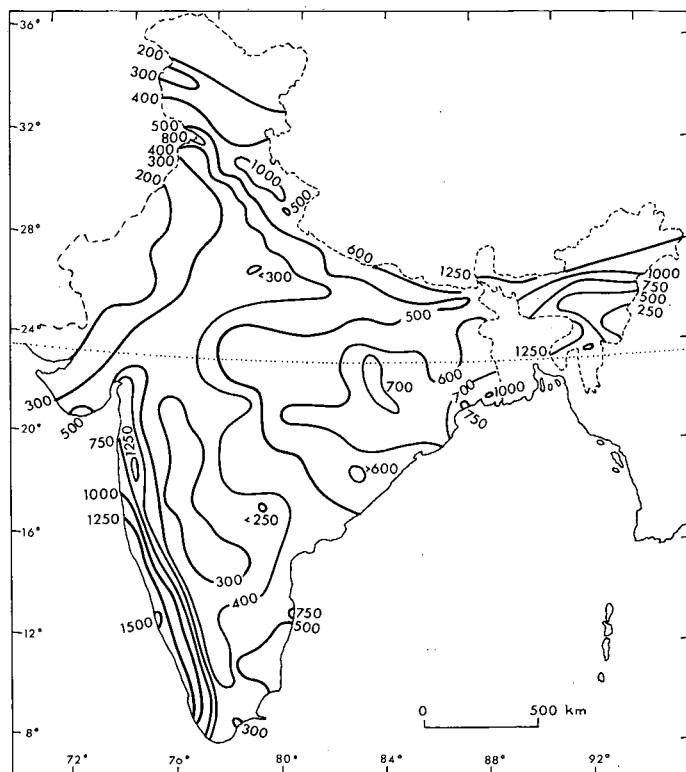


Fig. 24. Mean annual erosivity in India. EI_{30} in metric units (from Babu *et al.* 1978).

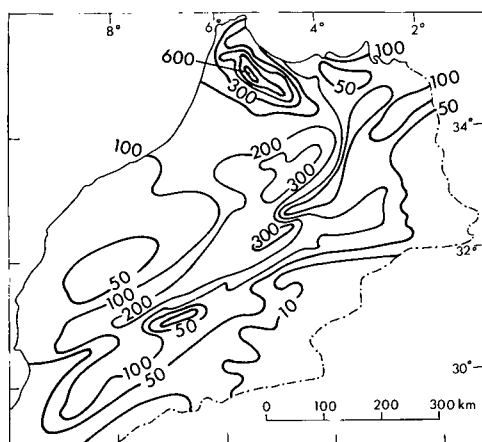


Fig. 26. Mean annual erosivity in Morocco. EI_{30} in metric units (from Arnoldus 1977b).

112 stations. Comparison with the values for California (Fig. 21), which has a similar climate, indicates that higher values occur in the mountains of Morocco.

As can be seen from the maps in Figs 21–26, the erosivity is much higher in tropical and subtropical high rainfall climates than in tropical and subtropical low rainfall climates, temperate humid and continental climates. This condition can be explained by differences in rainfall intensities and amounts.

The remaining factors in the USLE equation also vary with climate. The C factor for natural vegetation is high in semi-arid climates. In climates with seasonal rainfall and with snow, there are seasonal variations in the C factor, whereas in other climates the C factor may be more constant. The K factor is reported to be low in many tropical regions (cf. section 5.3.6). Crust formation in savanna climates will increase the K factor.

The Soil Loss Estimator for Southern Africa (SLEMSA) is a model which is similar to the USLE and which was developed for the Rhodesian high veld (Elwell 1979, Elwell 1981). Rainfall energy is used as a climatic factor, and the basic soil erodibility is modified by

taking into account the influence of tillage practice.

The Universal Soil Loss Equation was originally designed for more or less homogeneous fields. The great need for an equation which is applicable to drainage basins has resulted in attempts to adapt the USLE for use in small basins. When basins are considered, weighting of individual factors must be introduced. The weighting techniques are described in articles by Williams & Berndt (1972, 1976). However, the weighting is linear, whereas the factors may be non-linear. Furthermore slope length is roughly estimated in a catchment, even though sediment yield from a drainage basin is very dependent on how far the detached material is transported. Such transport is performed by overland flow, and theoretically, overland flow should therefore enter into the equation, and/or the factors should be weighted for only those areas which usually contribute to overland flow.

Some researchers have correlated sediment yield from catchments with different climatic and hydrologic variables. In two loessial basins, 1.9 km² and 1.7 km² in Nebraska, the best single predictor of sediment yield was $Q + q$, i.e. the volume of runoff in inches plus peak rate of runoff in inches per hour (Dragoun 1962). The parameter ($Q + q$) accounted for 81 % and 91 % of the variation in sediment yield in the two catchments while the EI_{30} index only accounted for 56 % and 54 %.

In a 0.32 km² loessial basin in Iowa, the peak runoff rate explained 80 % of the variation in sediment yield (Piest *et al.* 1975).

In a third study, dissimilar results were obtained in small and large basins (McGuinness *et al.* 1971). A high correlation was obtained between sediment yield from two basins of 0.04 km² and 0.08 km² in area and the EI_{30} multiplied by the vegetation cover factor C of the USLE.

Sediment yield from the large catchment of about 15,000 km² was closely correlated with runoff.

In agreement with the results of the studies by Dragoun (1962) and others, sediment yield models for small catchments which are based on the USLE, should include runoff in one way or another. One modified version of the USLE (Onstad & Bowie 1977, cf. Onstad & Foster 1975, and Foster *et al.* 1977b), is designed to be used in segments of small catchments on a single storm basis. The modified rainfall factor R_m is composed of one rainfall term and one runoff term:

$$R_m = aR_{st} + 0.22(1-a)Qq^{1/3} \dots\dots (101)$$

where

R_{st} = storm rainfall factor, EI_{30} units

Q = storm runoff volume, m³ per unit area

q = storm peak runoff rate, m³/h per unit area

a = parameter that indicates the relative importance of rainfall energy compared with runoff energy.

The constant 0.22 was determined from plot data in Indiana and Minnesota (Onstad & Bowie 1977). R_m combines more units than the original R in one index. However, by using runoff, it compensates for soil moisture differences. There is a comparable modification of the R factor in an annual version of the model (Onstad & Bowie 1977) which is intended to be used in combination with a routing function (section 10.2).

In order to overcome the problem of variations in sediment delivery ratio between different years, Williams (1974, 1975) and Williams & Berndt (1976) modified the USLE to a single storm sediment yield model for a whole catchment called the MUSLE, by exchanging the EI_{30} factor with a runoff factor:

$$A = \alpha(Qq)^\beta K L S C P \dots\dots\dots (102)$$

where

A = sediment yield from an individual storm, in tons

Q = storm runoff volume, in m³

q = peak runoff rate, in m³/s

α, β = model parameters

The K factor is calculated in the usual way by using the erosivity EI_{30} . The model parameters were found to be $\beta = 0.59$ and $\alpha = 11.8$ in metric units. In tests using these α and β values in 60 catchments in 6 states of the USA, the MUSLE generally explained about 80 % of the variation in storm sediment yield. In another test in two basins in Iowa α was found to be 7.24 and β , 0.84 (Onstad & Bowie 1977).

It is unsatisfactory that the variables of the MUSLE equation are weighted for the whole basin and not for the potential surface runoff area of the basin. Wigham & Stolte (1973) solved this problem by evaluating the factors in the USLE for areas within the catchment which had well-defined channels.

The MUSLE equation is a parametric black box model for drainage basins. The Stanford Sediment Model (Negev 1967) is a grey box parametric model where the erosion processes are divided into three separate processes: splash detachment, transport by thin overland flow and transport in rills and gullies. The functions of the key processes are:

$$D_{rain} = K_1 I^\alpha \dots\dots\dots (103)$$

$$T_{rain} = K_2 (\sum D_{rain}) Q^\beta \dots\dots\dots (104)$$

$$T_{rill, gully} = K_3 Q' \dots\dots\dots (105)$$

where

D_{rain} = hourly quantity of soil splash, in tons

T_{rain} = hourly quantity of soil splash pickup by thin overland flow, in tons

$T_{rill, gully}$ = hourly quantity of sediment contributed from rills and gullies, in tons

K_1 = a parameter that varies with the soil type and cover
 K_2 = a parameter of land characteristics affecting flow velocities
 K_3 = a parameter of topographical and geographical features of the land surface
 I = hourly rainfall, in inches
 ΣD_{rain} = quantity of available material, in tons
 Q = hourly overland flow, in inches
 α, β, γ = exponents

The exponents and parameters must be optimized by statistical methods, for application in a particular region and the parameters in the hydrological model used for the calculation of overland flow are lumped.

The sediment model by Donigian & Crawford (1976) is a later model that has been influenced by the concepts of Meyer & Wischmeier (1969), Onstad & Foster (1975), and the Stanford Sediment Model (Negev 1967). This model (Donigian & Crawford 1976) may be considered a white box parametric model as sediment yield may be detachment limited or transport limited.

All these models require estimations of surface runoff. The so called curve number technique is a lumped approach which is used to estimate surface runoff volume from basins (Holtan & Overton 1964, Hudson 1971, FAO 1976b). Curve numbers are derived directly as numbers from a table which lists all possible combinations of land use, conservation practice, hydrologic conditions of soil cover and hydrologic soil group. The probable quantity of direct runoff expected from different quantities of rainfall is then obtained from diagrams with curves representing the curve numbers. The estimations can also be corrected for variations in rainfall over a basin (Allis 1962).

The 'partial contributing area concept' has been the basis for some researchers

for modeling overland flow (Engman & Rogowski 1974, Beven & Kirkby 1979). The Stanford Hydrology Model (Crawford & Linsley 1966, Donigian & Crawford 1976) has functions that represent the percentage of areal variation in infiltration around a mean value. However, models with lumped infiltration values for large areas and parameters that need optimization are most common (cf. Fleming 1975).

A different approach—in comparison with the USLE and similar equations—was introduced by Bobrovitskaya *et al.* (1977). They derived an equation based on the assumption of a close connection between rate of erosion and rill magnitude. The result was an empirical relationship between sediment yield and runoff depth on slopes and in small basins:

$$SAY_p = aVcQ_p^N \dots\dots\dots (106)$$

where

SAY_p = sediment yield/area with the probability of a higher value p % of the runoff period, t/ha
 Q_p = runoff during the snowmelt or rainfall flood with p % probability of a higher value, mm
 a = coefficient taking into account the influence of the landuse practice of the previous year
 V, N = parameters depending on the type of rill network on the slope, agrotechnical practice and soil types (Tables 20 and 21)
 c = coefficient for the influence of slope gradient upon soil erosion. If the gradient $S > 10$ % $c = 0.01$, if $S < 10$ % $c = 1$

Type I rills run between the micro-relief at the soil surface. Type II rills are formed on the lower parts of slopes due to the confluence of the microrills. Type III rills are formed through the con-

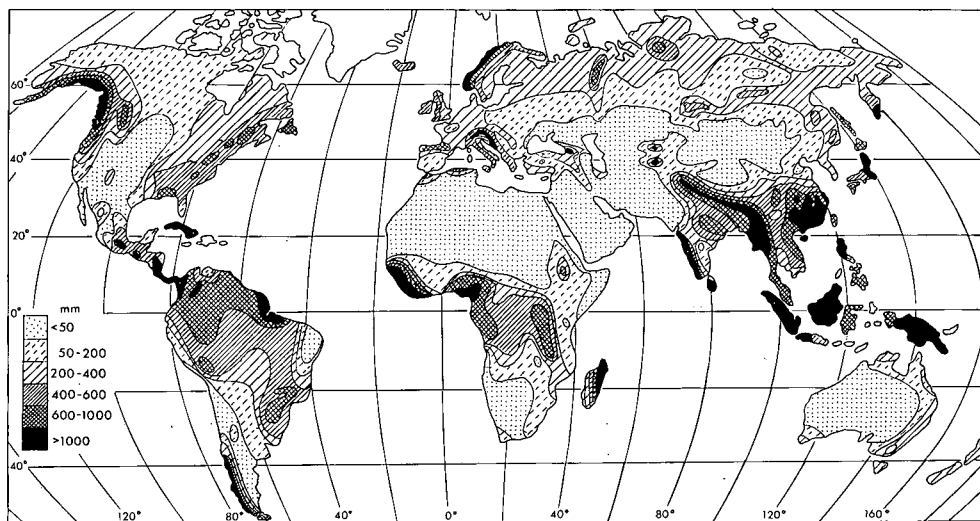


Fig. 27. Annual runoff (from Barry 1969, after L'vovich 1964). A less generalized picture is found in the Atlas of world water balance (1974) but the main general features are the same.

fluence of type II rills, and flow in natural depressions.

Equation 106 may be regarded as a variant of the sediment rating curve technique for a slope or a small basin,

involving a detailed examination of the rill network, soil, land use and cultural practices. SAY_p can be interpreted as being sediment yield or load per area during the runoff period, when discharge per

Table 20. Parameter values in equation 106 for different seasons, land use and size of rill network. The rills increase in size from type I microrills to type III large rills located in depressions (from Bobrovitskaya *et al.* 1977)

Land use	Type of rill network	V	N
<i>Spring period</i>			
Old sod field	I	$0.46 \cdot 10^{-6}$	1.6
	II	$0.38 \cdot 10^{-4}$	2.6
	III	$0.38 \cdot 10^{-3}$	1.6
Perennial grass	I	$0.28 \cdot 10^{-5}$	1.4
	II	$0.23 \cdot 10^{-3}$	1.4
	III	$0.23 \cdot 10^{-2}$	1.4
Winter crops	I	$0.42 \cdot 10^{-5}$	1.4
	II	$0.35 \cdot 10^{-3}$	1.4
	III	$0.35 \cdot 10^{-2}$	1.4
Stubble	I	$0.11 \cdot 10^{-4}$	1.3
	II	$0.93 \cdot 10^{-3}$	1.3
	III	$0.72 \cdot 10^{-2}$	1.3
<i>Summer period</i>			
Dense crops:			
old sod field, perennial and annual grass;	I, II, III	$0.3 \cdot 10^{-2}$	1.0
cereals (summer and winter crops)			
Furrow planting:			
potatoes, maize, beetroot, etc.	I, II	0.5	1.0
	III	4.9	1.0
Fallow land	I, II	1.4	1.0
	III	7.0	1.0

Table 21 Parameter values in equation 106 for different soil types and size of rill network in autumn-ploughed land (from Bobrovitskaya *et al.* 1977)

Land use	Soil types	Types of rill network	V	N
Autumn-ploughed land	Podzolic, light, podzolized forest-steppe and light chestnut soils	I	$0.30 \cdot 10^{-3}$	1.0
		II	$0.32 \cdot 10^{-1}$	1.2
		III	0.12	1.1
Autumn-ploughed land	Podzolized forest-steppe leached and dark chestnut soils	I	$0.30 \cdot 10^{-3}$	1.0
		II	$0.30 \cdot 10^{-2}$	1.6
		III	$0.40 \cdot 10^{-1}$	1.2
Autumn-ploughed land	Deep and common chernozems	I	$0.30 \cdot 10^{-3}$	1.0
		II	$0.63 \cdot 10^{-3}$	1.8
		III	$0.11 \cdot 10^{-1}$	1.3

area is higher than a value representing p % of a discharge-duration curve. The parameters V and N have been computed for snowmelt conditions (= spring period) and for rainfall floods (= summer period), as well as for different types of land use (Table 20). In the case of autumn-ploughed land, snowmelt parameter values have been determined for three types of soil with different re-

sistances to erosion (Table 21).

The study includes forest, forest steppe and steppe regions of the USSR. As runoff for a whole year varies from less than 50 to 400 mm in these regions (Fig. 27) and the exponent N, with two exceptions, varies between 1.0 and 1.6 and the V values ranges between $0.46 \cdot 10^{-6}$ and 7.0, the runoff values seem to be of less importance than the V values.

12 Deterministic models

Deterministic models are based on knowledge of the fundamental erosion processes. At present, models involving relationships developed from basic erosion theory are few and have not been thoroughly tested. However, the development of such models has been stimulated by the growth of digital computers in the last fifteen years, and perhaps improved theoretical models will be more common in the future. Since the climatic or hydrologic variables used in deterministic models are essentially the same as those used in empirical models and since deterministic models are only in their infancy, there will only be a brief review of deterministic models.

A pioneering work which has had great influence is that by Meyer & Wischmeier (1969). Their model is split into four separate, but interrelated phases which describe the process of soil erosion. The component processes are detachment by rain, detachment by runoff, transport by rain and transport by runoff. Equations describing the component processes were developed. Slopes are divided into slope segments, and total detached soil is compared to total transport capacity at the end of the increment (Fig. 28).

The basic equations proposed for the model are:

$$D_{\text{Rain}} = S_{\text{DR}} A_i I^2 \dots\dots\dots (107)$$

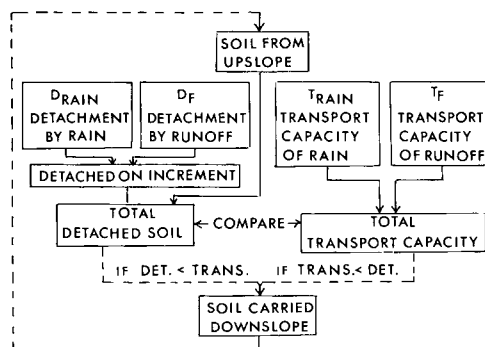


Fig. 28. Soil erosion model by Meyer & Wischmeier (1969).

$$D_F = S_{DF} A_i S^{2/3} Q^{2/3} \dots\dots\dots (108)$$

$$T_{Rain} = S_{TR} S I \dots\dots\dots (109)$$

$$T_F = S_{TF} S^{5/3} Q^{5/3} \dots\dots\dots (110)$$

where

D_{Rain} = detachment by rain

D_F = detachment by runoff

T_{Rain} = transport capacity of rain

T_F = transport capacity of runoff

A_i = area of slope segment i

I = rainfall intensity, in/h

S = slope %

Q = runoff, in/h

S_{DR} , S_{DF} , S_{TR} , S_{TF} = coefficients including soil effects

The numerical values of the exponents in the above equations were based on studies of processes as well as theoretical considerations, and were not determined through statistical calculations. A runoff submodel is required in the erosion model. In one application of the model, the soil constants were assumed to be: $S_{DR} = 0.01$, $S_{DF} = 0.0005$, $S_{TR} = 0.01$ and $S_{TF} = 0.0001$. The model has significant limitations, which the authors acknowledge. Vegetation effects, tillage and freeze-thaw effects are not intro-

duced into the model. Steady-state conditions are considered both with respect to rainfall intensity and infiltration rate, and soil detachment properties were not considered to be time-dependent.

Foster & Meyer (1972, 1975) attempted to express the notion that detachment is dependent on the sediment load in the flow in accordance with the equation:

$$D_F = c(T-G) \dots\dots\dots (111)$$

where D_F = detachment by flow

c = constant

T = transport capacity

G = sediment load

Many computer models of small drainage basins use the Meyer & Wischmeier conceptual model as a base, e.g. the model by Donigian & Crawford 1976 and the Colorado State University Model by Li *et al.* (1977), Simons *et al.* (1977), and Simons *et al.* (1979). However, there are still many unsolved problems to overcome before a comprehensive process simulating erosion model can be developed. Fleming & Walker (1977) stressed the importance of the armouring of the soil in the calculation of the available soil particles.

Without an accurate simulation of runoff, the exact modeling of non-point water erosion is impossible. Fleming (1975) provided a good outline of progress in the computer simulation of overland flow up to 1972. To my knowledge no perfect distributed model is available involving areal variation in overland flow as a consequence of saturation differences and throughflow conditions. However, in dry climates, distributed Hortonian models may be adequate as in the case of the Colorado State University model.

PART III CLIMATIC DIFFERENCES IN EROSION RATE. AN ANALYSIS OF EROSION MAPS

13 Introduction

Part III is an inventory and an analysis of erosion maps published in articles on soil erosion. The search for maps has been difficult as bibliographies and data bases seldom mention the existence of erosion maps. The inventory was arrived at by surveying the extensive literature on erosion. Nevertheless, the inventory does not include as many erosion maps as would have been desirable. Despite great efforts to make an exhaustive search, some maps have no doubt been missed.

The conception of climatic control dominates this study. When erosion maps are considered, attention is therefore paid to whether any climatically related pattern of erosion rates is discernable or not.

First, world maps of sediment yield are examined. Then regional maps covering major areas of the countries under consideration are examined and compared with the global maps. Finally other types of erosion maps, which do not show sediment yield, are also considered. These are of special interest because there are few regional maps of sediment yield and the existing maps do not cover whole continents. The third group of maps usually does not give any numerical values for the erosion rate. They only indicate a relative rate of erosion, the areas where sheet, gully and wind erosion prevail, or the erosion po-

tential, estimated on the basis of various factors. However, one map shows eroded topsoil in mm, and others indicate potential erosion in weight per unit area and time.

The text includes a brief description of the construction of each map. The reliability of a map is directly dependent upon the density and uniformity of the data points (Davis 1973). One main task has therefore been to examine the density of the basic data points to see if there is any bias in their distribution which would make the map unreliable in general or in any particular area. The reliability of a map is also dependent on the measurement techniques employed and the calculation methods used to determine the sediment yield values. These methods are therefore examined. However, the articles often give a limited amount of information on this matter. The text also includes a description of what methods are used to estimate erosion in areas which have not been measured.

Quantitative comparisons of erosion rates between maps are often difficult to make. One reason is that both global and regional maps use different intensity intervals. The maps would be much easier to compare if the same intervals were used in all the maps. Where changes have been feasible, revisions of the original maps are made, using the following basic intensity intervals: 0–10,

10–20, 20–100, 100–500, 500–1000, 1000–2000 and > 2000 tons/km²/year. The original maps have not been revised to any great extent. However, the same symbols are used throughout the study to indicate similar basic intensity intervals.

When more than one intensity class is included in one basic intensity interval, the visual impression of the basic interval is followed as closely as possible. In some instances, two or three intensity classes have been condensed.

14 Global maps based on sediment yield measurements

14.1 The maps by Fournier and Strakhov and the UNESCO map

Different methods have been used to survey land erosion by water on a global scale. The map by Fournier (1960) is a map of suspended sediment yield in basins > 2000 km² (Fig. 29). Suspended sediment yield from basins > 2000 km² was correlated with a precipitation seasonality index (cf. section 3.4.3). The correlation resulted in four equations for different types of climate and relief (equations 21–24 in section 9.2). In order to construct the map, Fournier used precipitation values from a great number of stations and calculated values of sediment yield by using the four equations. He plotted the computed figures on maps of the continents at scales of 1/12 000 000 to 1/30 000 000. Curves of equal sediment yield were easily traced (Fournier 1960). The final map was reproduced at a scale of 1/71 000 000 at the equator.

An examination of the rivers used for the deduction of the equations reveals that there are no rivers from areas having both an accentuated relief (orographical coefficient > 6) and an Af or Aw climate. Only one river, the Irrawadi, is classed as having an Aw climate and a low relief. Two river basins, the Ganges and the

Sikiang, are classed as having a Ca climate, which is nearly tropical in some parts. Thus, the whole of the tropical humid zone is mapped using two equations, which are based on practically no sediment yield data from that zone except for the data from Irrawadi. The distribution of basic sediment yield data for the equations is also seriously biased in respect to the different continents. There are no data from Africa, Australia or from South America, but there are data from 57 basins in the USA.

In agreement with what can be expected from the equations and their seasonality index (cf. Fig. 17), Fournier's map shows very high values of erosion (> 1000 tons/km²/yr) in some mountainous areas; in tropical areas which have strong seasonality and quite high annual precipitation; in humid tropical islands which have high rainfall; and in subtropical, monsoonal, high rainfall areas.

Another world map of erosion rates was constructed by Strakhov (1967), on the basis of the suspended load in 60 rivers (Fig. 30). A large number of the largest river basins in the world was included. Strakhov found that the geographic distribution of intensities of mechanical denudation was highly distinctive. He recognized two broad, parallel zones separated by the 10°C mean annual iso-

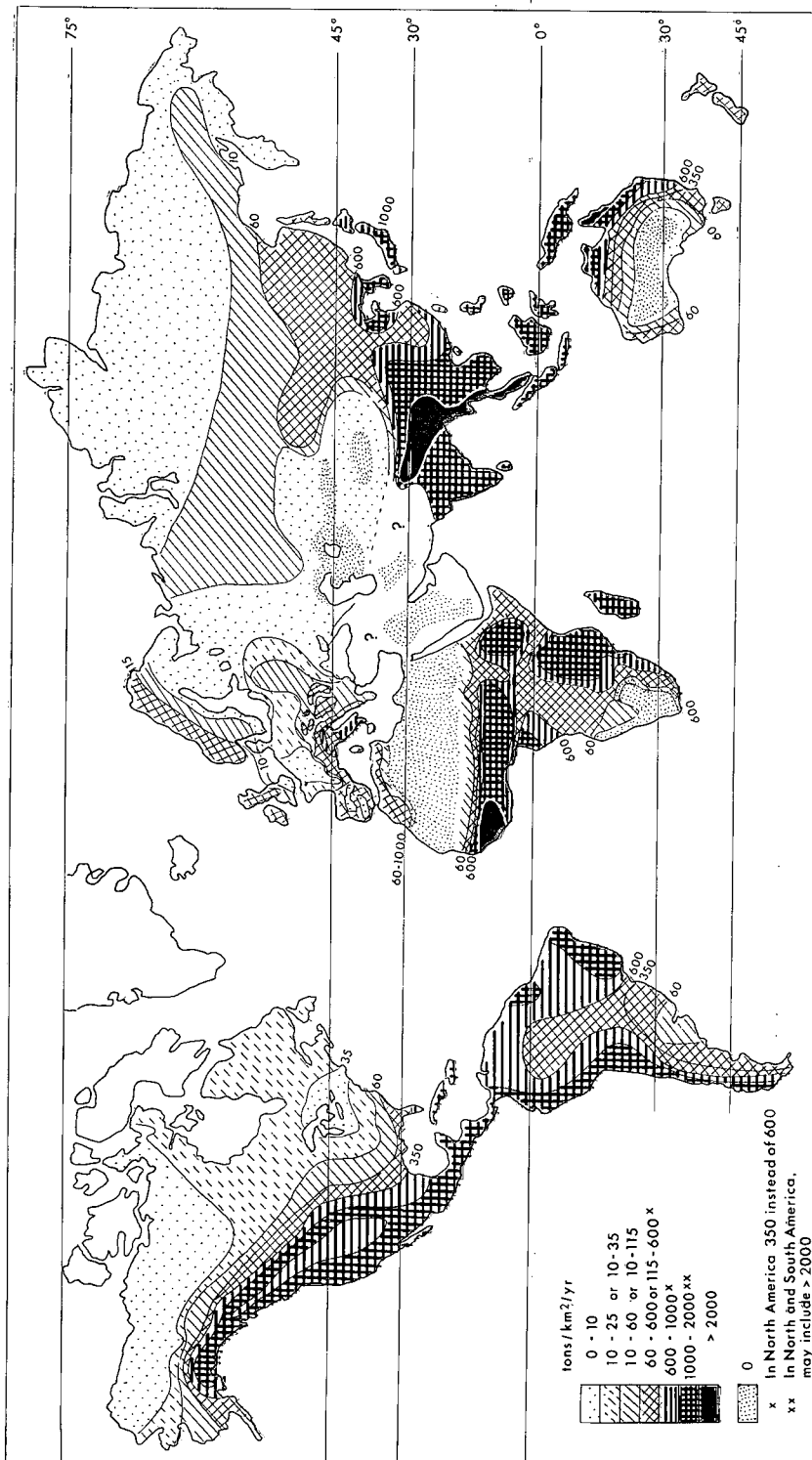


Fig. 29. Global erosion map by Fournier (1960).

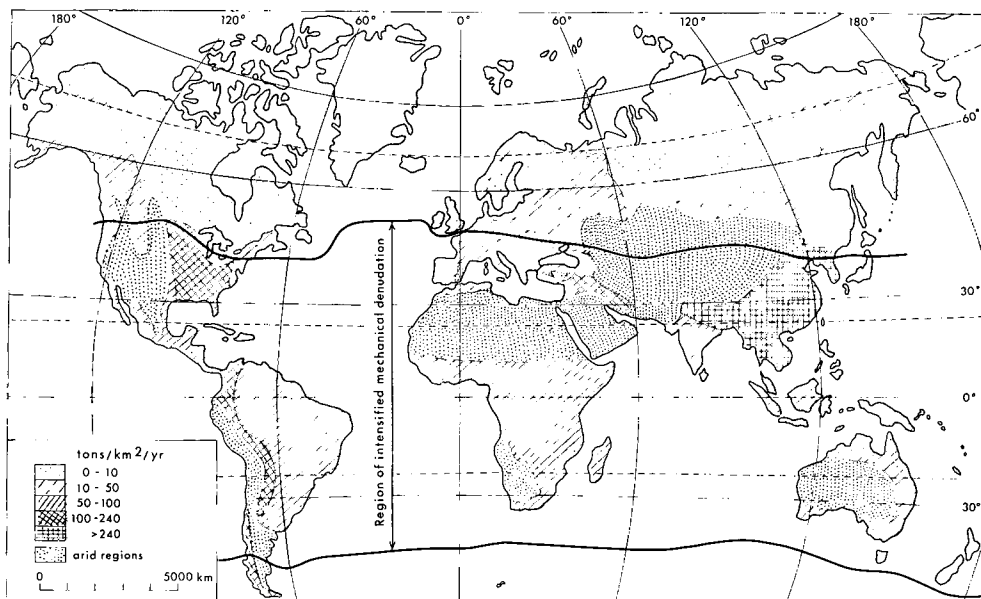


Fig. 30. Global erosion map by Strakhov (1967).

therms. One zone embraced the moist northern temperate belt with generally low erosion values and the other embraced the moist subtropical and tropical belt with high erosion rates. The 60 sediment yield values indicated the approximate values that should be assigned within the climatic belts. An explanation was also sought for exceptionally high values. In the Mississippi basin the high rate of mechanical denudation was thought to be caused by man's activity. In mountainous areas, erosion was expected to increase considerably. The occurrence of a dry period during the year and subsequent heavy rains, in combination with high relief, in China and Indo-China was thought to contribute to the very high erosion rates in these areas.

UNESCO & AIHS issued a preliminary global map in 1974 (not reproduced in this study). The object of the map was to elucidate the net suspended sediment transport to the oceans per unit area. The measured values of suspended sediment

yield to the oceans, divided by the area of the basins are given at the river mouths. The existence of reservoirs and lakes are not taken into account in the calculations.

Large areas on the UNESCO map have no values at all, e.g. Australia, Indonesia, Scandinavia, Mexico and Central America. In other areas, there are very few values. No attempts have been made to estimate the erosion within these areas. Some portions of this map are illustrated and discussed in comparison with other maps (chapters 15–17).

14.2 Evaluation of the global maps

The UNESCO map is incomplete and indicates sediment yield measured at the basin outlet to the oceans (tons/km²/yr). The sediment trapped in reservoirs and lakes is not included in the calculations. On the other hand, the areas above the reservoirs and lakes are included when sediment yield is divided by the basin

area. This makes the erosion rate values misleading in many areas.

The global maps published by Fournier and Strakhov are the results of generalizations based on the assumption of a simple relationship between erosion and climate-relief. Fournier plotted his map with the help of regression equations developed for areas with different climates and relief. Strakhov plotted his map by grouping his 60 basins with basic sediment yield data into groups with similar climatic and relief conditions and then extending similar values to similar climates and relief. In both cases, the distribution of the basic data is of importance to the outcome. Since only part of Strakhov's data is tabulated, the distribution of his data cannot be studied in any detail. However, Strakhov's data cover large areas of the world, as very large basins are included. The equations by Fournier are to a great extent based

on data from agricultural areas, a condition which involves the inclusion of high values. This is especially true within C climates. In areas with more natural vegetation the values might therefore be too high. In addition, Fournier's map must be considered as hypothetical in tropical regions due to the serious lack of basic data.

Since the global maps by Fournier and Strakhov occur in many textbooks, it would be of great interest to know their reliability in general. It would also be interesting to know where they give good or bad approximations of the real sediment yield. As regional maps are normally based on a denser data net within the particular region than global maps, regional maps must be considered to give a more reliable picture of sediment yield conditions. One way to check the global maps is therefore to compare them with regional maps.

15 Regional maps based on sediment yield measurements

15.1 Presentation of regional maps

15.1.1 Introductory remarks

There are comparatively few regional sediment yield maps. This scantiness may arise from a shortage of sediment yield data in combination with the fact that erosion varies to a great extent even within small areas. Some regional maps cover only a rather limited area, while others have the character of national maps.

The following presentation includes a description of the construction methods and an evaluation of the maps as well as a discussion of the reasons for the varia-

tions in erosion rates within the areas investigated. Comparisons are also made with other regional maps or with global maps. The presentation is made country by country and is arranged according to climatic and geographic location. This simplifies recurring comparisons of erosion rates in similar climates.

15.1.2 Yugoslavia

The erosion map of Yugoslavia in Fig. 31 is based on measurements performed during the period 1952–56 at 16 gauging stations on 12 rivers (Jovanovic & Vukcevic 1957). Scattered measurements were also performed on a few other riv-

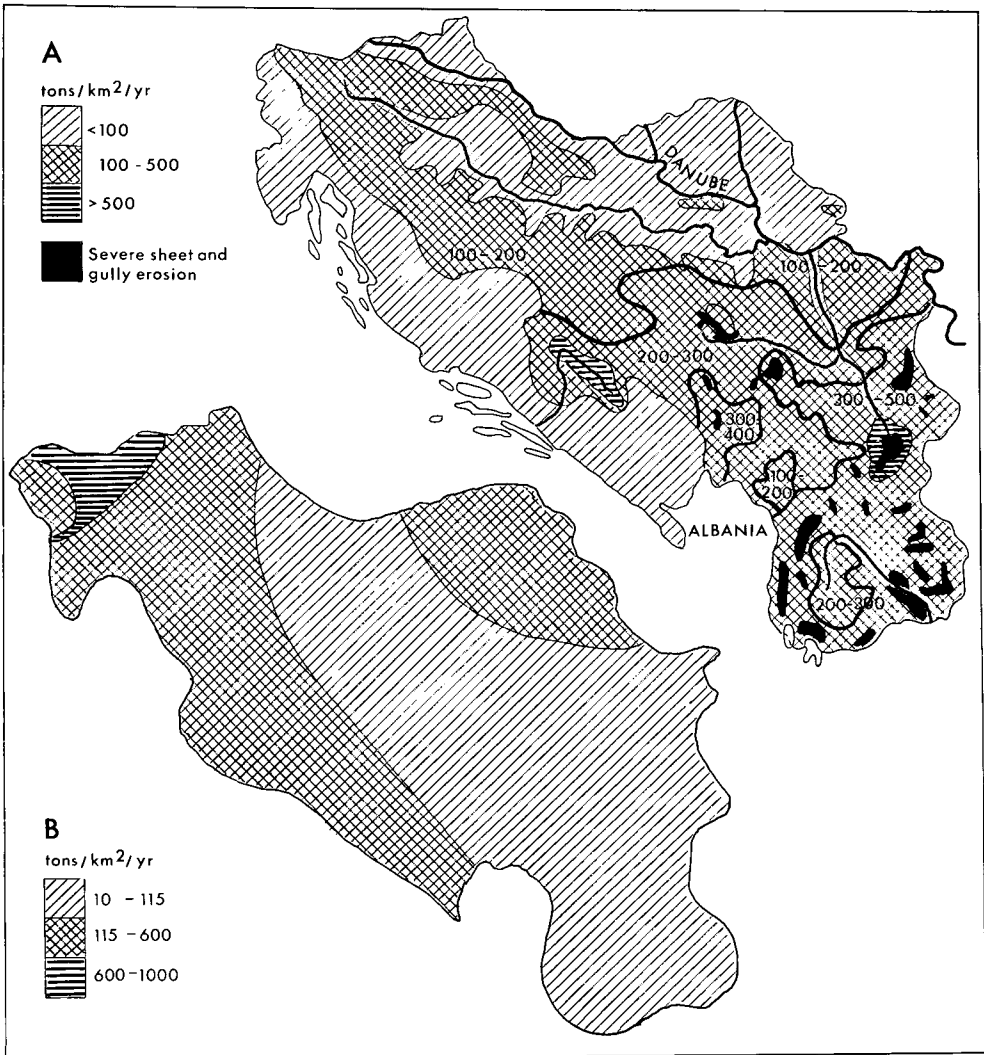


Fig. 31. Yugoslavia.

A: Map of sediment yield. 0.6 data sets/10,000 km² (from Jovanovic & Vukcevic 1957).

B: Detail from global map by Fournier (1960).

ers. The suspended load observations continued for one or two years. Sediment samples were taken on a daily basis. In periods of abrupt water level changes, sampling was undertaken several times a day, and in some places observations were made every one or two hours.

Sediment rating curves were con-

structed on a monthly or seasonal basis. Monthly mean values of suspended sediment yield and the water discharge of each river were correlated by the application of the equation

$$(SAM) = aQ^n \dots\dots\dots (112)$$

where (SAM) = mean monthly sediment discharge in kg/s, and Q = mean monthly

discharge in m^3/s . The values of the exponent n ranged between 1.7 and 3.0 for the different rivers. The mean value of the exponent was $n = 2.25$. By using this n value in each river basin the corresponding value of the erosion coefficient a was calculated. Its value varies with the characteristics of individual catchments.

Annual mean values of water discharge and sediment load were then plotted in a coordinate system. Points were analysed according to characteristics such as amount and distribution of rainfall, soil, geologic conditions, topography and land use, and the value of the erosion coefficient a was determined for each group. Average annual sediment yield was then calculated by utilizing long-term observations of water discharge. The quantity of bed load was estimated as a percentage of the total amount of suspended sediment load. This percentage was taken to be 10 to 20 %. In areas without available data, soil erosion was estimated by analogy with areas of similar characteristics.

In the southern half of Yugoslavia, the forests have been largely destroyed. In the northern half of the country, reforestation is more common, and as vegetation cover was found to be a very important factor in the retardation of erosion, this helps to explain the lower erosion in these areas. The authors also imply that the permeable limestone bed rock contributed to the low sediment yield in the north. In the northeastern part, there are flat areas with low sediment yield, and along the Adriatic coast, karst regions contribute to the low sediment yield.

Strakhov's map shows one symbol of 50–100 tons/km^2 for the whole country (Fig. 30), whereas the regional map has more variable values including a great area with much higher values than 100 tons/km^2 . Probably the mean value for the whole country is greater than 100

tons/km^2 . Fournier's map shows 10–115 tons/km^2 in southeastern Yugoslavia which has a continental climate (Fig. 31), whereas the regional map indicates values from 100 to $> 600 \text{ tons}/\text{km}^2$. Along the coast with its Mediterranean climate, the regional map indicates up to 100 tons/km^2 , whereas Fournier has the higher marking of 115–600 tons/km^2 . Thus the trends are reversed on Fournier's map. The reason for this discrepancy may be the dominant influence of geologic conditions on the erosion of solid material in karstic areas. Fournier did not use any sediment yield data from Yugoslavia, and his method disregards geologic conditions. On the other hand, the regional map includes sediment yield data from 16 stations in this particular area and also takes into account more factors in the estimation of erosion rates in areas without sediment load data. The regional map must therefore be regarded as more reliable.

15.1.3 Rumania

A similar approach was adopted by Diaconu (1969, 1971) in order to construct a map of mean annual suspended sediment yield in Rumania (Fig. 32A). The map is based on data from 202 stations, collected during the period 1952–67. Seventy-six of the 202 stations have sediment discharge data covering 12 years or more; 43 have data for 8–11 years; 61 have data for 4–7 years; and 22 stations for a period shorter than 3–4 years.

Two types of measuring techniques were adopted: *complete measurements*, which means that the concentration was determined at all points in the profile where water velocity was measured, and *simple measurements*, when the concentration was determined at two or three points. The number of complete measurements varied between 10 and 18 per year and profile. Simple measurements

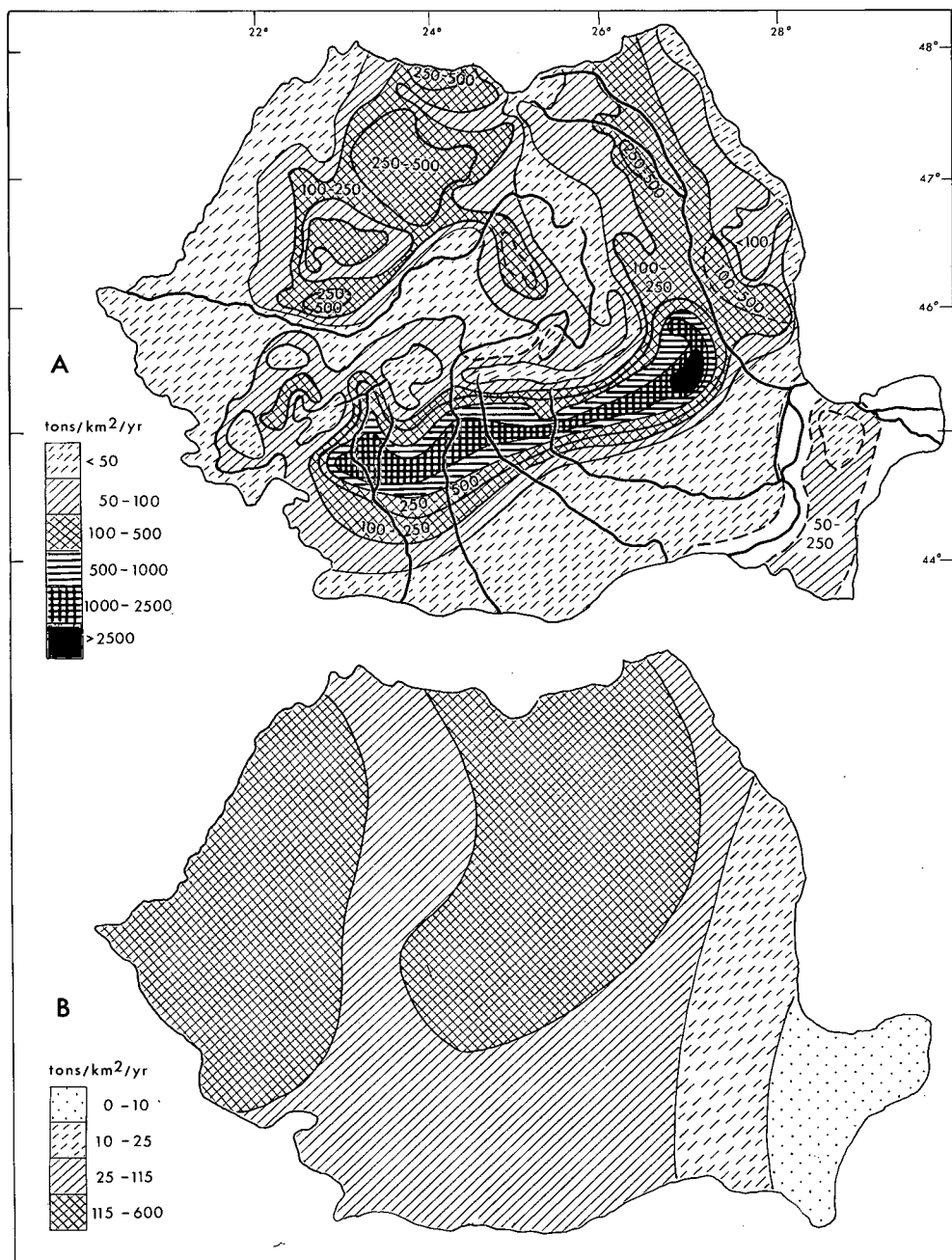


Fig. 32. Rumania. A: Suspended sediment yield. 8.5 data sets/10,000 km² (from Diaconu 1971). B: Detail from global map by Fournier (1960).

were obtained once a week in 1951 and were gradually increased to daily measurements from 1964 on.

Within 12 homogenous physical-geographical zones, relations between mean annual suspended sediment load SAY in kg/s and mean annual water discharge Q in m^3/s for small rivers were established with the following general formula:

$$\log SAY = b \log Q - \beta \dots\dots\dots (113)$$

The value of b varies very little and is close to 1.25 over the whole country. The value of β , however, varies within wide limits. Within each zone, there is a relation between β and the mean elevation of the basin enabling sediment yield values to be calculated for mean elevation intervals within each zone. The relationships between sediment yield and mean elevation of the basins were then used in the construction of the map. The method of utilizing a constant value of b is comparable to multivariate equations including mean annual water discharge (cf. the Appendix).

The map shows very high sediment yields > 1000 tons/ km^2 in the Transsylvanian Alps with a maximum value > 2500 tons/ km^2 . The Carpathians have values of 100–250 tons/ km^2 , and the lowlands < 50 tons/ km^2 .

In Strakhov's map Rumania falls entirely within the 50–100 tons/ km^2 intensity class. This gives values which are too low, as the mean value of the country according to Diaconu (1969) is 188 tons/ km^2 . On the other hand large areas of the country exhibit values which are less than 100 tons/ km^2 . Fournier's map shows values of 10–115 tons/ km^2 in large parts of Rumania and 115–600 tons/ km^2 in the Carpathians (Fig. 32B). In the western part of the country, the discrepancy between Fournier's map and the regional map is great. With the exception of the western areas, the order of magnitude of sediment yield seems acceptable, considering the degree of generalization inherent in the global maps.

14.1.4 Italy

The erosion map of Italy was drawn with the help of data from 70 gauging stations (Gazzolo & Bassi 1960). A map of erodibility was constructed, based on the assumption that rocks with a high density weather and form soils which are very resistant to erosion, and that rocks with low density break down to highly erodible soils. Three categories of erodibility were mapped. A graph with the percentage of the basins with a low erodibility as the abscissa and sediment yield in tons/ km^2 and m^3/km^2 as the ordinata was drawn. Values from the 70 river basins were plotted, and a median line was drawn. With the erodibility map as a base, the median values of the graph were used to construct the erosion map (Fig. 33). The measured values were also used. The map values have been converted from mm to tons/ km^2 by assuming a density of 2.5 tons/ m^3 , a density which was used in the calculations according to the article.

The areas with more than 2500 tons/ km^2 are situated in extremely erodible land with an almost complete lack of wood and grasscover, or areas which are devoted to intense agriculture. Most of Italy has values which are less than 500 tons/ km^2 .

On the UNESCO (1974) map there is an inset map of Italy. The boundaries of the basins and the numerical values were transferred from the UNESCO map to Fig. 33. Except for a single basin in Sicily, these values are in agreement with those indicated in the regional map. Strakhov's map indicates a range of 50–100 tons/ km^2 , which seems too low even for a generalized map. Fournier indicates 600–1000 tons/ km^2 for most of Italy but 60–600 tons/ km^2 in the Po lowland, along the Toscana coast down to Rome, and in Sardinia.

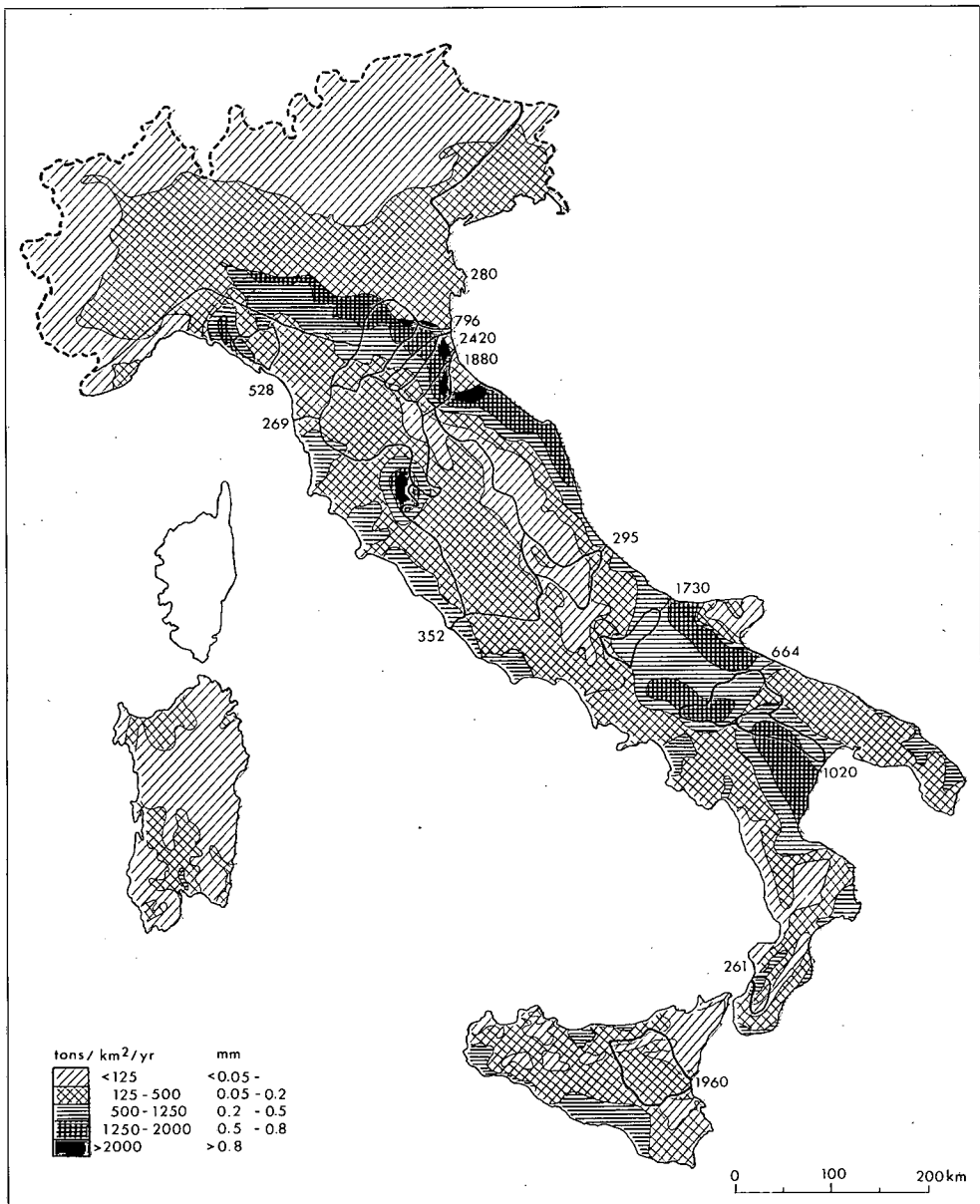


Fig. 33. Erosion map of Italy. 2.3 data sets/10,000 km² (from Gazzolo & Bassi 1960). The boundaries of a number of catchments and their sediment yield values are transferred from the UNESCO map (1974).

15.1.5 Cyprus

Another map in the Mediterranean area is the erosion map of Cyprus (Fig. 34). This map is also based on the assumption that erosion rates are related to rock

types (Phanartzis 1971). Sedimentation surveys of three reservoirs as well as results from a sediment sampling program from one river during 1968/69 and a reservoir survey undertaken by another author, are used as the basis for deducing

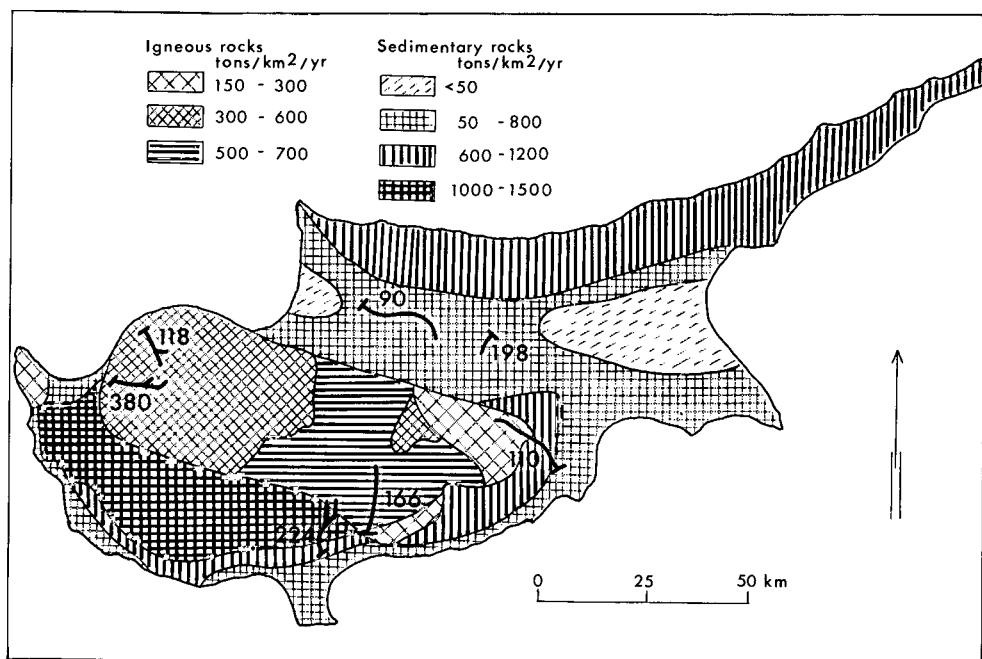


Fig. 34. Erosion map of Cyprus. 4,4 data sets/10,000 km² (from Phanartzis 1971). The individual sediment yield figures are taken from Konteatis (1974).

erosion rates for the whole island. Estimates were made by considering runoff and sediment concentrations from areas underlain by sedimentary and igneous rocks, respectively. Similarities of other areas to the basins surveyed with respect to rock types, rainfall and runoff characteristics, slope steepness, vegetative cover and cultivation and in some cases mere assumptions, constituted the basis for the division of Cyprus into erosion rate zones.

The values on the original map are given in mm/yr of semi-consolidated sediment. It is hazardous to convert the figures to an erosion rate in tons/km², since there is no information about the density. In order to be able to compare values from this map with erosion rates elsewhere I have assumed a dry weight of 1.0 tons/m³ so as to obtain values in tons/km². The figures plotted on the map have been taken from Konteatis (1974)

and are converted to tons/km² by making the same assumption (Fig. 34). The reservoirs in the latter survey had been in operation from 5 to 11 years. Some of the values are lower than the map values. However, the map is based on rather subjective assumptions.

The map of Cyprus shows that the highest erosion rates are associated with the flysch rock types and vine cultivation, steep slopes and a high rainfall. Landuse and slope may be responsible for the high rate of erosion, but these factors are related to rock type. The igneous rocks on the island are associated with mountainous terrain, high precipitation and slight agricultural land use. The main cause of the low erosion values, in spite of a quite high runoff in these areas may be the natural vegetation.

Cyprus was not included in either Strakhov's or Fournier's map.

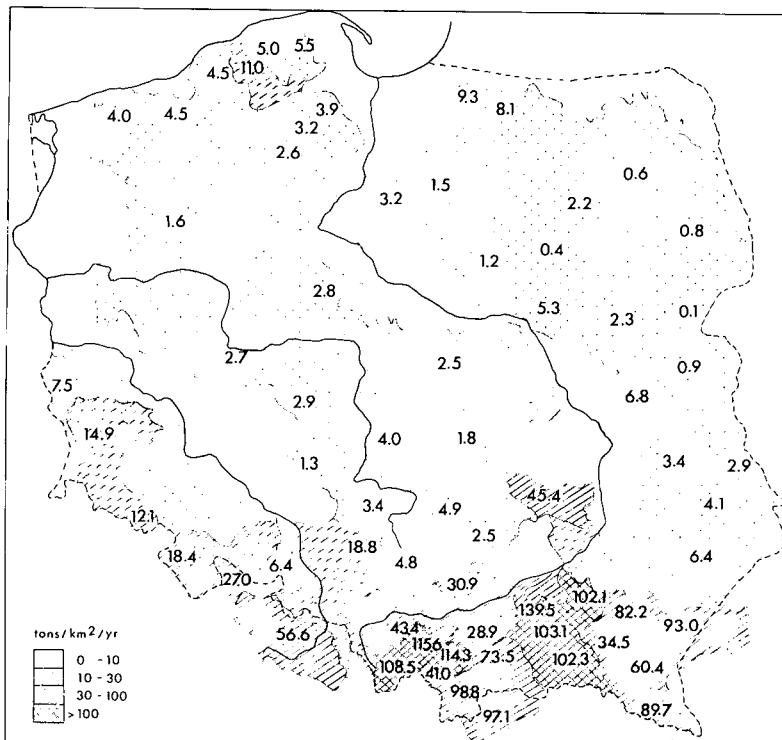


Fig. 35. Suspended sediment yield in Poland. 3.0 sets/10,000 km² (from Brański 1975).

15.1.6 Poland

The map of suspended sediment yield in Poland, by Brański (1975), is based on data from a dense network of sediment gauging stations and reservoir surveys (Fig. 35). Estimations of trap efficiency and the amount of suspended sediment accumulated in reservoirs were made by employing empirical equations and graphical relationships (cf. Brański 1980). The suspended load values plotted in Fig. 35 are averages for the period 1956–1970. In unmonitored years a graphical relationship of sediment yield against mean annual water discharge at each gauging station was used. No observations were made in several areas directly adjoining the Vistula and the Oder.

The Carpathian region and a hilly area with loess soils in southern Poland have

high sediment yields. In other areas in Poland sediment yields are low even in loess regions.

15.1.7 The USSR

An erosion map covering the Soviet Union is published in Atlas Mira (1964) and presented in this study as Fig. 36. This map is based on a map of sediment concentrations by Lopatin and a number of regional, suspended sediment concentration maps, as well as data from about 1200 gauging stations. The magnitude of the sediment yield per unit area applies only to medium-sized and large rivers. Even though the network of gauging stations is not very dense, the map is based on several additional regional maps and in combination this means that the accuracy of the map is probably quite

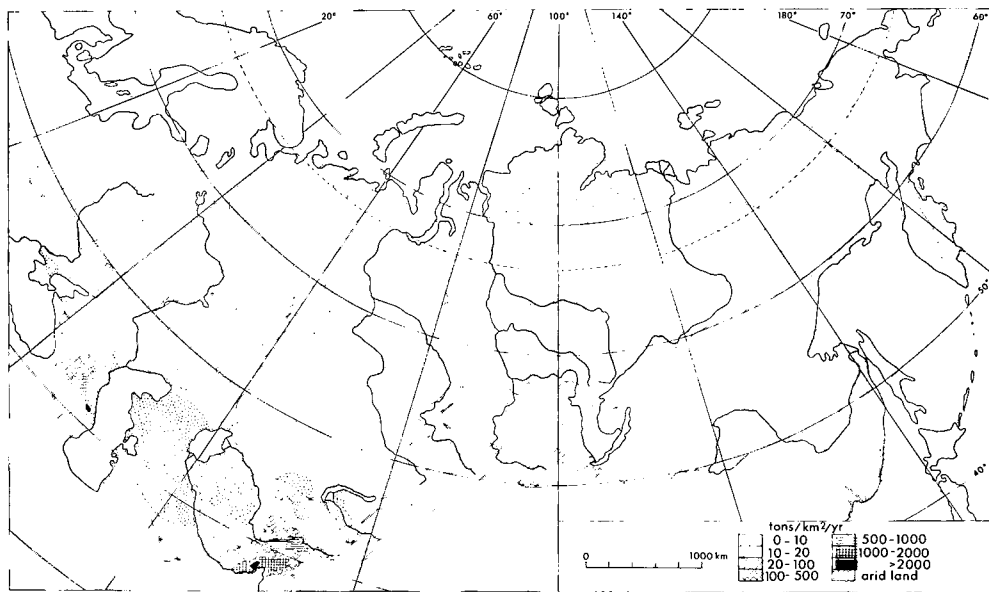


Fig. 36. The Soviet Union. Sediment yield of middle-sized and large rivers. 0.5 data sets/10,000 km² (from Atlas Mira 1964).

good. However, no information is available about the distribution of the gauging stations or the accuracy of the measuring techniques used.

There are large areas with very low annual sediment yields of less than 5 tons/km². The mountainous areas north of Pamir and the Caucasus have the highest rates of annual erosion in the USSR with values up to 2500 tons/km². There are even small areas with > 2500 tons/km² (Fig. 37).

The pattern of erosion rates in Atlas Mira is similar to that observed on Strakhov's map although the intensity intervals do not coincide. Some areas which are indicated 50–100 tons/km² in Atlas Mira are not illustrated in Strakhov's more generalized map. One large area in Strakhov's map, is mapped as an "arid region". Within this "arid region" are situated the mountains to the north of Pamir, which exhibit very high erosion rates in Atlas Mira. The pattern on Fournier's map of the Soviet Union does

not seem to fit the pattern in Atlas Mira.

Tkacheva (1974) divided Soviet Central Asia into erosion regions (Fig. 38). The limits of the different regions were outlined by considering the erosion controlling factors of climate, river flow, lithology, type of soil and elevation. Some of these regions coincide with identified physiographic regions in Soviet Central Asia. The sediment yield values are based on observations made in zonal rivers with drainage areas of 100–100 000 km². Data with records of less than 5 years were not used.

It is difficult to compare other maps with this type of map as measured sediment yields are not the basis of the division. The range of sediment yield values within the regions is considerable. One contributing reason for this is probably the great range of basin sizes. The highest erosion rates of 1000–2500 and > 2500 tons/km² in Atlas Mira occur in erosion region IV of Fig. 38 with 2400 tons/km² as an average value and

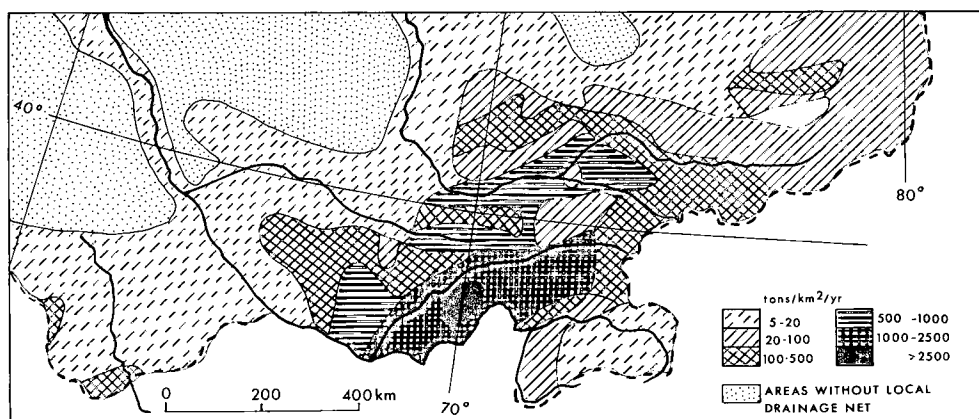


Fig. 37. Detail map of Soviet Central Asia (from Atlas Mira 1964).

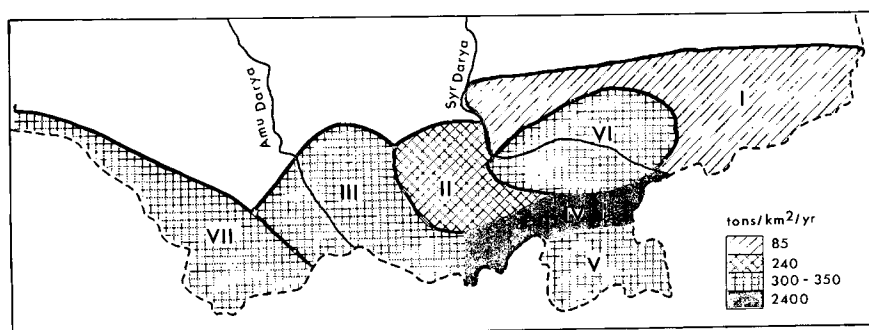


Fig. 38. Erosion regions of Soviet Central Asia. Extreme values in tons/km²/yr for the region are: I;5.9-290, II;13-880, III;43-940, IV;270-4500, V;11-1300, VI;130-780, VII;180-580 (from Tkacheva 1974).

extremes of 270 and 4500 tons/km². The reasons for the high erosion rates in region IV may be a combination of climatic, lithologic and topographic factors. A comparison with a map of the distribution of glaciers shows that the central part of region IV has a dense net of glaciers (US Army 1958). The western part of region IV consists of mountainous but cultivated areas. Two mountain ranges in region IV have high precipitation rates. The bedrock is of Cretaceous and Tertiary age and very susceptible to weathering. The low value of 270 tons/km² was probably derived from the eastern part of region IV, where Atlas Mira shows 100-500 tons/km² (Fig. 37).

As can be seen in Fig. 38, the average annual yield of four of the regions is 300-350 tons/km². Within these regions, there are, however, rivers with very low and with very high mean values. In the Pamir region, erosion is limited because of the dry climate, especially in the eastern high mountain plateau where the annual suspended sediment yield varies from 11 to 96 tons/km². Erosion is somewhat more intense in the western, strongly dissected Pamir region, where the annual sediment yield attains a value of 1300 tons/km² in one of the river basins. There are some large glaciers in this western part of the Pamir region V. In the eastern part of region I there are sev-

eral small glaciers in the Tien-Shan mountains, but sediment yields seem to be moderate.

In the text, the sediment yield of the Fergana Valley, region VI, is said to average 460 tons/km², but the map indicates a value of 330 tons/km². Loess and loess-like soils are common here. Considerable agricultural development of this region has led to widespread, accelerated erosion. Mudflows are typical of many rivers in the Fergana Valley (Tkacheva 1974). In addition, Atlas Mira indicates high values in large parts of the considered area (Fig. 37). Therefore, the higher value of 460 tons/km² seems likely.

In regions I, II and VI, the sediment yield decreases with increasing elevation, which is attributed to the increase in the proportion of non-erodible rock (Tkacheva 1974). The mountain ridges consist mainly of metamorphic and igneous rocks, and the piedmont plains and intermountain basins consist of loess-like rocks or other, equally unconsolidated deposits. More agricultural development in the basins might also be a possible explanation for the higher erosion at lower levels.

A similar type of map has also been constructed for the European USSR, excluding the Caucasus, by Lisitsyna & Alexandrova (1972). Nine physiographic regions with comparatively homogeneous sediment production conditions were delimited (Fig. 39). Climatic, geomorphological and land use factors were used as a basis for the division. Zonal rivers, i.e. rivers with drainage areas of 100 to 50 000 km², were used to characterize sediment yield within each region. A graphical relationship between sediment load and water discharge was used to determine the long-term sediment yield.

The lowest values of sediment yield are found in the forest and tundra zones. In the Carpathian mountain region VIII,

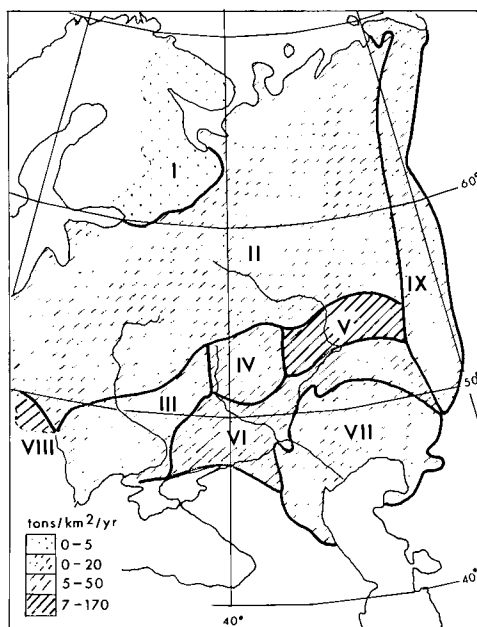


Fig. 39. Erosion regions of European USSR. Region VIII indicates 60–100 tons/km²/yr (from Lisitsyna & Alexandrova 1972).

erosion is said to be severe in large deforested areas where mudflows occur. In the steppe zone of regions III–VI, erosion is higher than in the northern forest zone. An average of 65 % of the area is farmland. Loessial material constitutes much of the soil. Gully dissection and landslides are common. Landslides commonly develop at the contact between loam and clay. In all regions of the steppe zone there is a direct relationship between sediment yield and river gradient. In loess areas, small basins usually have 10-fold higher values than the large zonal rivers.

Erosion is most pronounced in regions VIII and V. Mean annual sediment yield values of region VIII range between 60 and 100 tons/km², compared with 50–100 tons/km² on Strakhov's map and 20–100 and 100–500 tons/km² in Atlas Mira. The mean sediment yield of region V ranges between wide limits (7–170 tons/km²). Strakhov indicates 10–50 tons/km² in

region V. In Atlas Mira 20–100 tons/km² covers great parts of this region, but values of 5–20 tons/km² are also indicated. Region VII and most of region VI are mapped as arid land on Strakhov's map. This seems too great a generalization, even on the scale of a global map.

15.1.8 Canada

It is interesting to compare erosion rates in the USSR and Canada by using the USSR map in Atlas Mira and a map from Canada by Stichling (1973).

The erosion map of Canada is based on suspended sediment measurements (Fig. 40). Point integrating, depth integrating and equaltransit-rate measuring techniques were most frequently used to obtain samples. In many cases, the winter sampling programs were substantially reduced. Many rivers of small and average sizes have a winter sediment yield which is less than 5–10 % of the annual sediment yield, and some large streams have winter sediment yields which are in the range of 10 to 20 % of the annual load.

Twenty of the sediment gauging stations had 10 years of records, 40 stations had fewer than 10 years but more than 4 years of records and 60 stations had fewer than 4 years of records. A standard 10 year period, 1961–1970, was selected. Estimations of sediment yield in missing years were made. In some river basins sediment yield was estimated on the basis of hydrometric records and correlations with similar catchments where both hydrometric data and sediment observations were available, or by periodic sampling. Mean annual suspended sediment yield per unit area for the period 1961–1970 was computed. Parts of catchments were considered, if large reservoirs had been in operation in the last 10 years. In this study it was assumed that the procedure of constructing the

map was similar to that of the concentration map and that yields were plotted at the centre of gravity of the respective drainage basin. Subsequently, the boundaries of areas with similar sediment yields were determined.

The highest sediment yield, > 350 tons/km²/yr, is found in glaciated mountain regions. On the whole, the southern part of the Cordillera, which has high precipitation, high runoff and steep slopes, also has a high sediment yield.

Another regional map of Canada was prepared by Slaymaker & McPherson (1973). This map is based on published data from basins larger than 25 000 km² (Fig. 41). With the exception of the Yukon territory large areas on this map have lower sediment yield values than on the map by Stichling referring to smaller basins (Fig. 40). The differences between the maps may at least partially be only apparent and may be due to the overlapping scale intervals.

Heterogeneous sediment yield intervals make comparisons with the global maps difficult (Fig. 42). On Stichling's map, the Cordillera of Canada is mapped as 18–88, 88–350 and even > 350 tons/km². On Strakhov's map, most of it is mapped as 50–100 tons/km² and a small part of it as arid land. At that scale this seems to be quite a reasonable value. However, it would have been better if part of the Cordillera was shown as having a sediment yield of 100–240 tons/km² (cf. also Fig. 30).

Fournier's map shows very high values. Part of the Cordillera is mapped as having an annual sediment yield of 1000–2000 tons/km², and part of it, 60–1000 tons/km². These figures seem much too high. Stichling's map has the range of 18–88 tons/km²/yr marked in some areas around the Great Lakes, whereas two world maps both indicate lower values of < 10 tons/km². Strakhov has indicated 10–50 tons/km² east of the St Lawrence River. On Stichling's map,

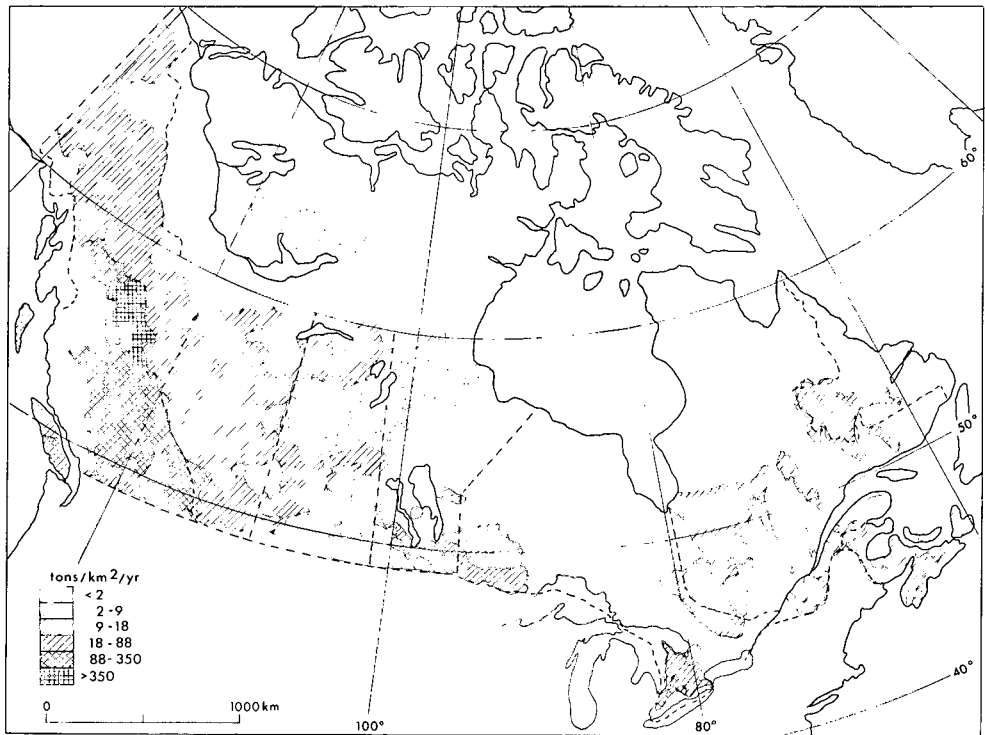


Fig. 40. Suspended sediment yield in Canada. About 0.04 data sets/10,000 km² (from Stichling 1973).

areas with 9–18 and 18–88 tons/km² are indicated east of the St Lawrence River and areas with these values are also shown in Quebec.

A comparison of erosion rates between Canada and the Soviet Union shows quite large areas with 7–27 (Fig. 41) or 18–88 tons/km² (Fig. 40) and 0–10 or 10–20 tons/km² (Fig. 36), respectively. The eastern, mountainous part of the Soviet Union, does not have as high erosion rates as the Western Cordillera of Canada. The difference may be due to a more maritime climate with more precipitation and a shorter frozen period in the Canadian Cordillera. However, there are some areas in the industrial regions of the Sayan mountains and the Carpathians which are marked as 100–500 tons/km².

15.1.9 The USA

Judson & Ritter (1964) divided the United States into a number of regions (Fig. 43). The value ascribed to each region is the mean annual suspended load plus bed load. Bed load was assumed to constitute 10 % of the total sediment load. Basic data from 28 rivers were taken from US Geological Survey, Water Supply Papers and from unpublished reports. These river basins represent approximately 60 % of the area of the United States, exclusive of Alaska. With the exception of the Columbia River, the sediment records cover at least three consecutive years. In the Mississippi drainage basin, the single station at Baton Rouge forms the basis for the estimate of rates of erosion in the entire basin. The value for the

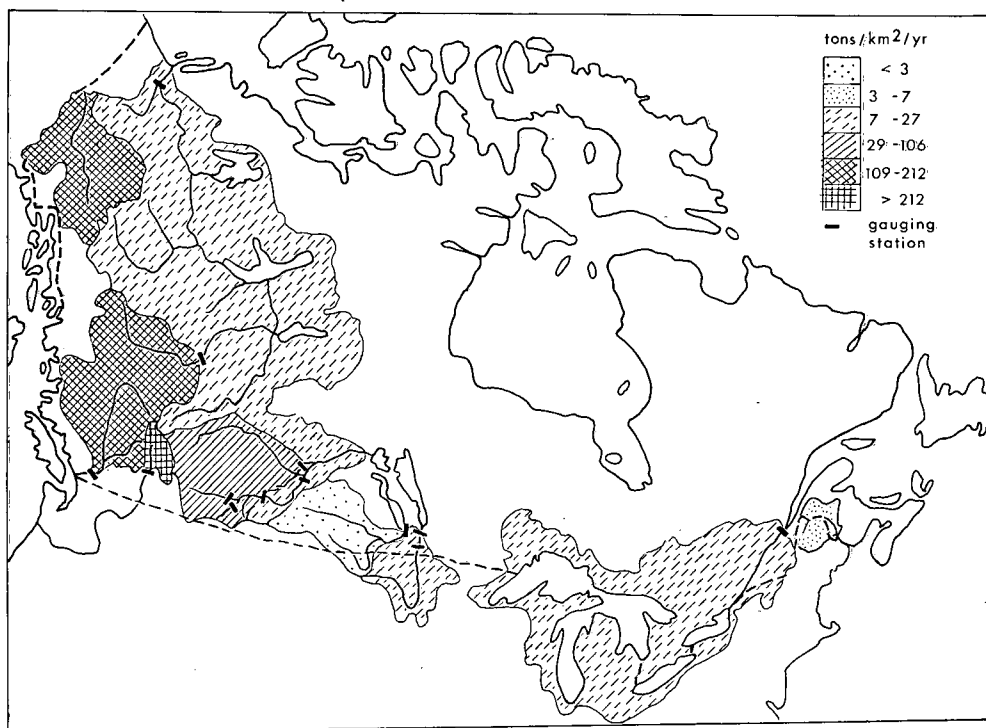


Fig. 41. Sediment yield from large rivers in Canada. About 0.04 data sets/10,000 km² (from Slaymaker & Mc Pherson 1973).

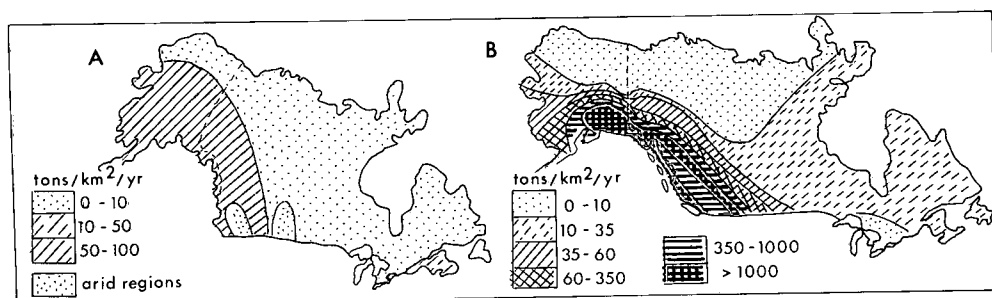


Fig. 42. Canada in global maps A: by Strakhov (1967), B: by Fournier (1960).

Colorado region is based on a 32-year continuous record at the Grand Canyon.

The UNESCO map of sediment transport into the oceans shows high values along the Pacific seaboard and small values along the Atlantic coast (Fig. 44). In the Atlantic states, values are lower than those estimated by Judson & Ritter. One reason for the low values is that there are

many reservoirs in the Appalachians, and sedimentation in those basins has not been included in the UNESCO calculations. The low value of 0.015 tons/km²/yr in the Colorado river basin is due to the same reason.

In Fig. 45, the United States has been divided into regions which exhibit broadly similar geological, meteorologi-

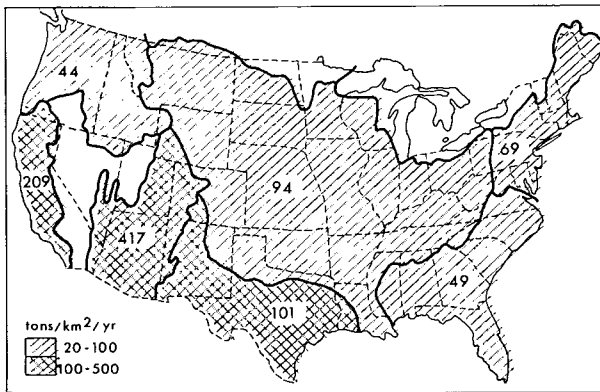


Fig. 43. Erosion regions of the USA. 0.03 data sets/10,000 km² (from Judson & Ritter 1964).

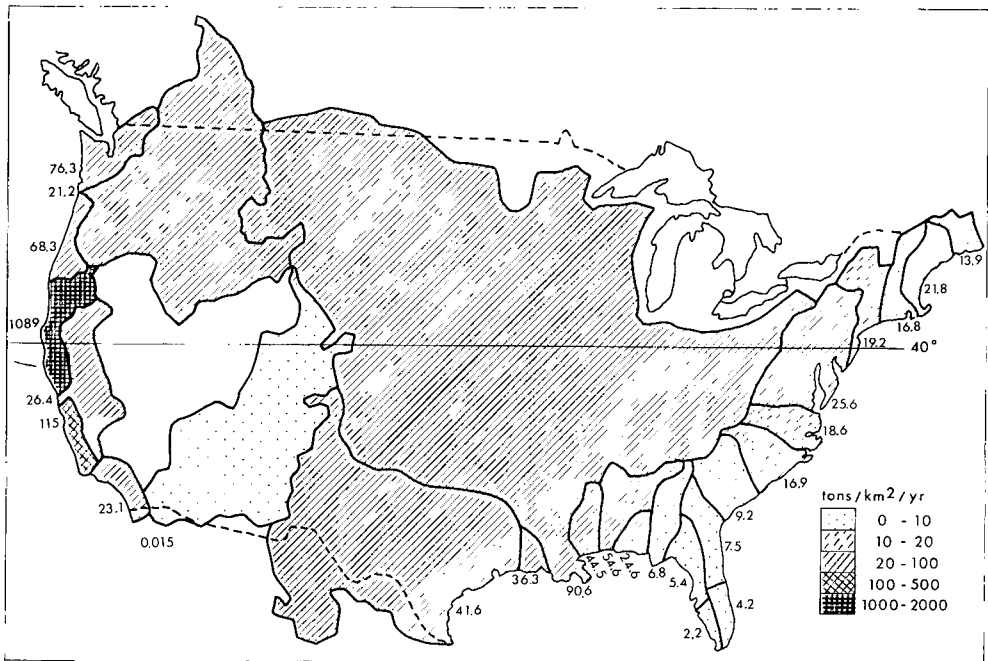


Fig. 44. USA. Sediment transport into the oceans. Detail from the global map. The values attributed to the delineated catchments are in tons/km²/yr (from UNESCO 1974).

cal, geomorphological and botanical characteristics (Geiger 1957). The sediment yield values are based on reservoir surveys in drainage basins ≤ 160 km². This erosion map is in agreement with the pattern shown in the erosivity map (Fig. 21). Regions III, IV and VII have rainfall erosivity values (EI_{30}) greater

than 500 in metric units. The average sediment yield indicated for regions III and IV may be much too high, because it is based on a small number of reservoirs and because problem areas within a region are usually the first to be measured. The loess soil of region V does not exhibit erosion values which are as high as in

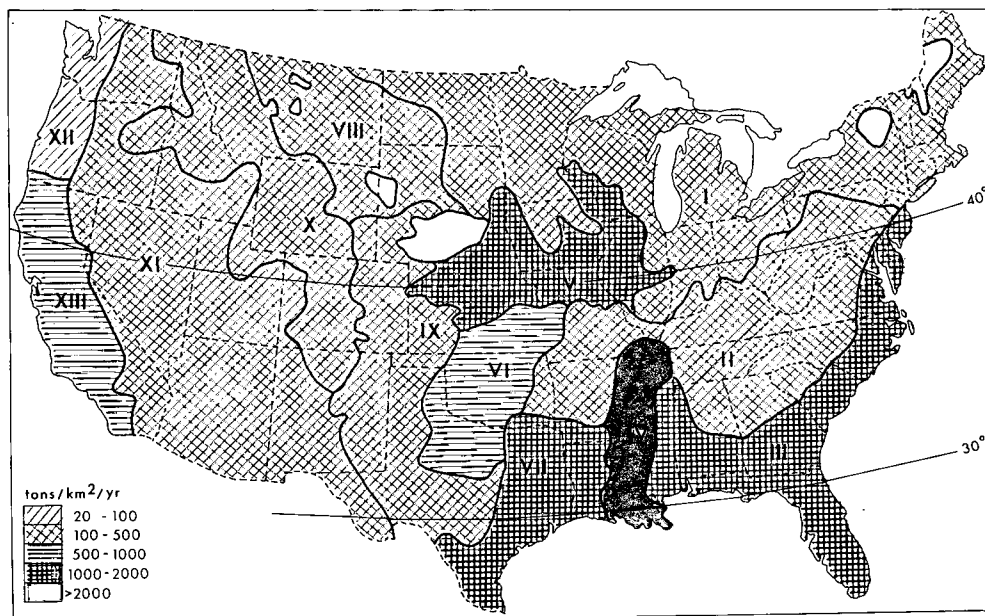


Fig. 45. Sediment yield in catchments less than 160 km² for regions in the USA. 0.4 data sets/10,000 km². The number of catchments used and the average sediment yield converted to nearest tens of tons/km²/yr for the regions are: I 36;210, II 80;350, III 9;1930, IV 14;5940, V 79;1060, VI 38;770, VII 29;1110, VIII 21;200, IX 21;220, X 5;130, XI 31;260, XII 6;60, XIII 37;630. The lowest and the highest recorded yields are: I 30; 540 II 40; 1110 III 830; 3170 IV 1200; 10,650 V 90; 2100 VI 120; 2100 VII 210; 2200 VIII 60; 520 IX 30; 600 X 20; 170, XI 20; 640 XII 30; 130 XIII 50; 1820 (from Geiger 1957).

region IV. Region V is represented by nearly 80 records, and the average sediment yield is about 1000 tons/km².

The values indicated by this map differ considerably from those shown on the UNESCO-map and on Judson & Ritter's map, and in fact they are not quite comparable. This is largely due to differences in selection of river basins used for the analyses, mainly the sizes of the basins investigated. Fournier's map shows high values in the western mountainous half of the country and lower values in the east, whereas the map of small basins indicates the reverse (Fig. 48B).

One large scale map of the Pacific area, namely of western Oregon is reproduced in Fig. 46 from Anderson (1954). It was constructed with the help of the multivariable equations 78 and 79 (cf. the Appendix). Equation 78 was

based on data from 29 basins. The basic data are discussed in the Appendix. The forest land was divided into 39 groups on the basis of differences in streamflow, topography, soil and land use. The results of the calculations with equation 78 were grouped to form 10 erosion zones in forest land. Agricultural land had an average rate of annual sediment yield of 84 tons/km²/yr as calculated by equation 79. A map of relative erosion was then used in order to calculate sediment yields in agricultural areas exhibiting slight, moderate and severe erosion.

The resulting map is very detailed (Fig. 46). A simplified version has been prepared for this study and is found in Fig. 47. High values of sediment yield, 100–150 tons/km²/yr, are indicated in forested areas in southwestern Oregon. Furthermore, there are quite large agri-

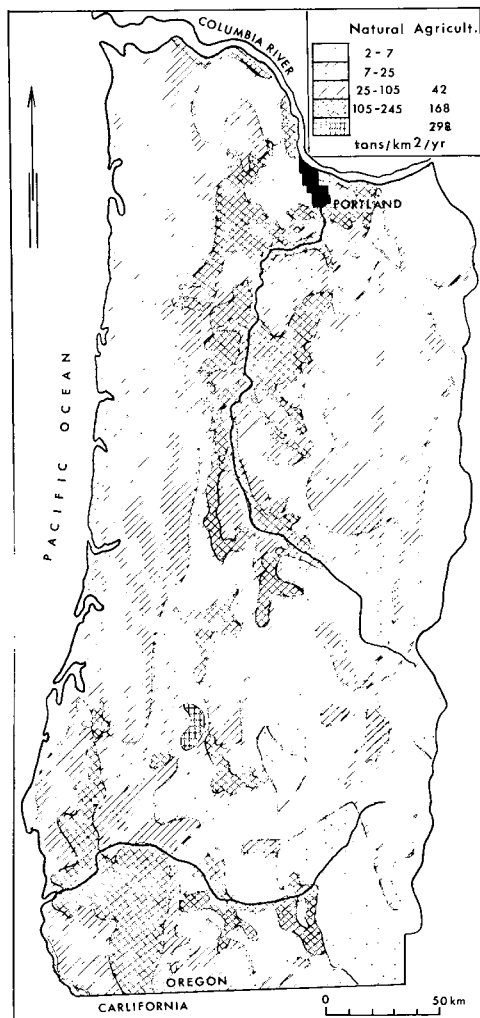
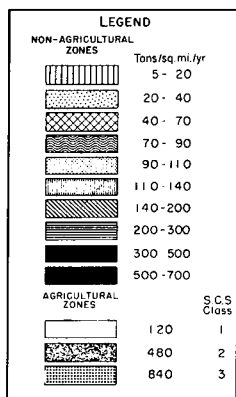
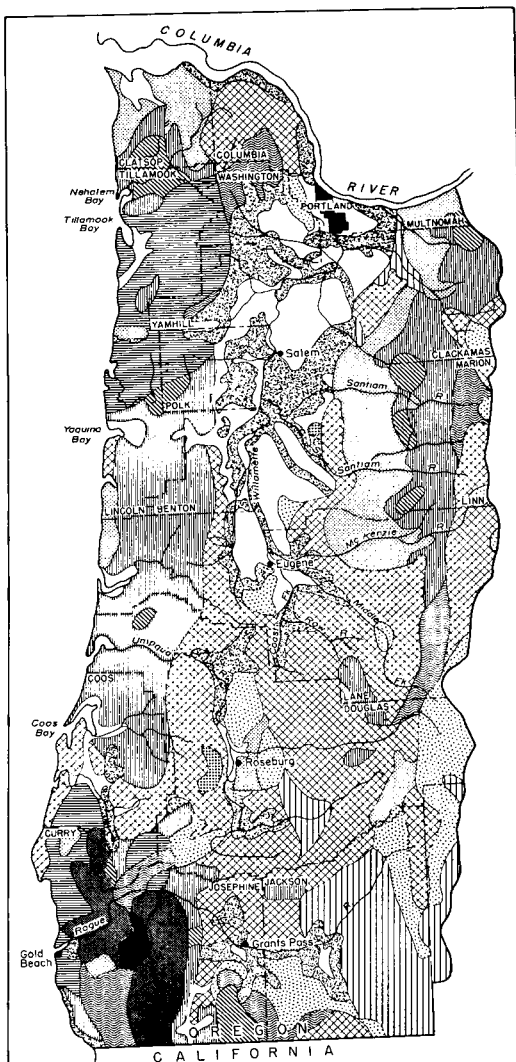


Fig. 47. Simplified version of Fig. 46.

Fig. 46. The original erosion map of western Oregon. 3.6 data sets/10,000 km² (from Anderson 1954).

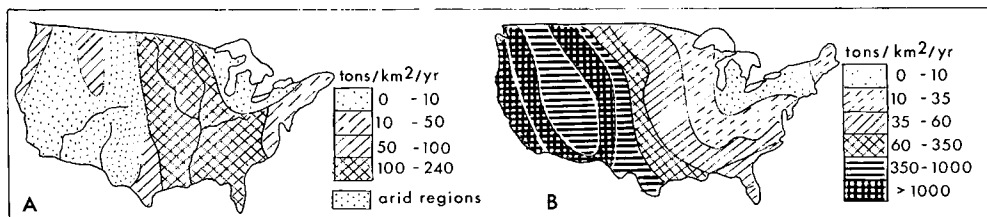


Fig. 48. The USA in global maps A: by Strakhov (1967), B: by Fournier (1960).

cultural areas with a mean sediment yield of about 170 tons/km²/yr.

Compared with the regional map in Fig. 46, both the UNESCO map and Judson & Ritter's seem to have acceptable values in the actual region. The value 50–100 tons/km² on Strakhov's map for western Oregon seems to be a quite acceptable generalization for a small scale map (Fig. 48). The value on Fournier's map, however, is 10 to 20 times too high (Fig. 48).

A map of a tributary basin to the Columbia river, the Walla-Walla river basin of 4558 km², in eastern Oregon and Washington, shows rather high sediment yields (Mapes 1969). The map is not reproduced in this study. The whole basin yields 824 tons/km² per year which is in the range of values on Fournier's map. Values > 1000 tons/km² are found in some intensely cultivated loess soil areas in the Walla-Walla basin. The area is marked as arid land on Strakhov's map.

Anttila & Tobin (1978) drew a sediment yield map of Ohio for drainage areas larger than 130 km² (Fig. 49). The minimum drainage area used to define annual sediment yield was 164 km². Maddock's classification was used for determining bedload (Lane & Borland 1951). Bed load was estimated to be < 10 % of the suspended sediment yield at 50 out of 51 sediment gauging stations. Mean annual sediment yields were calculated for the 25-year base period, 1946–1970, by means of sediment rating curves and flow duration curves. 62.5 %

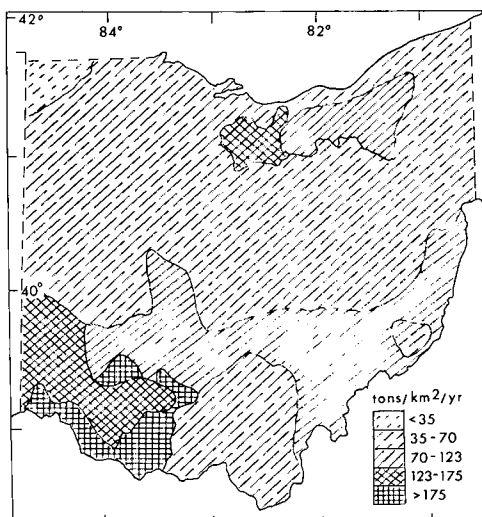


Fig. 49. Sediment yield in Ohio. 4.8 data sets/10,000 km² (from Anttila & Tobin 1978).

of the area has a sediment yield of 35–70 tons/km²/yr. Higher sediment yields are found in the southern part of the state. Fournier indicates 0–35 tons/km²/yr in Ohio. On Strakhov's map a line separating 10 tons/km²/yr from 100–240 tons/km²/yr goes through Ohio. The UNESCO map indicates 90 tons/km²/yr and Judson & Ritter's map 94 tons/km²/yr.

Fig. 50 is a map of South and North Dakota west of the Missouri (Livesay 1975). It shows sediment yields from basins larger than 260 km². Reconnaissance inspections were carried out by two or more specialists, who had the experience required for making estimates of sediment yield rates. Sediment sampling or

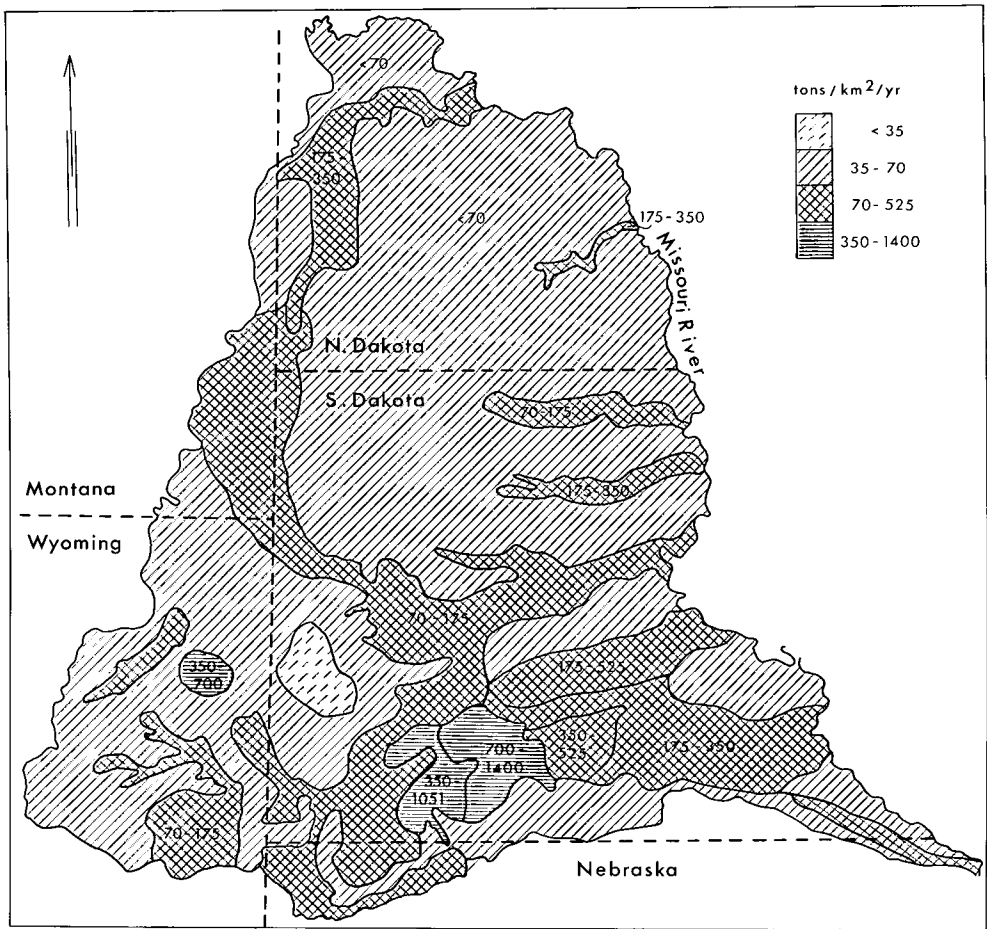


Fig. 50 Sediment yield in the Western Dakotas (from Livesay 1975).

reservoir surveys were made at representative points within the study area. A yield contour map was then developed. Using digitizing or planimetric processes, drainage area increments of sediment yields were summed and checked against measured rates in the area. Sediment yield values vary from < 35 tons/km²/yr to 700–1400 tons/km²/yr (Fig. 50). More than 50 % of the region has values < 70 tons/km²/yr. The UNESCO map shows 90 tons/km²/yr and Judson & Ritter 94 tons/km²/yr. Strakhov's map shows 100–240 tons/

km²/yr and Fournier's map 35–350 tons/km²/yr in the Dakotas.

The map of Wisconsin in Fig. 51 is based on suspended sediment yield measurements at 84 stations (Hindall 1976). Sediment rating curves and duration curves were used to calculate the annual suspended sediment yield. Sediment yields were adjusted to a 10-year base period, 1961–1970, and the mean values were plotted at the centre of each drainage basin. The highest yields in Wisconsin are indicated in the southwest, an area which was not glaciated during the

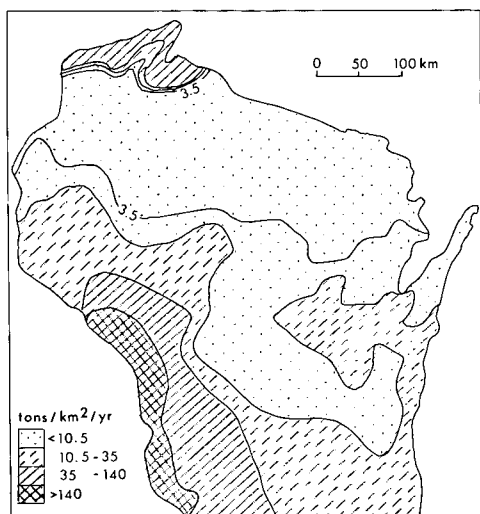


Fig. 51. Suspended sediment yield in Wisconsin. 6.0 data sets/10,000 km² (from Hindall 1976).

Pleistocene, and where the silty soils on very steep slopes are thought to promote high sediment yields. The Lake Superior red clay area also has rather high values. The red clay is thought to retard infiltra-

tion and promote rapid surface runoff. The lowest average annual suspended sediment yields are found in the Northern Highland province. Large areas of heavy forests, very little crop farming and low relief combine to keep yields low. The average suspended sediment yield in Wisconsin is about 28 tons/km² per year. In the UNESCO map, the south western part of Wisconsin shows a yield of 90 tons/km²/yr. Strakhov's map shows 10 tons/km²/yr and Fournier's map 10-35 tons/km²/yr (Fig. 48).

15.1.10 Pakistan

Ahmad (1960) prepared a map of northern Pakistan around 35° latitude (Fig. 52). This map is based on data from 17 sediment gauging stations, collected during the years 1953 to 1958. A bed load estimation of 20 % of the suspended load has been added. Most of the gauging stations are located in mountainous terrain, and only two are on the plains. Sediment

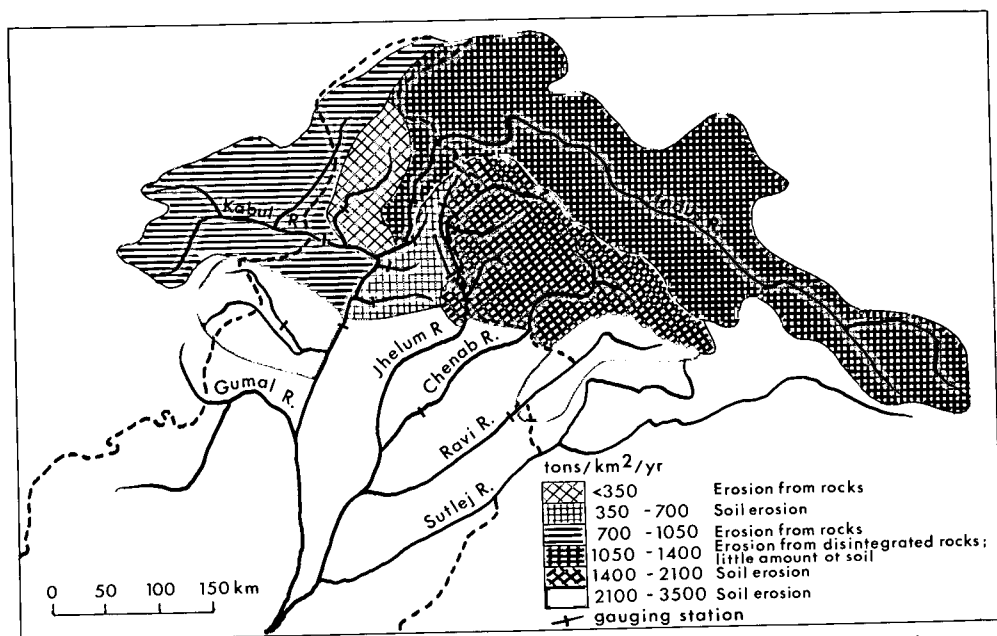


Fig. 52. Sediment yield in northern Pakistan. About 0.4 data sets/10,000 km² (from Ahmad 1960).

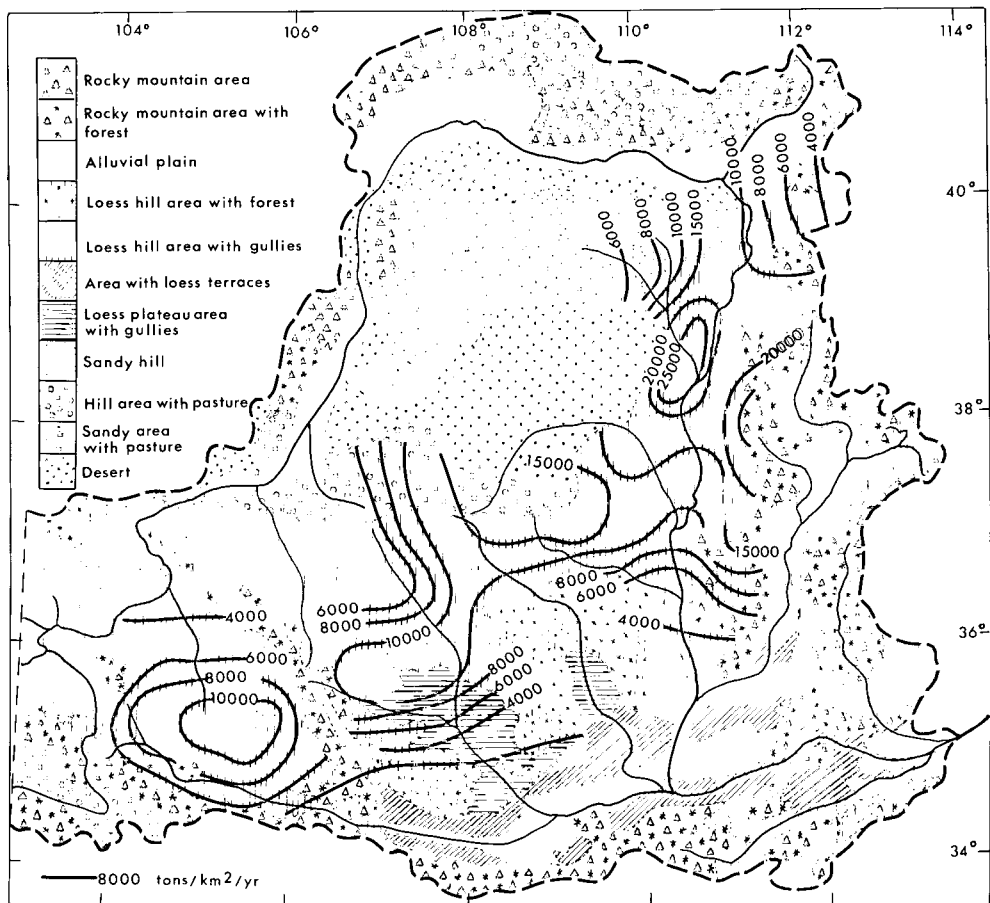


Fig. 53. Sediment yield in the Yellow River basin (from Gong & Xiong 1980).

in water diverted for irrigation from the Chenab river was estimated and added to the sediment measurements. A considerable part of the Ravi basin consists of flat cultivated land. The Ravi basin has a high erosion intensity. A second area with a high erosion intensity is found in a low mountain area with predominant grazing. Fournier's map does not show any erosion rate indication in this part of Pakistan. It is mapped as arid land on Strakhov's map.

15.1.11 China

The map in Fig. 53 of the upper and

middle-reaches of the Yellow River basin of China covers the area between 34° and 41°N (Gong & Xiong 1980, cf. Kung & Chiang 1977). It was constructed using data from sediment measurements in gullies, and in 1st, 2nd and 3rd order tributaries to the Yellow River. Values from gullies and small tributaries can be used because the sediment delivery ratio is very near 100 % (Gong & Xiong 1980).

The highest erosion rate of 25 000 tons/km²/yr is found in northern Shensi. The topsoil of this region is mainly Malan loess, which is very loose and has columnar voids. The silt content is generally over 60 %. On other parts of this map,

soil loss falls to 2000 tons/km². The very high sediment yield values in the Yellow River basin are certainly due to the erodible soil, the pronounced seasonality of the precipitation and the intense agriculture. The maximum value of 4500 tons/km²/yr reported from a basin in Soviet Central Asia is very much exceeded in large parts of the Yellow River basin.

Strakhov's map indicates > 240 tons/km² in the area covered by the regional map. Fournier also has comparatively low values in this region, although 1000–2000 tons/km² is indicated in the southwestern corner of the area.

15.1.12 Japan

The map of Japan in Fig. 54, is based on reservoir data and shows mean annual sediment yield values for rivers having two or more dams within their drainage basins (Kadomura 1980). The values have been transformed from m³/km²/yr, by assuming a rock density of 2.65 tons/m³.

The mean annual sediment yield for the whole of Japan is 1243 tons/km²/yr. Figure 54 shows especially high rates of sediment yield in the Central Highlands and in the outer zone of southern Japan, where quaternary uplift has been high, and where the fractured rocks in many cases yield sediment on the order of 2650 to 13,000 tons/km²/yr. According to Yoshikawa (1974, referred to by Kadomura 1980) the highest rate of "natural erosion", approximately 26,500 tons/km²/yr, is found in sub-alpine and alpine environments in the Northern Japanese Alps. This is a region with exceptionally steep slopes, heavy rains, and highly erodible material of volcanic origin. On the other hand, sediment yield values in some experimental catchments in hills and uplands covered with thick well-managed forests are of the order of 26–132 tons/km²/yr.

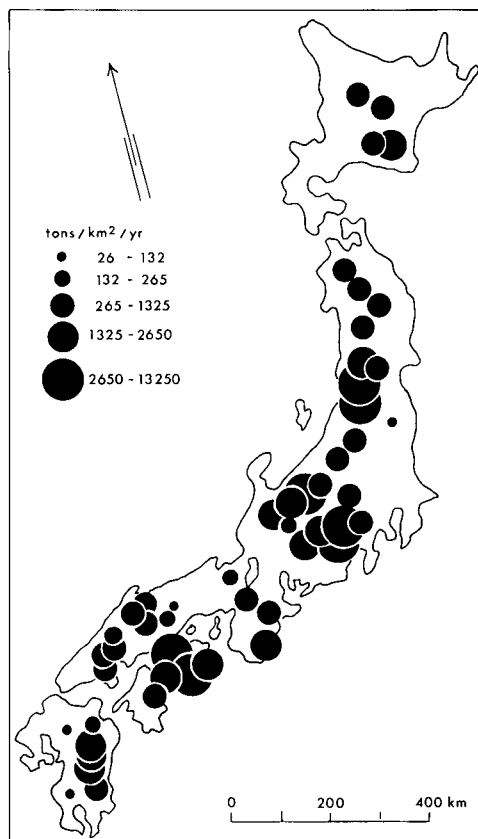


Fig. 54. Sediment yield in Japan. More than 2.8 data sets/10,000 km² (from Kadomura 1980).

On Fournier's map, the annual sediment yield in southern Japan is mapped as 1000–2000 tons/km²/yr and in the northern part of Japan as 600–1000 tons/km²/yr. This is in good agreement with Kadomura's map. Strakhov's map is difficult to interpret.

15.1.13 Morocco

The latitude of 35° cuts through Morocco. Fig. 55 is a reconnaissance map, which is based on known sediment yield values from 23 stations (FAO 1975). The highest values in Morocco are found

around 35° latitude in the northern part of the country. In the Low Rif basins the parent material is marl in several areas, and sediment yields are much higher than in the High Rif basins. Badlands give very high yields of up to 28,000 tons/km²/yr. Fournier's map shows 60–1000 tons/km²/yr. Strakhov's map shows 10–50 tons/km²/yr which is much too low.

15.1.14 South Africa

The latitude of 30°S cuts through South Africa (Fig. 56). Although this map is based on sediment yield surveys of reservoirs, it only indicates the relative rate of sediment yield (Schwartz & Pullen 1966). The division into zones was based on topography, climate, geological features and type of veld. Sediment yield values have been converted from morgen feet/sq mi/yr by assuming the dry weight to be 50 lb/ft³, a figure used by the authors. On comparison with this map, Strakhov's values of 50–100 tons/km²/yr are too low while Fournier's map coincides quite well.

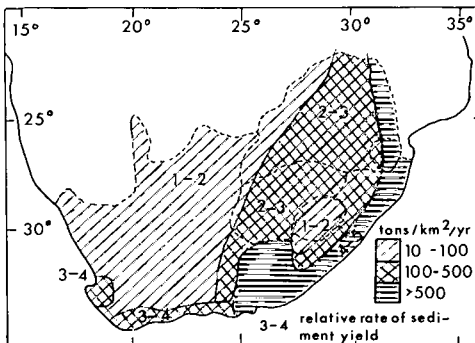


Fig. 56. Relative rate of sediment yield in South Africa, 1 = very low, 2 = low, 3 = moderate, 4 = high 5 = very high. 0.2 data set/10,000 km² (modified from Schwartz & Pullen 1966).

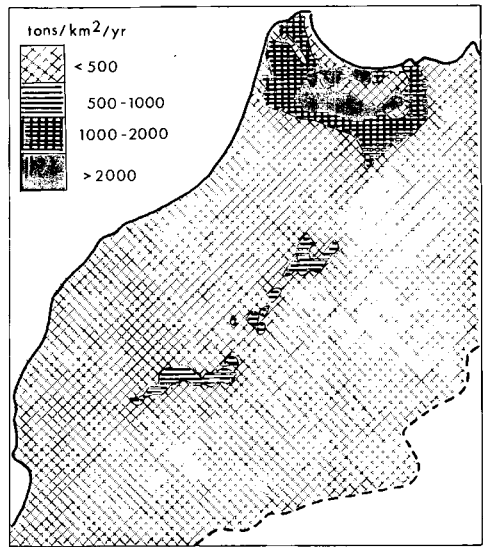


Fig. 55. Erosion map of Morocco. 0.5 data set/10,000 km² (from FAO 1975).

15.1.15 The Cameroon

The map of Cameroon in Fig. 57 is the only regional sediment yield map from a tropical climate included in this study (Olivry 1977). According to Olivry, erosion in the Cameroon is highly dependent

Some sediment yield values from the groups, in tons/km²/yr.

1. 12, 16, 28, 36
2. 73, 81
- 2-3. 121, 137, 161
3. 202, 202, 242, 242, 242, 242
4. 484
- 4-5. 645, 1209
5. 1612

on soil and vegetation. In a southeastern drainage basin of 77,000 km², which exhibits a value of 28 tons/km²/yr, the soils are ferralitic and the vegetation cover is 30 % forest and 70 % wooded savanna. In a neighbouring basin of 42,300 km² the sediment yield was 85 tons/km²/yr as a result of high erosion in a volcanic, mountainous area with a high percentage of cultivated land. Sediment yields were still greater in the northern Cameroon, which has a long dry season and seasonal vegetation cover.

Strakhov's map indicates 50–100 tons/km²/yr for the whole of the Cameroon. Fournier's map indicates a varied pattern with more than 1000 tons/km²/yr near the coast and in the northern part of the country, and 600–1000 tons/km²/yr in the central regions.

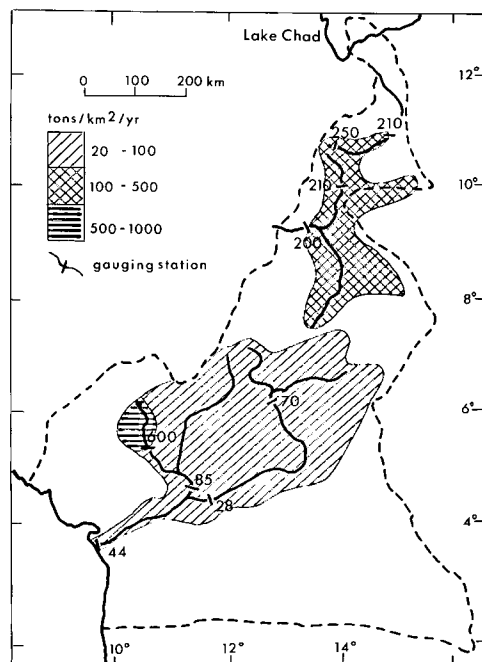


Fig. 57. Suspended sediment yield in the Cameroon. About 0.2 data set/10,000 km² (from Olivry 1977).

15.2 General discussion of the reliability of regional maps

A map can be assumed to be reliable, if it is based on a large number of well distributed gauging stations, with an appropriate time distribution of measurement taking and a reliable sampling technique for determining sediment concentration. The measured areas must also be representative of larger, unsampled areas. Examples of maps, which are based on a large number of basins per unit area (> 2 data sets per 10,000 km²) are Rumania, Italy, Cyprus, Poland, Western Oregon, Wisconsin, Ohio and Japan. The maps of Yugoslavia, Morocco, South Africa, the Cameroon, Pakistan, Canada (Fig. 40), the USA (Fig. 45) and the USSR (Fig. 36) are based on a less dense data net (< 1 data set per 10,000 km²). Canada (Fig. 41) and the USA (Fig. 43) have < 0.1

data set per 10,000 km², but the investigated basins cover large areas.

The methods of estimating sediment yield in unsampled areas are of great importance to the reliability of the maps, especially when there is a small number of measured values and when the distribution of measured values is biased. In studies where there is no information about the method of estimating values in unmeasured areas, the assumption of similar sediment yield in neighbouring or similar basins might have been used. In Ohio, extrapolations have been made after comparison within eight 'resource areas'. In the Western Dakotas, reconnaissance inspections by sedimentation specialists were undertaken to sort out high and low erosion areas within larger basins.

The map of Italy is an example of a map in which sediment yield is corre-

lated with one subjectively chosen variable. Estimations of sediment yield in unmeasured areas are based on a correlation with the areal percentage of ground in highly erodible rock types. However, agricultural land use might contribute to great sediment production and it might have been preferable to select land use in combination with lithology. As the map is based on so many measured values, it should nevertheless provide a good approximation of real sediment yield. Exceptions are Sicily with two gauging stations and Sardinia, where there is only one gauging station. Rock type is thought to be of great importance to the rate of erosion in other regions too. In Cyprus, flysch is associated with high erosion rates on sloping land. In Soviet Central Asia, Cretaceous and Tertiary deposits, loess and loesslike rocks are said to be susceptible to intense mechanical denudation.

In a few cases, equations are used to estimate sediment yield in unmeasured areas. In both Yugoslavia and Rumania calculations are based on exponential equations of water discharge, incorporating different coefficients for different regions. In Rumania there are even different coefficients for different mean elevations of the basins within the regions. The advantage of this method is that the equations can be used to estimate sediment yield in larger areas, since water discharge measurements are more widespread than sediment discharge measurements. The method is also quite simple to employ. However, there is a subjective component whenever unmeasured basins are included in the different regions.

A further development of this method would be the use of a multiple regression to calculate the coefficient in the discharge relationship. An equation of the type illustrated by equations 49–51 in the Appendix might either be used to calculate instantaneous sediment yield values

or to calculate monthly values. Equations 55, 58, 60 with mean annual water discharge as one variable are examples of multiple regressions in which other variables determine the coefficient in an annual relationship.

The map of western Oregon is based on multiple regression analysis. This may be regarded as the least subjective way of constructing a map. It is, however, a tedious method, as it includes many different variables. Even though the method of constructing the map is worthy of being more generally applied, the map of Oregon has some weaknesses because of infrequent sampling at high sediment concentrations and the questionable method of calculating high discharge sediment yield (cf. the Appendix).

15.3 Summary of the reliability of global maps based on a comparison with regional maps

Strakhov's map is in good agreement with the regional maps of the Soviet Union, apart from an area from the Black Sea eastwards over the Soviet Central Asia which is mapped as arid land. It also agrees with the regional maps of western Oregon, Wisconsin, southern Ohio, the Dakotas, Poland (although the values for Poland are somewhat high) and the southern Cameroon as well as with the UNESCO map of the Amazon, Orinocco and Parana basins. Strakhov's map indicates low values compared to the regional maps in northern Ohio, Yugoslavia, Rumania, Italy, the Yellow River basin, Morocco, Canada, the USA (but fairly good east of the Rocky Mountains compared to Judson & Ritter's map in Fig. 43) and South Africa. Part of South Africa is mapped as arid land, and this is also the case in Pakistan. The low values by Strakhov in the latter group may be

partly due to 'the area problem' (cf. chapter 1 and section 4.4). Strakhov's map is based exclusively on large rivers which have comparatively small sediment yields per unit area. Regional maps are generally based on small and intermediate basins and only in exceptional cases on rather large basins. Strakhov's map has no areas with considerably higher values than the regional maps.

Fournier's map agrees comparatively well with the regional maps of Rumania (except for the western part), Italy, Poland (although the values for Poland are somewhat high), Morocco, South Africa,

Japan, Argentina (in the UNESCO-map) and Wisconsin. It does not fit the pattern of the regional maps of Yugoslavia and the Soviet Union and the values are too high for the Cordillera of Canada and of the United States including Oregon. The values are also too high in the Cameroon and the Orinocco, Amazon and Parana basins where natural vegetation is preserved to a great extent. Low values are indicated in Ohio, the Western Dakotas, the eastern USA, Soviet Central Asia and the Yellow river basin compared to the regional maps. Pakistan is not mapped.

16 Qualitative soil erosion maps and other special types of maps

16.1 Maps of topsoil loss

16.1.1 *The USA*

The United States Department of Agriculture, Soil Conservation Service (USDA, SCS 1948), issued a map with three degrees of erosion (Fig. 58). This map has an original scale of 1:10 000 000 and is mainly based on data from an erosion survey from 1934, which includes twenty-nine types and degrees of erosion. A number of other soil conservation surveys were also considered.

In the Reconnaissance Erosion Survey 1934 mentioned above, soil loss was estimated by comparing the remaining soil horizon with the depth of a noneroded soil horizon. The field work was accomplished by soil surveyors. Wind erosion as well as sheet and gully erosion was investigated. 25 % or more of the land was affected as indicated on the map (Bennett 1939). This method of estimation is very subjective, and its precision is completely dependent on the

skill and the consistency of the field workers. They worked for the most part at a scale of 1:63 360. The great reduction in scale to 1:10 000 000 necessitated much generalization and the omission of many small areas (Conserv. Found. & FAO 1954).

A map showing average annual potential sheet and rill erosion on cropland is given in Fig. 59 (USDA, SCS 1977). It is a map of potential erosion but is presented here to simplify comparison with the map of topsoil loss. Soil erodibility, dominant slope lengths and slope gradients, crops and management systems were tabulated for different land capability sub-classes. The USLE equation was employed to calculate values within each sub-class. Mean values were calculated for each Major Land Resource Area. "West of the Rocky Mountains direct estimates of soil losses were provided by SCS personnel" (USDA, SCS, 1977, p. 12).

The low erosion rates in the West are

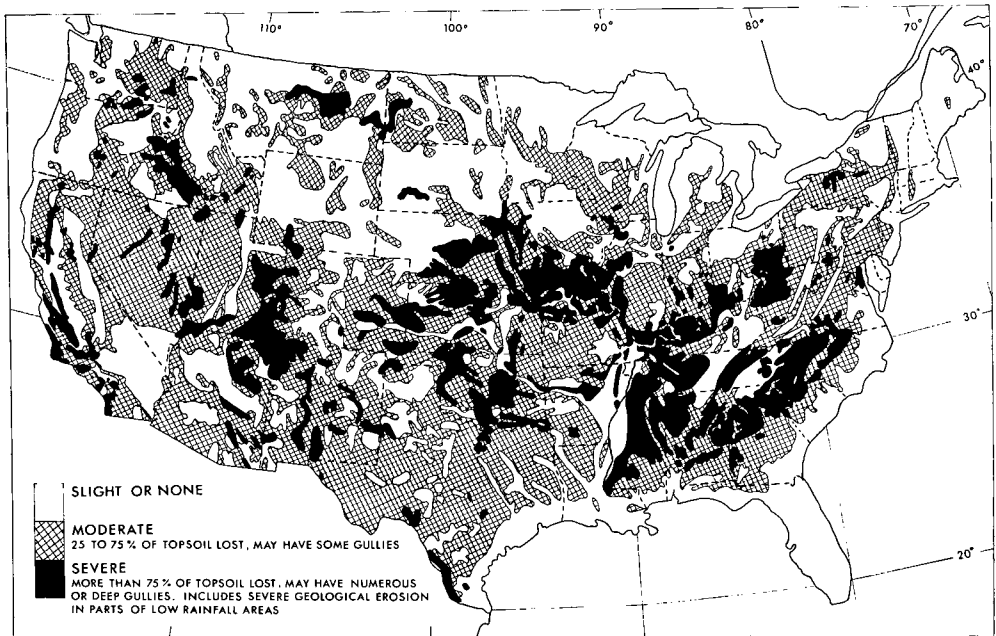


Fig. 58. Degree of topsoil loss in the USA (from USDA, SCS 1948).

partly due to low average erosivity values (cf. Fig. 21). One conclusion which may be drawn from a comparison between the two maps mentioned above is that in the western USA severe erosion may be caused by grazing or it may result from geological erosion in areas with a thin top soil. Such a comparison also indicates that in regions where cropland constitutes $\geq 25\%$ of the area, slight erosion means an annual soil loss from fields of 0–1000 tons/km²; moderate erosion, 1000–3000 tons/km²; and severe erosion, > 3000 tons/km². This is in agreement with the definition of proportions of eroded topsoil within the degrees of erosion.

Eroded soil from cropland contributes much to the sediment yield in streamflow in large areas of the USA. This contribution depends on the percentage of cropland which influences both the rate of gross erosion and the sediment delivery ratio (USDA, ARS 1975). Cropland increases surface runoff and provides little

obstacle to transport. Steepness is another factor affecting the sediment delivery ratio, as mentioned before.

Trimble (1974, 1975) drew a map of the depth of gross erosion in the southern Piedmont Plateau, covering the period of the European settlement to about 1940 (Fig. 60). The map was prepared from different erosion survey maps. Experienced soil scientists who had taken part in the surveys or who had considerable field experience in the general area of each survey were asked to assign average soil removal estimates to each erosion class or mapping unit. For each survey area, estimates were made by at least two soil scientists, in some cases three or four. The average depth from each survey area was then calculated.

In order to extrapolate the erosion rates from the 34 areas covered by detailed maps to the remaining 72.6% of the southern Piedmont, Trimble used the "Reconnaissance Erosion Survey" maps in 1:500 000, originally mapped in

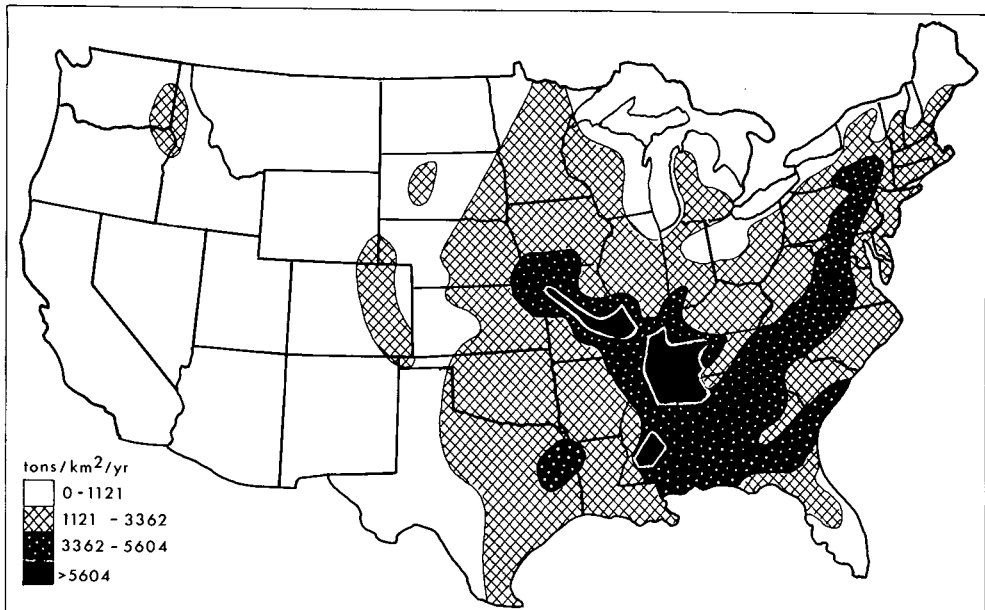


Fig. 59. Potential sheet and rill erosion on cropland estimated with the help of the USLE (from USDA, SCS 1977).

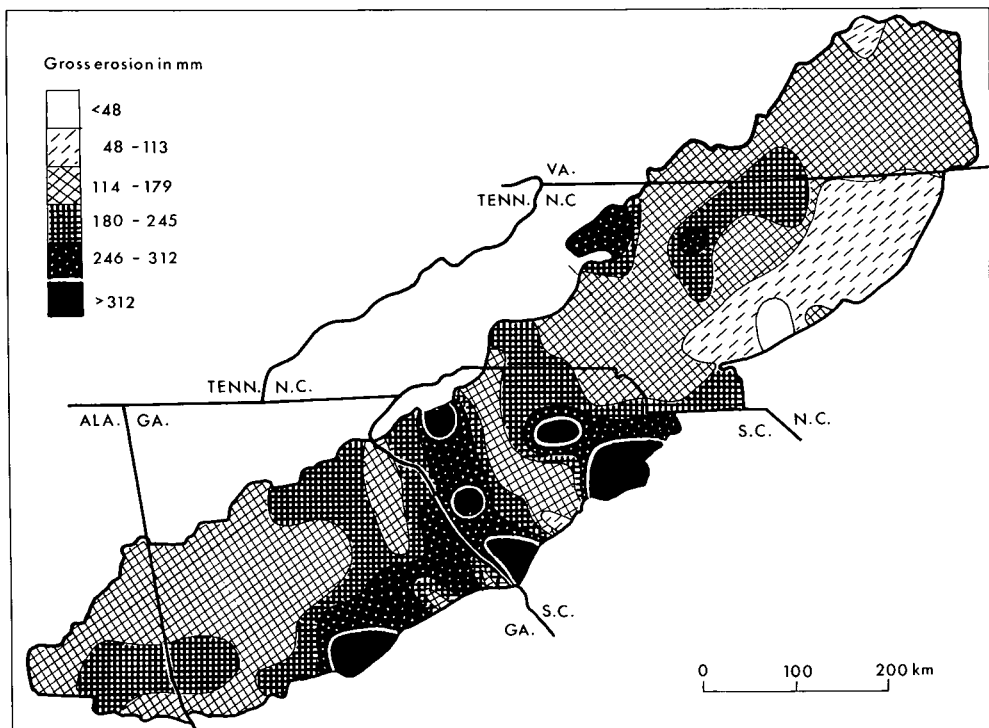


Fig. 60. Depth of gross erosion in the southern Piedmont Plateau. (from Trimble 1974). Assuming an erosion period of 150 years and a bulk density of 1.5 tons/m³, gross erosion for the intensity classes are: <480, 480-1130, 1140-1790, 1800-2450, 2450-3120, >3120 tons/km²/yr (present author).

1:63 360 by the USDA, Soil Conservation Service in 1934. The best fit values of the depth of soil removal in the 34 areas surveyed were:

Degree of erosion	Topsoil removed	Gross erosion in tons/km ² /yr assuming an erosion period of 150 years and a bulk density of 1.5 tons/m ³ (present author)
Little or none	0 mm	0 tons/km ²
Slight	106 mm	1060 ..
Moderate	150 mm	1500 ..
Severe	187 mm	1870 ..
Very Severe	399 mm	3990 ..
Destroyed by gullying	995 mm	9950 ..

In this way the whole area was covered and could be mapped county by county.

16.1.2 Latin America

Soil erosion maps of Latin America were prepared and published in 1954 (Conserv. Found. & FAO 1954). The methods of mapping were influenced by the Soil Erosion Survey of the United States in 1948. Only a small portion of the map is based on original field work. In many areas, where field work was not practicable, classification of degrees of soil erosion was conducted from aerial photographs.

The map is presented in two sections—Middle America and South America—including a general discussion of the influence of various factors upon accelerated soil erosion. The maps of Latin America cover A climates but they also comprise C and B climates. Exceptions are found in the mountains with regions of temperate and even cold climates. It is, however, difficult to compare this type of map with sediment yield maps, since maps of eroded topsoil show gross erosion experienced during a variable time period from a few years to hundreds of years. As the maps of Latin America were constructed about 30 years ago, population increase since then has certainly aggravated erosion today. Another difficulty is that the depth of the

original topsoil varies with climate, and accordingly the depth in mm represented by the respective degrees of erosion varies with climate.

The first section of the map covers *Middle America*. Middle America includes Mexico, Central America and the West Indies (Fig. 61). The Mexican part of the map is based on a soil association map prepared by the USDA and on observations during reconnaissance flights and interpretation of aerial photographs. Data for Puerto Rico were obtained from a soil erosion survey of the island.

Increasing population pressure in arid and semi-arid areas of Mexico has accelerated erosion, both because of overgrazing and because of the use of dry, marginal land for row crops. The loss of only an inch or two of the topsoil may give rise to severe erosion. Overgrazing is also the cause of soil erosion in the hilly lands of Cuba, although the climate there is humid. In the text concerning Mexico it was pointed out that "much of the soil erosion in the very humid mountain lands consists of landslips and slides rather than of sheet and gully erosion familiar in temperate regions" (Conserv. Found. & FAO 1954, p. 162).

Maize, coffee and tobacco cultivation in mountainous terrain is the second main cause of accelerated erosion. These are the conditions in Mexico and Central America, as well as in Jamaica, Puerto Rico and the Republic of Haiti. However, on flatter land in Cuba, Jamaica and Puerto Rico, where sugar cane has been cultivated for long periods, the topsoil normally remains thick.

The second section of the map covers *South America*. South America (Fig. 62) exhibits the same causes of great soil loss as Middle America. However, the extent of severe erosion is not so great as is the case in Middle America. Overgrazing is a problem in some areas but not in others. Much of the dry land around the Golfo de

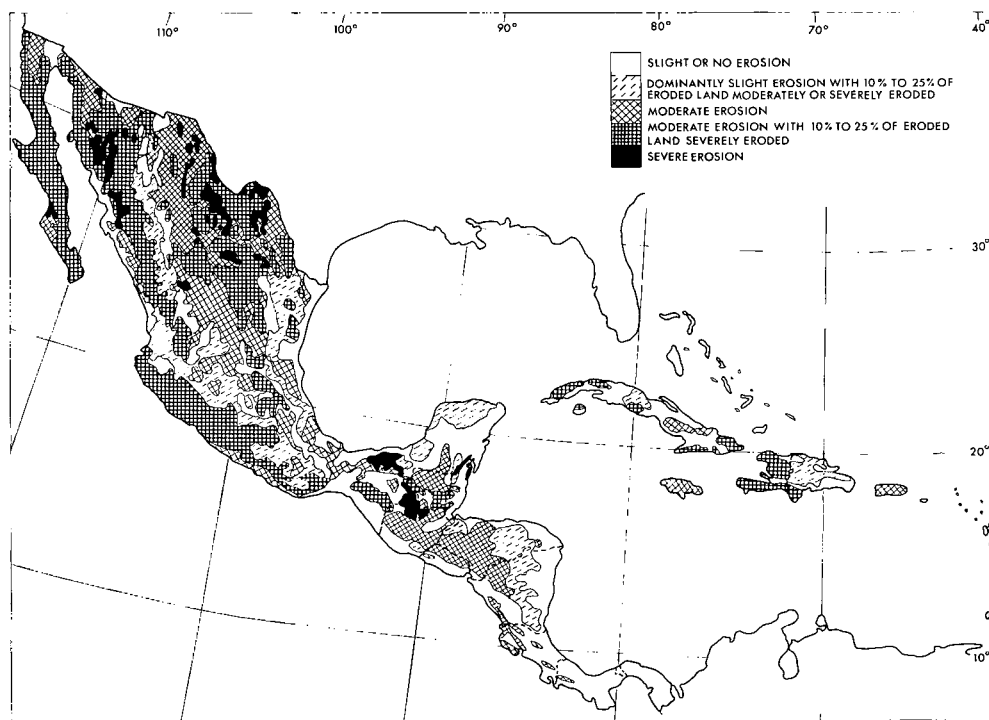


Fig. 61. Degree of topsoil loss in Middle America (from Conserv. Found. & FAO 1954).

Venezuela has been severely eroded as a result of overgrazing. In the drought zone of northeastern Brazil, which experiences torrential rain storms, cattle grazing has caused much accelerated erosion. In Patagonia, erosion is caused by wind in sheep-grazing areas. Another map of Argentina indicates that water erosion is also prominent (Inst. Natl. de Technol. Agropecu.). Compared to the dry lands of Mexico, erosion in this colder and less densely populated area is relatively small.

Sheep grazing in the Andes has caused much erosion. In northern Ecuador the soil lies above an impermeable cemented layer at a depth of about 0.9 m. During periods of heavy rainfall, the soil is super-saturated and becomes a semi-liquid mass which flows with gravity, leaving the hardpan exposed. In central Colombia, slides are common, particularly

where hard impermeable layers are found near the surface. By tradition, pastures are regularly burned in Colombia. On the plains of eastern Colombia, which experience extensive grazing, accelerated erosion is only localized in the vicinity of water holes. Nor is erosion a problem in the Venezuelan plains of the Orinocco, despite the unfavourable distribution of rainfall with extremely wet and dry seasons and the use of the land for cattle grazing. The population density is low. In Uruguay, overgrazing was not considered a major cause of erosion in 1954.

Row crop cultivation and coffee plantations constitute serious erosion hazards. Maize is a chief subsistence crop, and maize cultivation in steep terrain is the cause of much erosion in South America. Severe erosion is found in the Venezuela Highlands where about

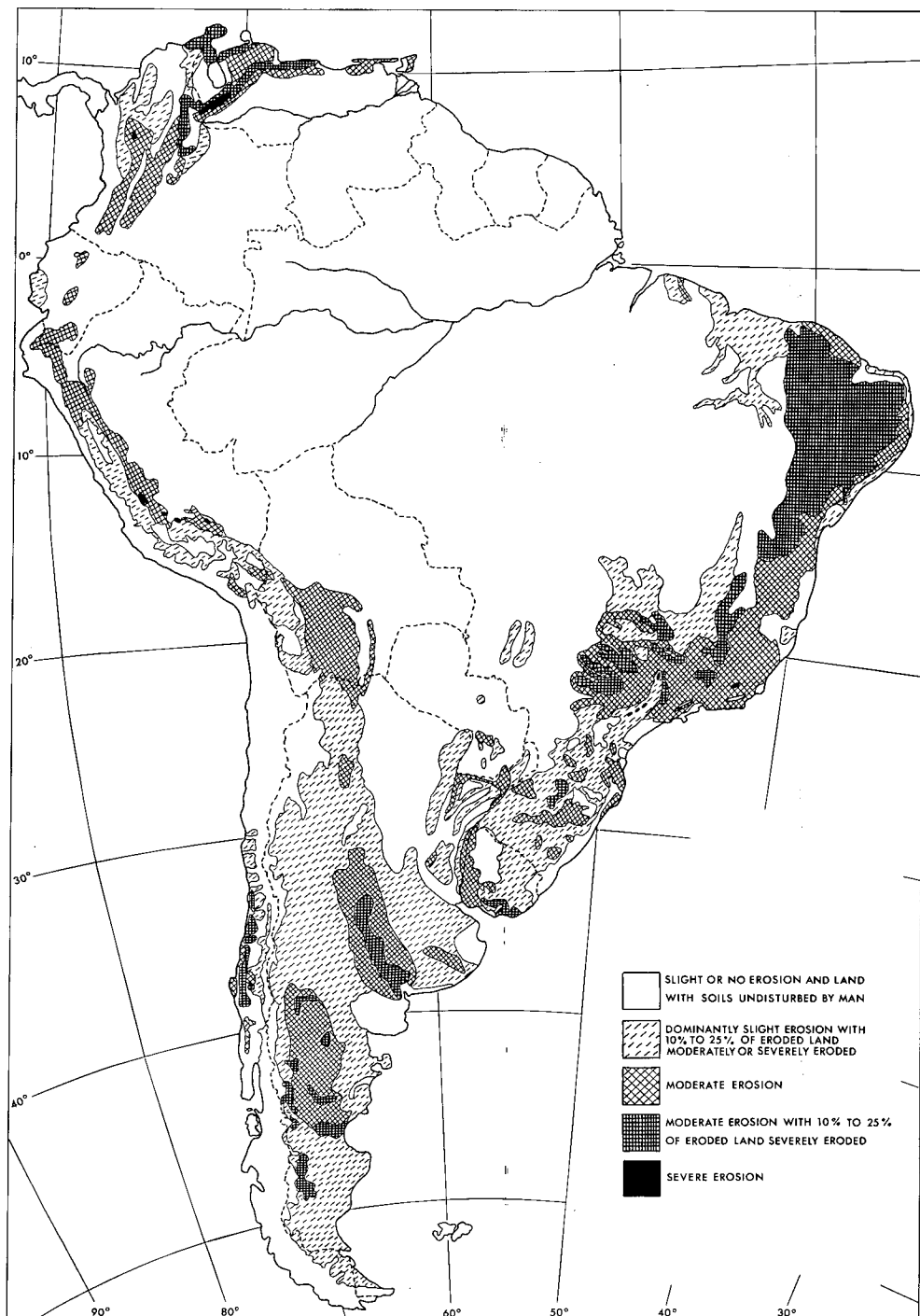


Fig. 62. Degree of topsoil loss in South America (from Conserv. Found. & FAO 1954).

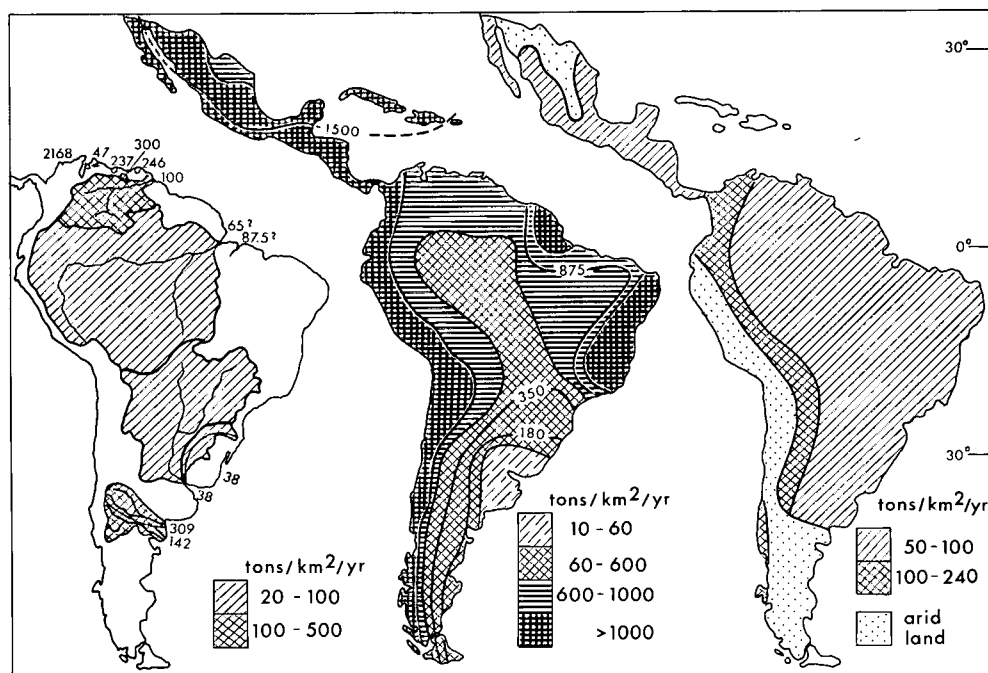


Fig. 63. Global maps of South America (from UNESCO 1974, Fournier 1960, Strakhov 1967).

75 % of the cultivated area is found on slopes $> 25\%$. Wheat-growing districts of Argentina and Uruguay experience much wind and water erosion. Coffee plantations have ruined much soil in Brazil and Colombia.

The global maps of South America give quite different impressions when studied together (Fig. 63). The few values on the UNESCO-map do not help in evaluating the other global maps. The great differences between the maps by Fournier and Strakhov can easily be observed as well as the differences between and similarities with the map of topsoil loss in Fig. 62.

16.1.3 Hungary

Stefanovits (1964) produced a map showing the degree of eroded topsoil for Hungary (Fig. 64). The work was undertaken using military maps on a scale of

1:75 000. Degree of erosion was determined from soil profile estimations, plot measurements of rill erosion, measurements of gullies per unit area and the extension of shallow gullies. Forested areas were not considered.

Soil profiles in representative regions were investigated. The depth of the A and B horizons was measured and compared to noneroded profiles of similar parent material in close proximity to the investigated profiles. Most of the noneroded profiles were situated on hillcrests and plateaus.

Rill erosion was estimated by measuring the cross-sections of rills along three lines across a plot and multiplying the mean cross-section area by the length of the plot. The volume was then multiplied by the soil density. The degree of erosion was considered to be slight, if measured values were < 3475 tons/km²; moderate, for 3475–8690 tons/km²; and severe for, > 8690 tons/km².

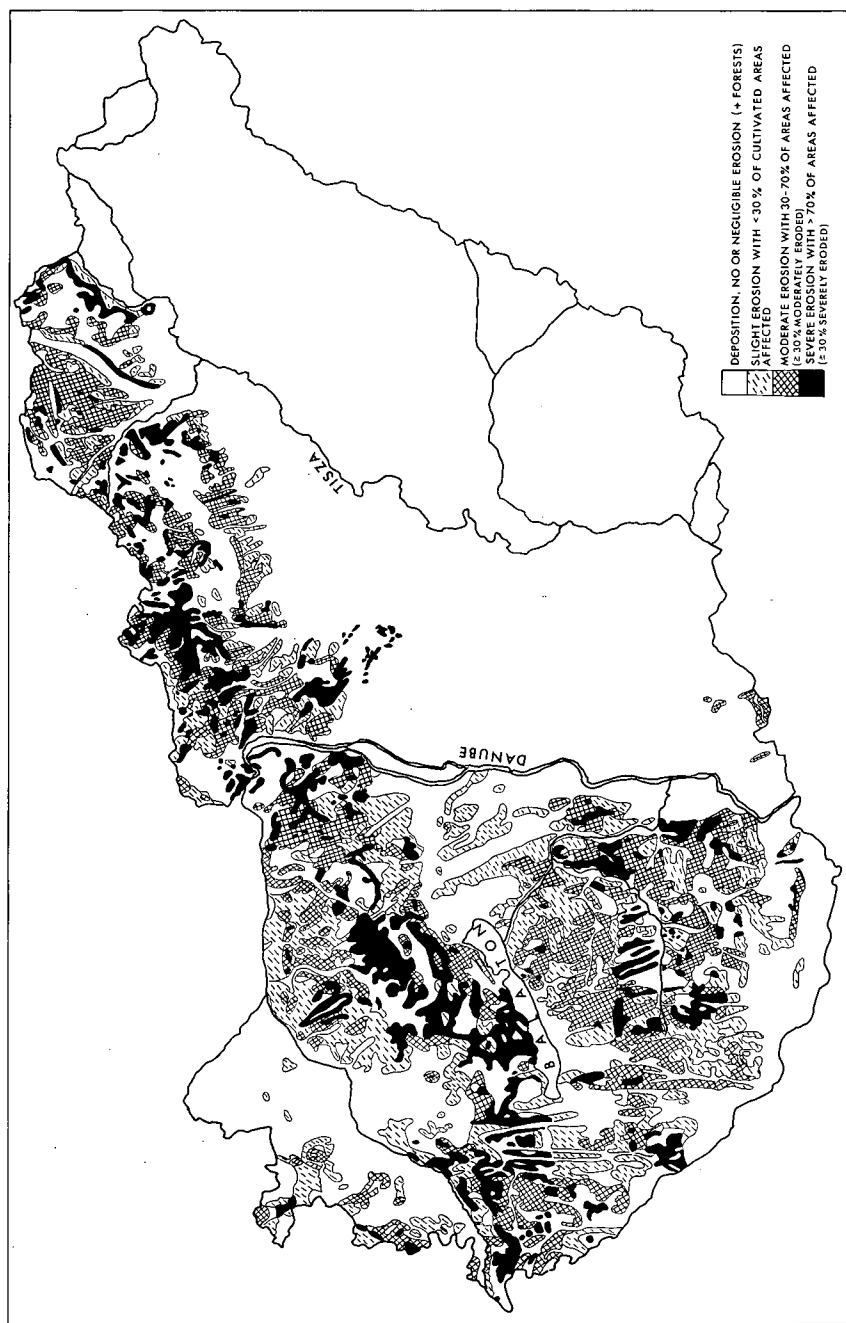


Fig. 64. Degree of eroded topsoil in Hungary (from Stefanovits 1964).

The degree of erosion by gullies was considered to be slight if the total length of gullies was shorter than 200 m/km²; moderate, if the length of gullies was 200–500 m/km²; and severe, if the length of gullies was greater than 500 m/km². In regions where soils are rich in sodium and crusts occur in the profile, the soil easily flows away if it is saturated. Erosion was considered slight if < 10 % of the area was affected; moderate, if 10–30 % was affected; and severe, if > 30 % of the area was affected.

In order to make the distribution of degrees of erosion clearer, a map was constructed to show the percentage distribution of severe, moderate and slight erosion within small districts of Hungary. There is also a map of gully dissection as well as a map of expected future erosion in Hungary which is based on the present degree of erosion and the expected evolution in different types of soils.

16.1.4 Northern Ghana

An erosion map of the savanna areas and the drier parts of northern Ghana has been derived from a soil map and a study of air photographs and field checking along roads and paths (Adu 1972). The map is not reproduced in this study. Moderate erosion was assigned to areas where 0.9 m of the original A and B horizons had been removed. Moderate erosion occurred in areas with gentle slopes of 2–3 % and was found in 40 % of the region. Severe erosion was assigned to areas where the A and B horizons, which were originally about 1.2 m thick, had been totally eroded exposing a stiff, weathered C horizon. Severe sheet and gully erosion was found around the more compact settlements.

16.1.5 China

A map of erosion zones was prepared by

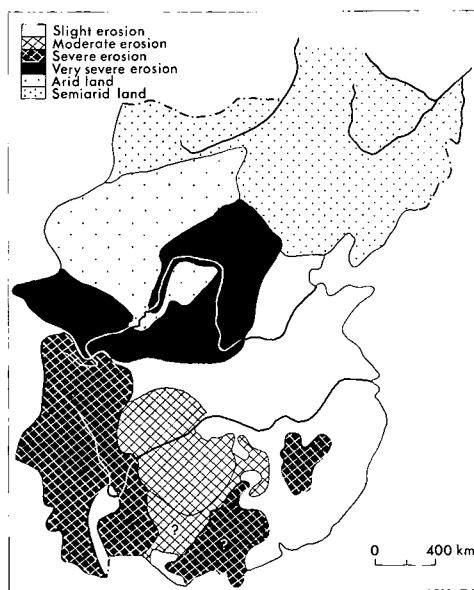


Fig. 65. Erosion zones of China (from United Nations 1951).

Chu and reproduced in United Nations (1951). The map is placed here despite the uncertainty about how the degrees of erosion were determined (Fig. 65). The original map has detailed indications of the distribution of different soils. It is stated in the text that the map is based on scattered information published by the National Geological Survey of China and on general descriptions of soil erosion given by Thorp, Lowdermilk and others (United Nations 1951). Very severe erosion is said to be associated with loamy loessial soils and sparse vegetation cover (Fig. 65). Severe erosion is said to occur where rainfall is abundant in the parts of the southern mountainous region which have reddish clay and meadow soils. Slight water erosion is assigned to the semi-arid region. In southern China, Fournier's pattern of erosion rate is in rather good agreement with Chu's map, but there are discrepancies around the Shantung peninsula and in the northern part of China.

16.1.6 Land capability maps

Maps constructed for land use planning—often called land capability maps—usually include codes of type of erosion and degrees of erosion such as slight, moderate and severe. It is difficult to extract information about soil erosion conditions from land capability maps, because they are large scale and they are cluttered with symbols. Therefore it is difficult to obtain a clear picture of the degree and type of erosion.

16.2 Maps based on visible erosion features

An erosion map of the Murray region, New South Wales shows different types of erosion (Higginson 1974). Gully erosion and sheet erosion occur in the deeply dissected alpine high plateau in the head waters of the Murray River (Fig. 66). Erosion in the eastern section is caused by excessive clearing of timber, overgrazing and damage by bushfires. The rest of the area is flat or gently undulating semi-desert land, which is subject to wind erosion.

The map gives no quantitative information on erosion intensity. Areas associated with wind erosion may also be affected by water erosion. In a river with a basin area of 422 km², just outside the western part of the map, the sediment yield is 254 m³/km²/yr of deposited sediment. A similar map indicating sheet, gully, wind, glacial and coastal erosion is prepared for India (FAO 1977).

On a geological map of Tunisia, symbols have been used which indicate the dominant type of erosion (Belaid 1970):

1. Diffuse surface runoff
2. Surface runoff with local gullies
3. Continuous gully systems
4. Gully systems, badlands in places
5. Wadis with limited gullies

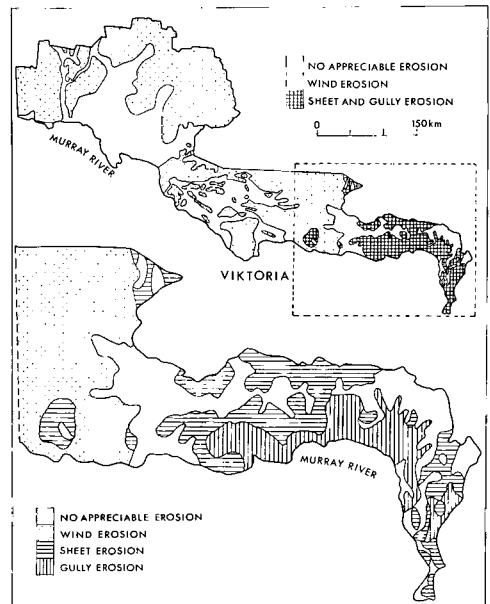


Fig. 66. The Murray region, New South Wales (from Higginson 1974).

6. Eolian deflation and accumulation of nebkas
7. Eolian deflation and accumulation of dunes

It can be concluded from the symbols that the map indicates types of erosion rather than erosion intensity. The amount of eroded material by diffuse surface runoff may be as high as the amount of gully-eroded material (cf. Rapp 1974).

There are many examples of local maps describing the extent of visible erosion and sedimentation (e.g. Doná 1962, Rapp *et al.* 1972). Williams & Morgan (1976) have constructed maps indicating type of erosion which include additional information about land use and which also include a slope overlay.

Hempel (1963, 1968) developed a scheme in West Germany to classify water erosion on fields 80–129 m in length, from visible erosion features as follows:

Degree 0: No erosion

- Degree 1: Visible erosion within the fields, only slight erosion forms
- Degree 2: Rills and furrows up to 10 cm wide and deep, erosion rills cross limits of the fields
- Degree 3: Rills and furrows more than 10 cm wide and deep, erosion rills cross the limits of the fields

By extrapolating the results to similar soils and inclinations, he developed outline maps of West Germany on a scale of 1:500 000. However, different soil types may show different erosion features for similar quantities of eroded soil. The map may be looked upon as a qualitative map depicting the extent of rill erosion on agricultural land.

There is also a map of East Germany on a scale of 1:500 000 which is based on field work concerning the extent of visible erosion, vegetation, soil and slope inclination (Flegel 1958). This map depicts the extent of severe and moderate water erosion as well as severe and moderate wind erosion.

16.3 Maps of potential erosion

16.3.1 Poland

Maps which show erosion susceptibility rather than actual erosion damage have been produced in many countries. Such potential erosion maps have been constructed on the basis of different erosion controlling factors.

Potential erosion in Poland was estimated on the basis of soil, topography and climate (Reniger 1951). The map shown in Fig. 67 was compared with sediment yield data by Debski (1957), but the quantitative measurements are too few and too approximate to be reliable. Comparison with the sediment yield map of Poland in Fig. 35 shows that some loess areas with severe potential erosion actually have low sediment yield.

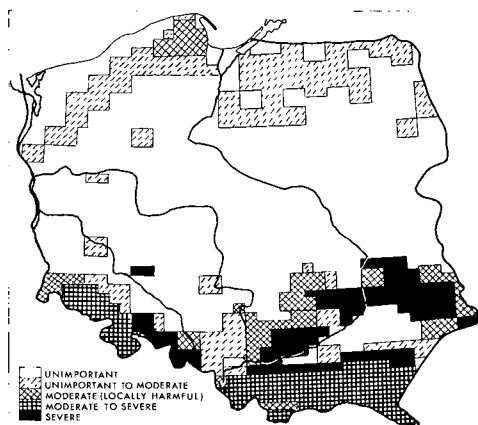


Fig. 67. Potential erosion in Poland (from Reniger 1951).

A comparison can also be made with information provided by Ziennicki *et al.* (1980). They state that soil loss from the slopes and the top of the loess upland is 2–5 mm/year (corresponding to 3000–7500 tons/km²/yr assuming a density of 1.5 tons/m³), but that nearly all the material is trapped and deposited at the bottom of valleys and gullies and that only 2–5 tons/km²/yr is transported away by rivers. This low figure is due to the fact that the gullies are heavily forested.

The same authors distinguish two additional regions of high erosion in Poland: the mountains in the south and the Lake district in the north. They state that in the mountainous region of southern Poland rainfall is greater than in the other parts of Poland and that farming reaches 600–800 m above sea-level. Steep slopes are often ploughed, but the soils are quite resistant to erosion. However, increased channel erosion and landslides have been observed. The amount of material washed out by the rivers is thought to be of the order of 150 tons/km². In the Lake district, steep slopes used as farmland exhibit distinct erosion furrows. The northeastern part of the Lake district has also been mapped by Ugglä *et al.* (1967, referred to in Skrodzki 1972). They dis-

tinguished between slight potential erosion, where slopes do not exceed 6 %, moderate potential erosion in areas with slopes between 6 and 12 % and serious potential erosion, where slopes are > 12 %.

16.3.2 East Germany

The map of the DDR by Kugler (1976) shows potential sheet and rill erosion (Fig. 68). Liability to erosion was estimated by combining the parameters inclination, soil and land use (see below).

Slope gradient is the modal value of the areally dominating gradient class within the respective district. Gradients were measured in the corners of 1 km² squares (Kugler 1969). Climate was not considered.

Potential severe erosion is associated with agricultural land with moderate to steep slopes in the loess region of the southern DDR and in the Lake district of the northern DDR (Fig. 68). Comparison of this potential map with the map by Flegel (1958), which was based on visible erosion features shows that the chief outlines of the patterns coincide.

There is also a map of potential gully erosion in the DDR which was constructed in a similar manner (Kugler 1976). A local map has been constructed in a similar way of the Swan Hill region, Alberta (St.-Onge & Lengelle 1974).

16.3.3 West Germany

Richter (1965) has produced a map of potential erosion in West Germany which is based on the factors of inclination, soil erodibility, land use, precipitation and wind parameters. A detailed description in English of the method of preparation is found in Richter (1980). The map in Fig. 69 is a generalized version from Richter's presentation 1978.

Richter (1978) analysed the relation between different factors and erosion in some typical landscapes of West Germany. His analysis may be summarized as follows: 1. The loess regions suffer severe erosion. Steep slopes in combination with wheat and sugarbeet cultivation on these silty clay soils contribute to the erosion severity. 2. The Alpine Foreland consists of a northern part, where agriculture on loess-covered hills and

LAND USE AND SOIL TEXTURE

Gradient	Cultivation			Forest			Grass		
	I	II	III	I	II	III	I	II	III
> 15°									
8°-15°									
4°-7°									
1°-3°									
0°									

POTENTIAL EROSION

	slight or none
	moderate
	severe

SOIL TEXTURE

- I Somewhat loamy to strongly loamy sand, especially loess soils and soils derived from loess.
- II Loamy to clayey loam.
- III Sand, normally coarse and medium sand without clayey admixtures.

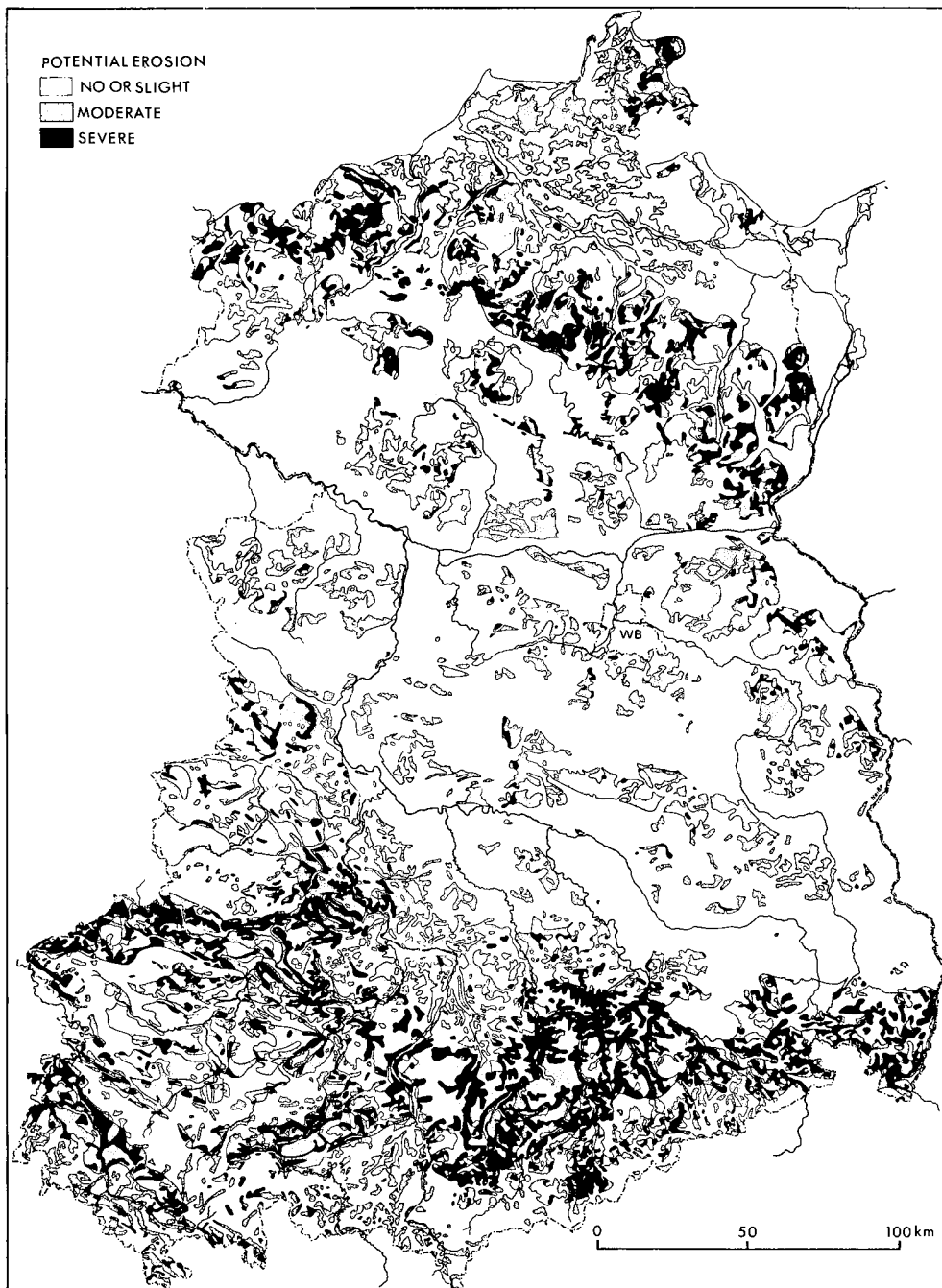


Fig. 68. Potential sheet and rill erosion in East Germany (from Kugler 1976).

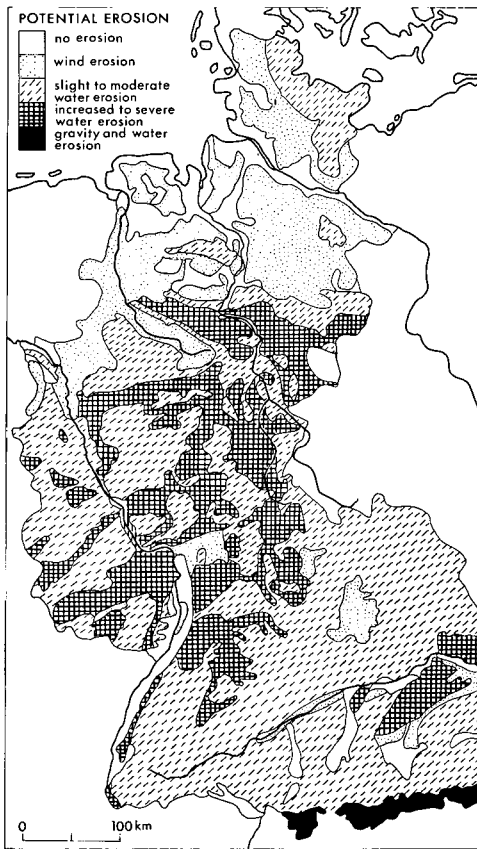


Fig. 69. Potential erosion in West Germany (from Richter 1978).

glaciofluvial plains contributes to moderate and severe erosion, and a southern part, where grass covered moraine undergoes slight to moderate erosion despite high precipitation. 3. The Bavarian Alps are devoted to grazing and forestry. Steep and unstable slopes promote the occurrence of rapid mass movements. It is the conjunction of rapid mass movements and water erosion that is characteristic of erosion in this region.

The areas of severe erosion on this potential erosion map do not always coincide with the highest degrees of erosion shown on Hempel's maps based on visible erosion features (cf. section 16.2).

16.3.4 Morocco

Arnoldus (1977b) prepared a potential erosion map of Morocco, which is based on the USLE, modified for use on a regional scale. This map illustrates maximum potential sheet and rill erosion, since both the C factor and the P factor are assumed to be unity. Three maps were constructed to illustrate the following parameters: the R factor, the K factor and the S factor. The R factor was calculated by using a modification of Fournier's index (chapter 11). The K factor was calculated from the nomograph by Wischmeier *et al.* (1971) for representative soils, and the boundaries of the soils were taken from the FAO/UNESCO Soil Map of the World. The slope factor was also obtained from the FAO/UNESCO Soil Map, where three dominant slope classes and combinations of slope classes are indicated. The three maps were superimposed. Multiplication of the various values resulted in a map of potential soil loss in tons/ha/yr (not reproduced in this study).

Topography is a factor of over-riding importance in this map. The maximum potential erosion does not conform with the distribution of sediment yield in Fig. 55 or with a map of soil degradation based on interpretation of landsat images (Riquier 1977).

16.3.5 North Africa, the Near and Middle East and the Kenyan Rangeland

Maps which cover North East Africa, North West Africa, and the Near and Middle East are issued by FAO *et al.* (1980). Water erosion, wind erosion, salinization and sodication are simultaneously indicated on three map sheets on a scale of 1:5 000 000. The potential rate at which material is removed by water from slopes under present land use conditions is estimated by using the Uni-

versal Soil Loss Equation. The intensity intervals < 1000, 1000–5000, 5000–20,000 and > 20,000 tons/km²/yr are indicated on the maps, which are not reproduced in this study.

The USLE calculations are adapted to the small scale of 1:5 000 000 (FAO *et al.* 1979). A correlation of the EI₃₀ erosivity factor of the USLE with the modification

of Fournier's index, $\sum_{i=1}^{12} p_i^2/P$ by Arnoldus (1977, cf. chapter 11), was used to calculate the erosivity factor from meteorologic data for 615 stations. The soil erodibility factor K was calculated from Wischmeier's *et al.* (1971) nomograph or taken from a rough estimation of the K factor attributed to each soil unit of the FAO/UNESCO World Soil Map.

The slope factor S was based on the slope classes in the FAO/UNESCO World Soil Map and was corrected according to length of slope from contour maps or satellite imagery if this information was available. Parametric values of the cropping-management factor C at this scale of 1:5 000 000 were assigned only to those classes which are shown in the World Atlas of Agriculture. Cropland in areas of very seasonal rain (Sahel and savanna areas) was assigned 0.8 and in more humid areas, without a long dry season, 0.4. In pasture, grassland and woodland or forest, different values were used for different percentages of ground cover, evaluated from satellite imagery.

A similar, more detailed map of potential water erosion, based on the USLE equation, has been constructed for the Kenyan rangeland (Dunne *et al.* 1981). The work in Kenya revealed a tendency toward obtaining exceptionally high values in steep terrain. This might also be true for the FAO *et al.* map (1980).

16.3.6 Czechoslovakia

A map of quantitative soil erosion hazard

in mm/yr was constructed for the western part of Czechoslovakia by Stehlik (1975). The calculation formula included variables of climate, geologic conditions, soil and inclination. Another map of the whole country (Bucko *et al.* 1964) shows the percentage of cultivated land which is susceptible to sheet and rill erosion as well as to gully erosion.

16.4 Evaluation

Maps showing the percentage of eroded topsoil indicate where further erosion may be harmful to agriculture, and may serve as a useful guide for conservation measures. The quantitative rate of erosion measured in tons/km²/yr varies very much within different degrees of erosion. One reason for this is that the original depth of soil varies with different types of soil and with climate. Another reason is that the symbol "moderate erosion" includes areas with 25 % as well as 75 % of the topsoil eroded. This implies a very great variation of the topsoil losses between regions with the same symbol. A third reason is that the extent of land, which is really affected as the symbol indicates, varies between areas. There are also difficulties in combining and comparing areal sheet erosion and linear gully erosion. The reliability of the map depends on the amount of fieldwork performed, the soil scientist's knowledge of erosion processes, and his skill in interpreting aerial photographs. The technique involves many sources of error. It is difficult to find non-eroded soil profiles. Estimation of the original soil depth is also difficult because soil thickness varies with gradient and aspect.

The map of China indicates severe erosion in agreement with the extremely high values on the sediment yield map of the middle and upper reaches of the Yellow river basin.

Land capability maps may be useful in the planning of erosion control measures

and may also be used as preliminary stage in preparing erosion maps. However, they cannot be used directly to obtain a clear picture of erosion.

Maps showing different types of erosion indicate areas where areal and linear erosion is most pronounced. They do not give much information about the intensity of erosion. The quantitative importance of sheet erosion may, for example, be dominant even in a gully-eroded area.

Maps of potential soil erosion are qualitative maps showing erosion hazard. The indicated intensity depends on the factors used and how the factors are combined. The maps can be based on all the main factors that affect erosion as in the map of West Germany in Fig. 69 or on a few selected factors. The combination of different parameters may, however, be a problem, as the parameters do

not have comparable units. The procedure must be accomplished more or less subjectively. Maps of potential erosion, combined with descriptions of erosion conditions, are valuable guides to soil conservation planning.

Potential erosion maps based on the USLE are of varying accuracy depending on the scale of the map and on the reliability of the basic data used. In many areas, a rough potential erosion map, based on the USLE and covering large areas might be useful to organizations which intend to undertake erosion control measures to improve agricultural production. The map of cropland erosion in the USA based on USLE calculations is probably quite accurate as careful subgrouping of slope lengths, gradients, soil and other factors have been employed.

17 Concluding discussion of map analysis

17.1 Introduction

The review and the analysis of existing maps on soil erosion led to several general observations and conclusions. This chapter includes some of them. The sediment yield maps in chapters 14 and 15 are not directly comparable. Different catchment sizes have been used and reservoir surveys or sediment station data have been included. Different sampling techniques and sampling intervals have been employed and the total duration of the sampling period has also varied. Sometimes the sampling technique and calculations are not described. Sediment rating curves may have been constructed for seasons or for an entire year, and different duration curve intervals may have been used in the calculations of sediment yield. Some maps include bed

load but most do not. Different methods have also been used to extrapolate to unsampled areas. The methods used in the construction of qualitative maps also vary. Nevertheless, the survey of maps, both quantitative and qualitative, has been most valuable for a better understanding of global variations in erosion and the need for erosion control.

The rate of erosion results from a combination of several factors. In the preceding chapters, different regions have been looked at individually and the factors responsible for the erosion rates have been discussed in connection with each map. In the following sections, the main outline of the global distribution of erosion will be briefly treated. Three principal erosion factors, viz. climate, relief and soil, in this case the occurrence

of loess, have been chosen for discussion even though many other conditions contribute to differences in erosion rates.

17.2 Sediment yield in different climates

Erosion rate values, taken from the sediment yield maps, have been grouped together in Table 22. As can be observed from the table, Db, Dc and Dd climates usually have values less than 100 tons/km²/yr. Values greater than 100 tons/km²/yr are characteristic of Da, Ca, and Cs climates, especially where there is considerable rainfall seasonality. However, Ohio and western Dakotas have large areas with fairly low sediment yield. BS climates in the USSR, the dry lowlands of Soviet Central Asia and South Africa have values less than 170 tons/km²/yr. BS climates in the United States, Venezuela and the Cameroon usually have values between 200 and 400 tons/km²/yr, while loess soil areas in the Yellow River basin, which is on the borderline between BS and Da climates have values greater than 2000 tons/km²/yr. Soviet Central Asia and Pakistan, which are highland areas surrounded by semi-arid climates, have sediment yield values greater than 500 tons/km²/yr. Lower values (240–350 tons/km²/yr) are found in Fig. 38.

There are too few sediment yield maps from A climates to permit any general conclusions. The sediment yield of the Amazon River basin seems quite high, taking into consideration the flatness, the natural vegetation and the size of the basin. It is comparable to the sediment yield of the Mississippi basin, which has large areas of cultivated land (Fig. 43).

17.3 Erosion in mountainous terrain

Map analysis has indicated high sediment yield and moderate to severe top-

soil erosion in mountainous areas, e.g. in China, Japan, Pakistan, Soviet Central Asia, the Caucasus, Poland, Hungary, Italy, Morocco, Canada and Latin America (Table 22 and Figs 61, 62, 64). The eastern Soviet Union is exceptional because, although mountainous, it does not have high sediment yield. One cause of the high erosion rates in mountains may be the united action of snow avalanches, rapid mass movements and erosion by water (cf. Richter 1978).

In some mountains, glaciers may contribute to the high sediment load in the streams.

High surface runoff because of high rainfall, low evaporation and steep terrain may be another cause contributing to high erosion rate in mountains. In Rumania, sediment yield is correlated with elevation. Sediment yield for small basins is greatest at about 400–800 meters above sea level and decreases for both higher and lower elevations (Diaconu 1974). In Soviet Central Asia, sediment yield decreases with increasing elevation in four regions. This is thought to be due to resistant rocks at higher elevations (Tkacheva 1974 cf. Fig. 38). In Italy, erosion-resistant rocks at high elevations are thought to be the cause of lower sediment yield there. Areas with Cretaceous clay with shells are said to be very susceptible to erosion (Gazzolo & Bassi 1960). The highest sediment yields in Soviet Central Asia are found in mountain ranges from Cretaceous and Tertiary age. These ranges also happen to have high precipitation and many glaciers. In Morocco, the parent material is marl in many areas of the low Rif basins, which yield much sediment (FAO 1975) and in Cyprus, flysch areas are said to experience high erosion intensity (Phanartzis 1971).

The possibility of cultivating mountainous areas depends very much on soil conditions. Farming and grazing in mountainous terrain often cause severe

Table 22 Sediment yields obtained from maps, arranged according to climate

Climate cf. Table I, Fig. 1)	Region	Fig. No.	Sediment yield in tons/km ² /yr		
			Major areas	Minor areas	Comments
Db, Dc, Dd	Poland	35	0.1–11		lowland
				45	loess area
	USSR	46	0–20	19–140	mountains
				20–100	lowland and eastern ranges
	Canada	40	18–88	–	–
		41	7–27	88–>350	Cordillera
			29–>212	–	Cordillera
	USA (small basins)	45	210	–	–
	Wisconsin	51	3.5–35	–	Forests and dairy farming
Da, Ca	UNESCO-map (1974)	–	420;660		Yangtzekiang; Sikiang;
			1190;1400		Red River; Ganga-Bramaputra
	Japan	54	265–1325		most values
				1325–2650	several values
				2650–13,250	seven values
	USA	43	94;49	–	–
	USA (small basins)	45	350–1000	–	–
			>1000	–	agricultural areas (loess)
	Ohio	49	35–125	–	–
				>125	south west
	Western Dakotas	50	35–175	–	–
				175–350	–
				350–1400	Black hills
	Yugoslavia	31A	100–500		mountains
				>500	small areas
				<100	lowland
	Rumania	32A	50–500		mountains
				<50	lowlands
				>500	Transsylvaneian Alps
	Caucasus, USSR	36	100–500		mountains
	Eastern South Africa	56	>500		dense population, sloping land

erosion. Latin America, Hungary, Italy, Morocco and the Fergana Valley of Soviet Central Asia illustrate this clearly. In Poland, rainfall in the mountains is greater than in other parts of Poland. Farming reaches 600–800 m above sea level. Since surface runoff is increased because of cultivation but soil erodibility is low, linear erosion of waterways is increased and landslides are often observed (Ziemnicki *et al.* 1980).

17.4 Erosion in loess areas

Extensive loess areas are found in the Yellow River basin, Soviet Central Asia, the grasslands of the Soviet Union, Hun-

gary, West and East Germany, Poland, Argentina, the Walla-Walla basin in Oregon–Washington, along the Missouri and to the east of the lower Mississippi river.

The loess areas of the Yellow river basin indicate sediment yield values of more than 2000 tons/km²/yr with maximum values of 25,000 tons/km²/yr (Fig. 53). The area is situated in a low mountain area with BS to Da climates with low precipitation and a monsoonal regime. Climatic, topographic, land use and soil factors combine to generate the high erosion rates. Erosion is also severe on loess soils in the wheat growing areas in the transition zone between the humid and the arid Pampas (Fig. 62).

Climate cf. Table 1, Fig. 1)	Region	Fig. No.	Sediment yield in tons/km ² /yr		
			Major areas	Minor areas	Comments
Cs	Italy	33	125-500	500-2500 <125	mountainous, dense population Cretaceous rock slopes Sardinia
	Cyprus	34	100-800 500-1500	-	-
	Yugoslavian coast	31	<100	-	karst areas
	Morocco	55	<500	-	-
	California, USA	45	630	>1000	precipitation >400 mm
	California, USA	44	23;115;1089	-	mountains
	California, USA	43	209	-	-
	South Africa	56	200-500	-	cultivation
BS	USSR	36	0-20		lowland
BS (Da)	Yellow River	53	2000-25,000		loess, mountains, cultivation
H, BS (Ca)	Pakistan	52	1050-3500	<350-1050	mountains
H, BS	Soviet Central Asia	37	500-2500 5-20		mountains dry lowland and dry high plateau
	Soviet Central Asia	38	240-350	-	-
				85	-
BS	USA (small basins)	45	203;217;259		regions 8,9,11
	USA	43	417	-	-
	South Africa	56	15-170		highlands and ranges
	The Northern Cameroon	57	200		low plateau
	Venezuela	63	237;246;300	47;2168	- grazing land
A	South America	63	65 or 88;100		Amazon; Orinocco
	Congo (UNESCO 1974)		37		-
	The Cameroon	57	30-85		mountains
	India (UNESCO 1974)		384;513		Godawari; Mahanadi (reservoir sedimentation not included)

The loess area along the Mississippi river is indicated to have severe erosion on the map of eroded topsoil although the area is lowland (Fig. 58). The map of small basins indicates a sediment yield of nearly 6000 tons/km²/yr from this area (Fig. 45). The Mississippi loess area has higher rainfall erosivity than the Missouri loess area where 1000 tons/km²/yr is indicated on the map of small basins (Fig. 45).

Loess areas have high potential erosion in East and West Germany and in Poland, but the sediment yields from loess areas in Poland are low (Fig. 35). In the grassland area of the Soviet Union, sediment yield is also quite low. However, sediment yields from smaller tributary basins sometimes have tenfold higher values than the zonal rivers in some loess areas (Lisitsyna & Alexandrova 1972).

18 Summarizing remarks and proposals for future research

Erosion intensity is the result of the combined action of the many elements which are discussed in Parts I and II. The elements may be grouped into the major factors of climate, relief, soil, vegetation and man's activities. The influence of the different elements on erosion is very complex. One element may influence many other elements, which in turn, affect erosion and which may affect it in both a positive and a negative way. The maps in Part III have given some indication of the global distribution of erosion which may be explained by the integrated effects of the elements.

A few final remarks on the influence of high intensity rains, precipitation seasonality, relief, soil, natural vegetation, and cultivation are presented below. Special attention is paid to differences in the degree of influence in relation to climate.

1. High intensity rains

B, Dc, Dd, ET and, to some extent, Db and Cb climatic areas receive little precipitation energy (cf. Fig. 5). In these areas, erosion is limited due to low precipitation energy. In arid areas, long distance transport of material is difficult because showers are often local. For example, aridity rather than relief or structure is considered to be the major control on the low rate of sedimentation in the Peru–Chile Trench (Galli-Olivier 1969). However, in the semi-arid areas which have a large amount of high intensity rain, spot erosion may be high (Figs 12, 15, 16) but the transport distance of the eroded material may be short. In humid tropical areas, a large amount of the rainfall occurs as high intensity rain (sections 3.1.3 and 3.1.5).

2. Rainfall seasonality

In climates with distinct wet and dry sea-

sons, vegetation varies seasonally. The reduced vegetative cover at the end of the dry season causes high erosion when the rainy season begins. In monsoonal and tropical wet-and-dry areas, precipitation is often great at the onset of the rains. Another common result of the climatic conditions in wet and dry tropical climates is duricrust formation, which in turn, influences erosion. The second rise of the sediment yield curves in Figs 15 and 16 is caused by data from seasonal climates. High sediment yields in areas with rainfall seasonality are indicated on the maps of Italy (Fig. 33), Cyprus (Fig. 34), Pakistan (Fig. 52), China (Fig. 53), Morocco (Fig. 55), South Africa (Fig. 56) and the northern Cameroon (Fig. 57).

3. Relief

The maps in Part III clearly show that sediment yield and topsoil erosion are especially great in mountainous areas (Figs 31–36, 40, 41, 52, 54–56, 61, 62, 64). In areas where glaciers contribute to the sediment yield, sediment yield may be high (Figs 37, 52). Another source of sediment may be material which comes from rapid mass movements. In cultivated and timber-cutting areas, steep relief may increase the erosion and sediment transport by surface runoff as well as the tendency for rapid mass movements to occur. Topography affects precipitation. High precipitation is therefore another reason for high erosion rates in mountains.

4. Soil

Silt is both highly erodible and easily transported (Figs 8, 9). Loess is a silty material with a high susceptibility to erosion (section 17.4), which means that erosion tends to be severe in loess areas where the groundcover is sparse and the rainfall intensity high.

Gravel and stones provide good protection against water erosion. This is of importance in many alpine and semi-arid areas where the natural vegetation is sparse. Sodium-rich soils are easily dispersed and therefore highly erodible. Soils derived from gypsum in arid areas are found to be extremely erodible (Flaxman 1966).

5. Natural vegetation

Natural vegetation usually provides an effective protection against erosion. The geologic erosion may, however, be comparatively high in semi-arid areas which have a sparse ground cover (Conserv. Found. & FAO 1954) and especially so in those areas which experience high but variable precipitation. Overgrazing in semi-arid areas causes accelerated erosion, which may be disastrous, because the topsoil is often thin and the formation of a new topsoil is slow in this climate. Regions with seasonal snow often exhibit little ground cover for part of the year. However, erosion from meltwater is less effective than erosion from rainfall overland flow (section 3.1.4).

6. Cultivation

Cultivation implies little ground protection for at least part of the year. In most cases, the soil is also deprived of organic matter, the loss of which weakens the soil structure. A decrease in the amount of the clay-humus complex increases the

leaching of nutrients. Intense erosion in cultivated areas is a common phenomenon but, it may be especially pronounced on sloping land in highly erosive climates (Figs 22–24) and in highly erodible loess areas.

Some scattered examples of the occurrence of the above mentioned factors in regions with different climatic and environmental settings are found in Table 23. The numerical values are taken from the maps in chapter 15.

Many problems on soil erosion and sediment transport remain to be solved. More research is needed in order to clarify the influence of the elements on soil loss in different climatic environments. Overland flow is closely related to erosion and transportation. Research on the extension of areas contributing to surface runoff within catchments in different climates and with different topography and land use would therefore elucidate erosion differences as well. Special attention should be paid to the great variations in precipitation within catchments in regions which receive convective rains. This is of importance for the length of sediment transport and also for the development of extrapolation techniques using data from raingauges. In regions which experience frontal precipitation, rainfall is more evenly distributed, but local, topographical conditions can similarly promote local variations.

Table 23 Examples of coincident occurrence of different erosion-promoting factors and sediment yield in selected regions

Region	High intensity rains	Seasonal rains	High relief	High erodibility	Sparse natural veg.	Intense cultivation	Sediment yield tons/km ² /yr
Upper Yellow River	×	×	×	×	×	×	2000–25,000
N. Pakistan (mountains)	×	×	×		×		<350–3500
N. European USSR							0–20
N. Poland						×	0.1–11
Italy	×	×	×		×	×	<125–2400
S. Morocco			×		×		<500–1000

The relationship between chemical solution and throughflow, overland flow, sediment erosion and transport in different climates requires more research. Bremer (1973) considers that intensive deep weathering in the tropical rainforests of Amazonia leads to differential sinking. The effect of piping on gully formation has been reported especially from savanna and semi-arid climates (Parker 1963, Barendregt & Ongley 1977, Stocking 1981, Chakela 1981). Piping is also present in other climates but is often due more to mechanical than chemical processes (cf. Wenner, 1941). Another chemical influence on sediment erosion and transport may be found in semi-arid areas as a result of the greater salinity of the water. Its effects on flocculation and sedimentation in the fields and in the water-ways and reservoirs need more research.

Laboratory experiments on temperature effects on overland flow properties, erosion and transportation for different soils and slope inclinations would be of great value in arriving at a better theoretical understanding of climatic differences in erosion.

The relative importance of different sources of erosion in different climates needs to be studied more. Channel eroded material seems to constitute a significant part of the sediment yield in semi-arid regions of the United States. Furthermore, channel erosion has increased as a result of soil conservation in other parts of the United States (Robinson 1977). Brown (1960, referred to in Smith 1962) declared that sediment yield in the USA is dominated by sheet erosion in agricultural areas with more than 510 mm of precipitation. In forests, pasture and other land with less than 510 mm of precipitation, gully and channel erosion contribute most of the sediment load. Similar observations from other parts of the world would provide useful information. Stream erosion is a con-

spicuous feature in the landscape of northeastern Ghana and elsewhere in the savanna areas (Adu 1972). The quantification of the contribution of mass movements to sediment yield in different environments would also be of interest.

The choice of methods of constructing equations and preparing maps should be based on our present knowledge of erosion and transportation processes. It should also depend on the purpose of the equations and the maps and on the size of the areas. For example, equations can be constructed with the aim of predicting the life of reservoirs in a region or they may form a basis for choosing conservation techniques in cultivated areas. Maps can be constructed in order to give an indication of the variations in sediment yield within a region or on a global scale or to indicate soil conservation needs.

At present, deterministic models of soil loss from smaller areas such as hill slopes have not been extensively developed and still include gross simplifications. Therefore, predictions of erosion are most easily made with reference to empirical equations such as the USLE or the SLEMSA (chapter 11). The order of magnitude of erosion rates can at least be estimated with the help of the USLE even outside the USA but it is necessary to be most careful in evaluating the results of the calculations.

In very small catchments, the MUSLE equation could be used to predict sediment yield (chapter 11). However, the factors should be estimated and weighted only for areas contributing to overland flow. Channel and gully erosion must be added. A statistically developed multiple regression can also be established for small basins in a region and can then be used to obtain estimates of sediment yield in single catchments in the same area. Table 16 and the Appendix should be consulted in the choice of variables. In intermediate basins (100–1500 km²) and large basins

(> 1500 km²) it is necessary to develop multiple regressions. Climatically homogeneous environments should be treated together. It is not advisable to include humid and arid areas in the same equation. Several different variables which describe the five major factors must be tested in order to arrive at a reliable equation. It is especially important to include a variable for the percentage of cultivated land.

A global map of erosion can be constructed by using equations of types 58–63 (Jansen & Painter 1974). However, a variable for the percentage of cultivated land should be included. A soil variable might be determined in a way similar to what has been used to calculate K values in Morocco (section 16.3.4). When erosion is compared on a global scale, basins of comparable size should be used. However, the choice of basin size affects the resulting erosion rate and the range of variation and diversity in erosion rates, which, in turn, affects the required density of the data points. Sediment yields from basins which are too small might be affected by isolated local elements, and very large basins might experience basin sedimentation in reservoirs and channels and on flood plains and in other depositional environments. Therefore, a basin size in the lower range of large basins (2000–5000 km²) would probably be an

appropriate size range for a global map. Regional maps may also form the basis for a global map in areas where they are available. However, adjustments must be made which allow for scale and basin size differences between regional and global maps.

In the construction of regional sediment yield maps, the procedures used by Jovanovic & Vukcevic (1959) in Yugoslavia and Diaconu (1971) in Rumania seem to be acceptable ways of generalizing from the available data (cf. section 15.1). The use of monthly values would probably give considerably more accurate values than the use of mean annual values. An accompanying description of the sources of sediment would be valuable to the user of the map.

Maps of eroded topsoil or potential erosion maps may be suitable as a basis for conservation practices (cf. sections 16.1 and 16.3).

An understanding of erosion processes in different climates is exceptionally important. Much progress has been made and will be made in this field. Man's interference with the subtle ecological balance in nature must be based on a sound knowledge of erosion processes. Present and future knowledge can be used to improve agricultural conditions and to cope with the need of an ever increasing world population for food and fabrics.

APPENDIX

Short review of multiple regressions of sediment yield with a discussion of the variables

Sediment yield is calculated in different units: sediment yield for a long period (S), sediment yield/area (SA), sediment yield/area/year (SAY), sediment load/area/day (SAD), and sediment load/area/month (SAM). Regressions involving sediment concentration (S_c) are also included.

In many cases, a logarithmic transformation has been employed because it is easier to compute and test the significance of linear regressions than non-linear regressions. The equations are given in their original forms. This means that many of the variables are expressed in different units, but their exact meaning is given in the symbol list at the end of the Appendix. Comparable symbols are used for comparable variables.

The equations are arranged in accordance with basin size. The basins are assumed to be small if

they are $< 100 \text{ km}^2$, intermediate if they are $> 100 \text{ km}^2$ but $< 1500 \text{ km}^2$, and large if they are $> 1500 \text{ km}^2$. Equations including intermediate basins are grouped as small basins if the equations also include small basins, and as large basins if they also include large basins. The equations which refer to small basins are divided into climatic types. The equations which refer to large basins are arranged according to author. This way of grouping precludes repetition of comments on the variables and facilitates a comparison of what variables have been used in different climates by the same author, who has used the same test variables.

Short comments have been inserted under the equations but those who wish to follow the discussions of the researchers more in detail should turn to the original works.

A Small drainage basins

A.1 Grasslands in Da and Ca climates

Source	No. of basins	Location	Size km ²	Equation	R ²
Glymph 1954a from Gottschalk&Brune	30	Loess Hills, W.Iowa	0.098-107	$\text{LogS}_1 = 1.0545 \log E_1 + 0.7664 \log 100A_1 + 0.7867 \log AGE_1 + 0.3701 \log (TR)_2 - 2.9127 \dots (26)$	93%
Glymph 1954a from Glymph et al.	36	- " - and E.Nebraska	0.093-7252	$\text{LogSAY}_4 = 1.0078 \log E_1 - 0.1354 \log 100A_1 + 0.6460 \log 10P_1 - 1.4130 \dots (27)$	88%
Glymph 1954a from Maner&Barnes	7	Blackland Prairies, Tex.	1.11-22.56	$\text{LogSAY}_1 = 0.9898 \log E_1 - 0.1407 \log A_1 - 0.2400 \dots (28)$	93%
Stall&Bartelli 1959	20	W.central Ill.(loess)	0.23-668	$SA_1 = 3.9 + 12.2E_3 + 0.25R_1 + 0.74AGE_1 - 1.16D_1 + 26.9 (TR)_3 - 8.21 \log A_4 \dots (29)$	88%
Ackermann&Corinth 1962	12	S.and central Ill.	23-2347	$SAY_3 = \frac{0.053E_3 (TR)_1}{E_1} (TR)_5^{0.25} R_5^{0.47} \dots (30)$ $SAY_3 = 0.00856E_3 (TR)_1 (TR)_5^{0.52} R_5^{0.74} \dots (31)$	97% 95%
Farnham et al. 1966	22	W.Iowa N.W. Mo.	0.223-6.86	$SAY_2/E_2 = 0.494 + 10.255R_4 + 0.072 (TR)_8 - 1.041A_5 - 1.258 (TR)_4 \dots (32)$ $SAY_2/E_2 = 0.494 + 6.310R_4 + 2.934R_6 + 0.071 (TR)_8 - 0.980A_5 - 1.445 (TR)_4 \dots (33)$	76% 79%
Glymph 1954a from Kohler	14	Clarinda, Iowa, Bethany, Mo.	0.01-55	$\log SAY_1 = 3.0858 \log P_1 + 1.8896 \log 100Q_1 + 0.7029 \log C_1 - 0.0563 \log R_3 - 0.0130 \log 1000A_1 - 4.6646 \dots (34)$	

In this group of equations which refer to grasslands in Da and Ca climates, both bed load and suspended load are calculated from reservoir and lake surveys. Trap efficiency variables are included in most cases. The sediment yield is corrected for trap efficiency in equation 27. In 6 out of 7 studies gross erosion (E) is included in the equations. Variables describing climate, vegetation, soil and slope are employed to derive gross erosion. Only area or relief, and in one instance chances of deposition in non-incised channels (D₁) are taken into account as variables of transportation of the eroded material. Thus, these equations consist of gross erosion and sediment delivery ratio variables.

In equation 34, P₁ is a climatic variable which is used to describe precipitation at the time of year when the soil is most exposed. There is also a runoff variable in equation 34 which is more a transportation variable. Area is also included in this equation.

As can be seen from the first three equations sediment yield/area varies with E^{1.0} and with A^{-0.23}, A^{-0.14}, A^{-0.14}. When the area values are inserted into the equations, sediment yield per area is 5, 3 and 1.5 times greater in the smaller areas than in the larger basins in the respective studies.

Two studies use linear arithmetic models (equations 29, 32, 33). It is difficult to evaluate the importance of the variables in Stall & Bartelli's equation (29) as gross erosion is in tons/acre per year, but sediment yield is in tons per acre. Thus, the annual erosion is not multiplied with the number of years of reservoir operation. One explanation might be that sediment yield during the average age of reservoirs is calculated and adjusted to the exact number of accumulation years with the AGE term. Equations 30, 31, 32 and 33 are dimensionally correct.

A.2 BS climate. Middle latitude, semi-arid climate

Source	No.	Location	Km ²	Equation	R ²
Branson&Owen 1970	17	Badger Wash, Colo.	0.049- 1.25	SAY ₅ =40.87R ₆ +0.03C ₃ (35)	86%
Hadley&Schumm 1961	84	Cheyenne R, Wyo., S.D., Nebr.	<7.77	log10SAY ₅ =0.74+0.53log10(TR) ₆ + +0.47logE ₄ -0.68log100A ₁ (36)	71%
- " -	22	- " -	<7.77	log10SAY ₅ =0.20+0.47log10(TR) ₆ + +1.37log100R ₆ -0.69log100A ₁ (37)	77%
- " -	13	Twenty-mile Creek	<7.77	log10SAY ₅ =0.18+0.48log10(TR) ₆ + +1.02 log100R ₆ -0.53 log100A ₁ (38)	85%
Flaxman&Hobba 1955	38	Columbia R.	0.036- 16.77	logSAY ₆ =0.9426+0.4694log(TR) ₇ - -0.7374logAGE ₁ -0.2059logA ₂ + +0.8705 logP ₃ +0.2082logC ₄ (39)	80%

In this climatic group, gross erosion (E) is not calculated. In one equation, 36, there is a subjectively evaluated upland erosion variable, E₄, which is defined with reference to arbitrary scales of infiltration, of vegetation density, and of channel condition. E₄ varies between 1 and 4. Relief/length is as good a variable as E₄. Relief/length is also used in the arithmetically linear equation developed by Branson & Owen (35). Flaxman & Hobba tested average slope in % as a relief variable but it was not statistically significant. This agrees with the findings of Branson & Owen, who stated that there was a very low correlation between weighted mean slope of contour belts and sediment yield/area.

Area is used in two studies and shows a highly significant correlation ($r = 0.89$) in the third study, but is not incorporated in the equation because of the masking effect. A correlation of sediment yield to $A^{-0.2}$ (equation 39) is a normal relationship (cf. Table 5). The circumstance that sediment yield is proportional to $A^{-0.68}$, $A^{-0.69}$ and $A^{-0.53}$ (equations 36-38) can be explained by the fact that the trap efficiency variable is simple reservoir capacity and not related to unit area or inflow.

A climatic variable is only used in one study. However, climatic variables are very difficult to define in semi-arid climates. Precipitation is very erratic. Much of the eroded material is deposited within the drainage basin because showers are short and localized. In a high magnitude (perhaps a 20-50 year) flood event, most of the deposited material will be transported to the basin outlet. However, long records are not available in most cases. Branson & Owen found that there was a poor ($r = 0.34$) positive, non-significant correlation between annual runoff and sediment yield per area in 17 catchments in the Badger Wash basin, where

annual precipitation is about 216 mm.

The average annual precipitation in equation 39 varies from about 203 mm to 762 mm, and much of this falls as snow which accumulates in the mountains until spring. Sudden thaws, frequently associated with warm "Chinook" winds, cause high runoff and severe erosion. In dry areas rain variability between years is great. Because stockpond accumulation is used in equation 39, mean annual values must be employed. In addition to precipitation, the course of the thaw (with or without rainfall) would affect the amount of erosion.

As climate influences natural vegetation in semi-arid regions, cover density will include the effect of both climate and natural vegetation. Land use or vegetation variables are used in two studies. In equation 35 the amount of bare soil is used as a variable. The small basins that equation 35 are based on are climatically homogeneous. Sandy soils have a relatively dense cover, while clayey soils support only sparse vegetation. Therefore, besides being a vegetation variable "the amount of bare soil" reflects soil characteristics.

Apart from the implied intercorrelation of the vegetation variable with soil there are no soil variables in the equations from B climates. Flaxman & Hobba examined % of land with shallow, medium and deep soils, but it was not found to be significant in explaining sediment yield.

In conclusion, it may be said that in the studies which refer to semi-arid climates, relief/length, vegetation cover and area are the variables which are most commonly included in multiple regressions with sediment yield. A precipitation variable is included in one case but soil variables are not included.

A.3 Cs climate - Mediterranean climate

Source	No.	Location	Km ²	Equation	R ²
Scott et al. 1968	8	S. Calif.	8.65- 1100.8	$\log \text{SAY}_5 = -6.48 + 1.70 \log P_2 + 0.70 \log I_1 + 1.61 \log K_1 \dots (40)$	86%
Anderson 1949	8	S. Calif.	11.7- 523	$\log \text{SAY}_5 = 1.041 + 0.866 \log q_1 + 0.370 \log D_2 - 1.236 \log C_5 \dots (41)$	90%
- " -	"	- " -	- " -	$\log \text{SAY}_5 = 2.161 + 0.071 \log A_1 + 1.619 \log I_2 + 0.410 \log P_4 + 0.370 \log D_2 - 1.974 \log C_5 (42)$	

The catchments studied by Anderson are on the southfacing slopes of mountains in southern California. The basins are characterized by steep topography and shallow, coarse-textured soils. Vegetation varies from thin chamise and sage on the lower, drier sites to mixed-chaparral, oak-chaparral and conifers on the higher, wetter sites. Equation 41 is dimensionally correct. The equation includes climatic, vegetation and drainage density variables. There are no soil or relief variables (provided D_2 is not considered as a topographic variable). Peak discharge during the year can be calculated by the equation:

$$\log q_1 = 1.293 + 1.082 \log A_1 + 1.870 \log I_2 + 0.474 \log P_4 - 0.852 \log C_5 \dots (43)$$

This expression is included in equation 41 instead

of $\log q_1$ (cf. equation 42).
 When I re-calculated equation 41, I got higher multiple correlation coefficient for non-logged data, $R=0.98$, than for logged data, $R=0.94$. The plot of residuals was better with non-logged values and standard errors of the coefficients were lower. On statistical grounds a linear model of non-transformed variables would have been preferred.
 Five of the reservoirs used by Anderson were also included among the reservoirs used by Scott, Ritter & Knott (1968), but they used average annual values and not sediment yield from separate years. To obtain a significant result with only eight catchments, three independent variables were considered to be the maximum permissible number, thus only climatic and soil erodibility variables were considered.

A.4 Forest land in Ca, Cb, Cc and Da climates

Source	No.	Location	Km ²	Equation	R ²
Cooper&Snyder 1956	1	White Hollow, E.Tenn.	6.94	$S_2=52.4e^{-0.167AGE_2}P_5^{0.93}P_6^{-0.38}$ $T_1^{-0.78}q_2^{0.89}.....(44)$	73%
Imeson 1971	1	Hodge Beck, E.UK	18.9	$\log Sc_1=1.024\log Q_2+0.365\log q_3+$ $+1.2118.....(45)$	79%
"	1	Catchwater Drain, E.UK	15.4	$\log Sc_1=0.413\log Q_2+0.236P_7+1.154....(46)$	92%
Striffler 1963	20	Tobacco R., Mich.	2.6- 347.8	$\log Sc_2=3.066+0.191\log Q_3+0.135\log EB-$ $-0.003K_2-0.007K_3-0.003K_4+0.001C_6-$ $-0.756QS.....(47)$	52%
"	"	"	"	$\log Sc_2=2.996+0.131\log Q_3+0.154\log EB-$ $-0.767QS-0.002K_5.....(48)$	49%
"	"	"	"	$\log SAD_1=3.831+1.190\log Q_3+0.134\log EB-$ $-0.003K_2-0.007K_3-0.003K_4-0.764QS.(49)$	76%
"	"	"	"	$\log SAD_1=3.729+1.128\log Q_3+0.153\log EB-$ $-0.778QS-0.002K_5.....(50)$	76%
"	"	"	"	$\log SAD_1=3.771+1.138\log Q_3+0.131\log EB-$ $-0.778QS-0.003K_6.....(51)$	76%
Douglas 1973	13	S.E. New South Wales	1.19- 226.6	$\log_{10}SAY_2=2.912+1.538\log P_{11}-$ $-0.2367\log B-2.0791\log_{10}R_{10}-$ $-6.718\log D_4+3.7431\log_{10}R_{11}+$ $+3.1161\log C_9-1.2221\log G_2.....(52)$	80%

Cooper & Snyder (1956) reported on a TVA project. Tree planting was undertaken in a severely eroded area in Tennessee. Since the cover density increases to some final value as an exponential function of time, its effect on the production of sediment was assumed to be a decreasing exponential function of time. Thus, $e^{-0.167 AGE_2}$ was used as an index of vegetational cover in this case (44). In addition to the effect of cover changes, suspended sediment load per storm was found to be a function of peak discharge, temperature 10 days prior to the storm and amount of rain divided by duration of rain. A low temperature yields more sediment. That must mean that winter rains give more sediment load per storm. The reduction in suspended load in 1952 was 93 % of the 1936 value.

In a stream with the quite high sediment yield, for English conditions, of 480 tons/km²/yr (equation 45) concentration increases almost linearly with discharge. In the stream with 8.9 tons/km²/yr in sediment yield (equation 46) concentration increases at a decreasing rate for higher discharges. Antecedent precipitation is of importance in this case.

Striffler (1963) used individual observations of concentration or the instantaneous value of sediment discharge converted to a daily value as the dependent variable (equations 47-51). It is a type of extended sediment rating curve technique for 20 rivers, where differences in the coefficient of $\log Q_3$ between rivers and differences in the constant of the sediment rating curve function are compensat-

ed by variables which describe bank erosion, soil, cultivation and stage. The only relief variable examined was the gradient of main channel but it was not found to be significant; this was also the case with area. No seasonal variables, explaining differences in cover, time since cultivation, temperature, frost and snowmelt and differences in baseflow at different seasons, were examined, nor any variables describing antecedent moisture conditions or rainfall intensity. Bank erosion contributed 26 to 30 percent of the average sediment load.

Equation 52 by Douglas (1973) is included in this group of climate even though there is altitude-determined ET climate within some basins. Seven river basins are < 100 km² and six are intermediate in size, i.e. > 100 km². The best multiple regression for predicting suspended sediment yield in N S Wales (52) contains Fournier's precipitation index (P_{11}). However, element analysis shows that P_{11} is non-operative while mean annual runoff and mean annual precipitation both are operative variables in this region. Relief/length and bifurcation ratio are also operative variables according to element analysis. Nevertheless relief/length is not included in the equation. It was shown in the article that principal component 2 is very much a topographic component but the significance of this component to mean annual suspended sediment yield is not clear since sediment yield is not significantly associated with the first five components but with the sixth (Douglas 1973).

A.5 Equations from more than one climatic environment

Source	No.	Location	Km ²	Equation	R ²
Flaxman 1966	30	9 states in W. USA	0.049- 621.6	logSAY ₁ =1.5945+0.6789I ₃ +0.0190R ₇ - -0.00655K ₇ +0.1552K ₈ (53)	81%
Flaxman 1972	28	10 states in W. USA	0.01- 130	log (SAY ₅ +100)=6.21301- -2.19113log (P ₃ /T ₂ +100) + +0.06034log (R ₇ +100) - -0.01644log (K ₇ +100) + +0.04250log (K ₈ +100) (54)	92%
Douglas 1973	11	N.E. Queensland	5.59- 585.34	log10SAY ₈ =-8.783+3.810logQ ₆ - -1.538logR ₉ +4.822logD ₄ (55)	83%
"	11	"	"	log10SAY ₈ =-8.410+2.704logP ₁₁ + +5.603logB+2.967logD ₄ (56)	82%
"	24	N.E.Queensl. + N S Wales	1.19- 585.34	log10SAY ₈ =1.124logP ₁₁ +1.733logB- -1.107..... (57)	55%

The first equation (53) is from nine western states of the United States (Flaxman 1966). Precipitation in the catchments varies from about 118 mm to 2540 mm/year. Cultivated catchments are excluded as well as basins with substantial amounts of channel and gully erosion. Equation 53 contains one rainfall intensity variable, one slope variable and two soil variables. A trap efficiency variable and runoff were not accepted as significant. Runoff was expected to be indicative of protective ground cover. No real vegetation variable was investigated. The high standard error of estimate of 0.45 log units shows that the equation is not a reliable predictor of sediment yield. It is erratic for high yields and where there are soluble salts >1 % or where average runoff is < 610 mm. Soils, high in salts, derived from marine sediments, specifically gypsum, develop a low density salt "mulch" when dry. The surface layer is highly dispersed and readily movable with rainfall and runoff (Flaxman 1966, p 827).

The second equation (54, Flaxman 1972) has a very low standard error of 0.001 log units and explains as much as 92 % of the variation in sediment accumulation. Even here there is one climatic, one relief and two soil variables. P₃/T₂ is intended to reflect the response of vegetation to climate. The higher the ratio between annual precipitation and annual temperature, the greater the vegetative protection. P/T is adjusted downward when vegetation is disturbed and modified upward for snowmelt.

Trap efficiency and volume-weight of deposited sediment failed to add significantly to the explanation of sediment yield variation. This was also the case with rainfall intensity and runoff variables. Equation 53 is a semi-logarithmic relation while equation 54 is written as a logarithmic relation. Weber, Fogel & Duckstein (1976) criticize Flaxman's choice of the log-log model without statistical examination of a linear model. They point out that in this case (54), adding 100 to the values of the

variables before converting to logarithms results in an approximately linear transformation of the original data, because the values of the variables are small relative to 100:

log (x + 100) ≈ log 100 + $\frac{x}{100}$

I have run the raw data in equation 54, without invoking a log transformation or adding 100 to the values, in a forward stepwise regression program. R² was 0.90 compared to 0.92 in the original article. The residuals had a very similar configuration to that when 100 was added before log transformation.

Equation 55 and 56 (Douglas 1973) are from one region, N.E. Queensland. In some parts tropical cyclonic rains are of great importance, while in most parts the south easterlies play the dominant role. The region is within the Am (Aw), Ca climatic types. Three climatic variables, five relief variables, one vegetational and one geology variable were examined. The superiority of the study lies in the exhaustive statistical analysis. Multiple regression analysis as well as principal component analysis and element analysis were used to select variables.

Equations 55 and 56 seem to be equally satisfactory predictive equations (R² ≈ 88 %). However, the variables in the first equation are significantly associated with principal components 1 and 4 while the variables in the second equation are significantly associated with components 1, 2 and 4. Suspended sediment yield has highest load on component 2 and bifurcation ratio (equation 56) is also associated with component 2. Element analysis also shows that equation 56 is better since P₁₁ is an almost ideal operative variable in this region while Q₆ is classed as inoperative.

When catchments from both Queensland and New South Wales are considered (57), suspended load is most highly correlated with P₁₁ but none of the climatic variables meet the optimum conditions for an operationally significant variable. B, associated with component 2, is operational showing

that transport efficiency is important in affecting sediment yield. Principal component analysis shows that most of the variance in the data set used to develop equation 57, which makes up compo-

nent 1, includes wetness variables, suspended sediment yield and relief/length. Component 2 has bifurcation ratio and ruggedness number as its highest loadings.

B. Large drainage basins

Source	No. of basins	Cli- mate	Size km ²	Equation	R ²
Jansen & Painter 1974	14	A	>5000	$\text{LogSAY}_7 = 4.354 + 1.527\log Q_4 - 0.302\log A_3 + 0.296\log R_8 - 3.417\log T_3 \dots (58)$	94%
	14	B	>5000	$\text{LogSAY}_7 = 12.133 - 0.340\log Q_4 + 1.590\log H_1 + 3.704\log P_8 + 0.936\log T_3 - 3.495\log C_7 \dots (59)$	86%
	25	C	>5000	$\text{LogSAY}_7 = -3.078 + 1.003\log Q_4 + 0.686\log R_8 + 4.287\log T_3 - 5.031\log C_7 \dots (60)$	63%
		C R ₈ ≤ 3		$\text{LogSAY}_7 = 6.976 + 0.444\log Q_4 - 1.674\log A_3 + 3.860\log H_1 - 1.665\log R_8 + 3.425\log T_3 \dots (61)$	76%
		C R ₈ > 3		$\text{LogSAY}_7 = -3.055 - 1.125\log A_3 + 0.585\log H_1 + 1.104\log R_8 + 3.056\log P_8 - 3.053\log C_7 \dots (62)$	95%
	26	D	>5000	$\text{LogSAY}_7 = -5.073 + 0.514\log H_1 + 2.195\log P_8 - 3.706\log C_7 + 1.449\log G_1 \dots (63)$	65%
	79	All cli- mates	5430-6 130 515	$\text{LogSAY}_7 = -2.032 + 0.100\log Q_4 - 0.314\log A_3 + 0.750\log H_1 + 1.104\log P_8 + 0.368\log T_3 - 2.324\log C_7 + 0.786\log G_1 \dots (64)$	58%

During the development of equations 58–64 Jansen & Painter (1974) tested a rough vegetation variable and a rough geology variable. The reason for not using a soil variable was that adequate soil maps were not available. However, even if adequate soil maps were available there would have been great difficulty in constructing a variable that could be calculated easily from maps. As soils are very varied, weighting problems are especially great in large basins. Instead of a soil variable, a rock softness variable was used with values 2, 3, 5 or 6 which was only significant in D climates and in the equation for all climates. Natural vegetation was used as a variable with the values: forest, 4; grass, 3; steppe, 2; desert, 1. The values were not weighted according to area. Even in B climates basins were ascribed the value 4. Man's influence was not accounted for.

From the equations 58, 59, 60, 63 it can be seen that discharge, temperature and vegetation are used in three of four climates, precipitation, slope and altitude in two climates and area and lithology in one. In D climates discharge and temperature are not significant variables, as they are in the three other climates. Area appears again in equation 64 which refers to all climates and in the two equations 61 and 62 when C climate is separated according to relief. But in the latter cases the high coeffi-

cients of logA must be the result of some intercorrelation. The coefficients of logA in equations 58 and 64 are more normal (cf. Table 5).

Equations 58–64 show that sediment yield increases with increasing runoff, altitude, relief, precipitation, temperature and rock softness and decreases with increasing area and vegetation. However, inverse relationships are found with temperature in A climates, with annual total discharge in B climates, and with relief in C climates with low relief (61). One explanation for inverse temperature relationship in A climates is that the density of vegetation will increase with temperature. Another explanation is that rainfall seasonality increases with decreasing temperature. The negative sign accompanying discharge in B climates may be due to the fact that a low annual discharge will comprise a few very high intensity rain storms which produce little vegetation while higher discharge is indicative of more vegetation cover. It is possible that man's activities at low altitudes are sufficient to have changed the sign accompanying relief in C climates with R ≤ 3.

The mean annual suspended sediment yield in A climates was 72 tons/km²/yr, in B climates 169 tons/km²/yr, in C climates 714 tons/km²/yr and in D climates 47 tons/km²/yr (Jansen & Painter 1974).

Source	No.	Climate, veg., region	Size, km ²	Equation	R ²
Lisitsyna & Alexandrova 1972	8	D; forest; Kola-Karelia	100-50 000	$SAY_7 = 0.71Q_5 - 6.95 \dots (65)$	74%
	76	D; forest; N. Russian Plain	100-50 000	$SAY_7 = 0.89Q_5 - 0.41C_8 + 5.07 \dots (66)$	53%
	21	D; forest; Ural	500-20 000	$SAY_7 = 0.56Q_5 - 1.67R_8 + 21D_3 - 1.98 \dots (67)$	67%
	11	D; 50% forest; Carpathians	100-10 000	$SAY_7 = -1.83Q_5 + 125 \dots (68)$	52%
	14	D; steppe; the Don	1500-35 000	$SAY_7 = 13.3Q_5 + 27R_8 - 30.5 \dots (69)$	59%
	7	- " -	<1500	$SAY_7 = 93.8Q_5 + 76.6R_8 - 390 \dots (70)$	88%
	11	D; steppe; the Volga	500-25 000	$SAY_7 = 31.8Q_5 + 43.6R_8 - 79.3 \dots (71)$	94%
	14	D+B; steppe; N. of Caspian Lowland	1500-35 000	$SAY_7 = 5.25Q_5 + 39.4R_8 - 6.0 \dots (72)$	83%
	26	D+B; steppe; Russian Upland	500-35 000	$SAY_7 = 2.91Q_5 + 0.6 \dots (73)$	47%
	16	B; semidesert; Caspian Lowland	500-25 000	$SAY_7 = 8.29Q_5 - 0.17 \dots (74)$	42%
Tkacheva 1974	26	H(B); S. Tien-Shan mountains	140-4650	$SAY_7 = 9.58Q_5 - 0.18H_1 + 0.04A_3 + 496 \dots (75)$	77%
	14	B; Fergana Valley	1000-31 600	$SAY_7 = 21.4Q_5 - 0.13H_1 + 100 \dots (76)$	66%
	9	H(B); S. Tien-Shan foothills	2600-13 500	$SAY_7 = 38.4Q_5 + 0.03A - 120 \dots (77)$	90%

In contrast to equation 63 by Jansen & Painter the equations 65-71 use discharge per km² as a variable in D climates (Lisitsyna & Alexandrova 1972). The only variables examined for equations 65-74 were discharge per area, slope, weighted average drainage basin elevation, density of river system and % lakes and bogs. Variables of % cultivated land, soil and vegetation (except in the form of the discharge variable) were not tested. The equations are intended to measure zonal sediment yield, i.e. sediment yield from drainage basins, both natural and developed (cultivated, artificial flow regulation etc.) typical of a given zone. However, within the regions, cultivated land constituted

0-50 %, 0-30 % and 30-80 % of the basin areas. Therefore, land use is a cause of variation in sediment yield within the regions and should have been tested as a variable.

The lower limits of zonal drainage areas have been established as 100 to 1500 km², depending on regional characteristics. For instance a characteristic feature of the steppe zone is the dense ravine-gully network. The lower limits of the steppe regions vary with dissection of relief and the shape of the longitudinal profile of slopes. Strong gully dissection and convex slope profiles imply a large catchment area.

Source	No. Climate, reg.	Size, km ²	Equation	R ²
Anderson 1954	29 C; W. Oregon	145- 18 855	LogSAY ₉ = -3.721 + 0.116 log A ₁ + 1.673 log q ₄ + + 1.244 log Q ₇ + 0.401 log R ₁₃ + 0.0486 K ₁₀ + + 0.482 K ₁ + 0.0280 C ₁₀ - 0.0036 C ₁₁ + 0.942 C ₁₂ + + 0.0086 C ₁₃ (78)	
	17 Willamette R., W. Oreg.		SAY ₁₀ = -1.639 + 0.240 C ₁₄ + 0.00514 EB ₂ (79)	
Guy 1964	1 D; Scantic R. Conn.	255	LogSc ₃ = 1.551 + 0.143 log Q ₈ + 0.177 log Q ₉ + 0.037 q ₆ - - 0.113 log 10P ₅ (80)	
	1 C; Brandywine Creek, Del.	813	LogSc ₃ = 0.406 + 0.719 log Q ₈ - 0.00150 M ₁ + + 0.00484 T ₄ - 0.124 log Q ₉ + 0.196 q ₆ - - 0.232 log 10P ₅ (81)	
	1 C; Hazel River, Va.	741	LogSc ₃ = 0.642 + 0.513 log Q ₈ - 0.00256 M ₁ + 0.0097 T ₄ - - 0.178 log Q ₉ - 0.421 log 10P ₅ + 0.533 log 100I ₄ ... (82)	
	1 C; Pappahannoch R., Va.	1595	LogSc ₃ = 1.333 + 0.187 log Q ₈ - 0.00211 M ₁ + + 0.00711 T ₄ + 0.204 log 100I ₄ (83)	
	1 C; Rapidan R. Va.	1204	LogSc ₃ = 0.355 + 0.524 log Q ₈ + 0.00970 T ₄ + 0.429 q ₆ - - 0.275 log 10P ₅ (84)	
	1 C; James R. at Buchanan, Va.	5398	LogSc ₃ = 0.960 + 0.532 log Q ₈ - 0.00530 M ₁ + 0.00777 T ₄ - - 0.566 log Q ₉ + 1.019 q ₆ (85)	
	1 C; James R. at Scottsville Va.	11839	LogSc ₃ = 0.211 + 0.560 log Q ₈ - 0.00297 M ₁ + + 0.00286 T ₄ - 0.196 log Q ₉ + 0.770 q ₆ (86)	

One equation that includes variables which cover all the major factors of climate, relief, soil and land use is that developed by Anderson (1954). Even though the basins were large Anderson employed a soil sampling technique which involved that sampling was restricted to some standard conditions of soil forming processes in each of 14 major geologic types. Thus the number of needed soil samples was reduced to feasible limits. As man's activities caused accelerated channel erosion in some of the rivers, there was some discrepancy between measured suspended load and that calculated by equation 78. The Willamette river was the river with the greatest discrepancy. By using the discrepancy of measured sediment yield and sediment yield calculated by equation 78, and then applying equation 79, Anderson (1954) estimated that 24 % of suspended sediment load came from 14,140 km² of forest land, 22 % from 4710 km² of agricultural land and 54 % from the 62.5 km of eroding main channel in the Willamette catchment.

The period of sediment observations is only 1 to 3 years, but as it refers to a maritime forest climate this is not such a great drawback as the necessary observation period will probably be lowest in this type of climate (cf. Lisitsyna & Alexandrova 1972). Average suspended sediment concentration was calculated for only seven discharge classes and multiplied by the volume of water in the respective discharge classes. This method is very rough compared to a combination of the sediment rating curve

and flow duration curve methods, especially with Piest's duration curve intervals (Walling 1977).

Guy (1964) ascertained which climatic factors influenced mean storm sediment concentration in individual rivers in C or D climates in eastern USA. Equations with coefficients which are highly significant and have the least standard error of estimate are tabulated. Time trend, temperature and peakedness index are not log transformed while discharge and rainfall variables are. When different equations for one single stream are compared, the variation in coefficients for a specific variable is largely due to the effect of intercorrelation among the independent variables. Intercorrelation is also the cause of the change of sign and significance of coefficients in exceptional cases.

Net storm surface runoff is the only variable that occurs in all the equations (80-86). In six of the seven equations mean air temperature is a significant variable. Concentration is greatest in seasons with high temperature. This is contrary to conditions in equation 44, where storm load decreases with increasing temperature. Groundwater runoff is thought to be a variable describing antecedent conditions, but it is not a good variable, as it is intercorrelated with air temperature. The significance of the regression coefficients indicate that a peakedness index or peak flow are better measures of storm intensity than mean storm precipitation intensity.

Equation 87 (Cooper & Snyder 1956) shows that

Source	No.	Climate, reg.	Size, km ²	Equation	R ²
Cooper & Snyder 1956	1	C; Chestnec Creek, Tenn.	347	$SAM_1 = -8.70 + 29.88e^{-2.8M_2} + 0.433P_{12} \sqrt[3]{P_{13}} \sin M_3$ (0.833) $P_{12}^2 \sqrt[3]{P_{13}}$ (87)	80%
Mc Pherson 1975	36	D; S. Alberta, Canada	75-1427	$\text{LogSAY}_{11} = 0.5532 \log R_2 - 0.5859 \log A_6 + 0.8332 L_1 + 1.37878$ (88)	25%
Lal et al. 1977	5	C; N. India	1860-57 000	$SAY_{12} = 0.01 (TR)_3^{0.22} Q_4^2$ (89)	
	5	"	"	$SAY_{12} = 155 (TR)_3^{0.22} Q_4^{0.68} A_3^{-0.2}$ (90)	
	5	"	"	$SAY_{12} = 0.774 (TR)_3^{0.58} Q_4^{1.23} R_{12}^{-0.13}$ (91)	

sediment load decreases by an exponential function of time after tree plantation, as the cover density increases.

McPherson (1975) developed a regression equation (88) for intermediate sized streams. Only one stream basin is smaller than 100 km². Equation 88 explains only 25 % of the variation in suspended sediment load (R=0.5). One reason for the low explained variance is incorrect basic data because of the low number of observations used in the construction of sediment rating curves. In five cases sediment rating curves for the whole year were constructed from ten or less observations. This is too few observations for reliable sediment yield estimations. Another reason is that only shape, relief and discharge variables were examined. There were no variables describing land use, al-

though both coniferous forest and grassed prairie are found in the basins, and no bank erosion or soil variables. Besides differences in the natural vegetation there are certainly cultivation differences which probably explain some of the variance in sediment yield. Climatic variables, other than discharge might also have been tested.

Equations 89-91 (Lal, Banarji & Narayanan 1977) refer to north Indian Ca climate (one basin from Aw climate). The introduction of an additional parameter describing area or relief (equations 90, 91) did not improve the results in northern India but when the equations were tested on two rivers in southern India, the new equations with additional parameters gave better results than equation 89.

List of symbols

Sediment yield variables

S = *sediment yield*

- 1) tons
- 2) storm sediment load in tons

SA = *sediment yield/area*

- 1) tons per acre

SAY = *sediment yield/area per year*

- 1) tons/sq mi/year
- 2) inches/year
- 3) tons/acre/year
- 4) tons/sq mi/year, corrected for trap efficiency
- 5) acre ft/sq mi/year
- 6) cubic yards/sq mi/year
- 7) suspended sediment yield in tons/km² per year
- 8) mean annual suspended sediment load in m³/km² (rock density)
- 9) suspended sediment load in 10 tons/sq mi/year
- 10) measured average suspended sediment yield minus sediment yield read from map, adjusted for recent cutover. (Map, based on equation 78)
- 11) suspended sediment load in tons/sq mi/year
- 12) annual sedimentation of reservoir m³/km² per year

Sc = *sediment concentration*

- 1) mg/l
- 2) instantaneous sediment concentration, PPM (individual observations)
- 3) mean storm sediment concentration PPM

SAD = *sediment load/area per day*

- 1) lb/sq mi per day

SAM = *sediment load/area/month*

- 1) suspended sediment load per month in tons/sq mi

E = *gross erosion (gully+sheet)*

- 1) tons/sq mi/year
- 2) inches/year
- 3) tons/acre/year
- 4) upland erosion evaluated by veg.density, soil texture and channel aggradation. E_4 ranges from 1-4

- 3) longterm average annual prec., inches
- 4) 21-day precipitation antecedent to the max. 24-hour prec. of the year, inches
- 5) storm rainfall in inches
- 6) storm rainfall duration in hours
- 7) rain 24-28 hours before sampling
- 8) annual average rainfall, mm
- 11) the ratio of squared precipitation of the month with the greatest rainfall and mean annual precipitation = Fournier's index p_m^2/P
- 12) rainfall for the current month in inches
- 13) rainfall for the previous month in inches

I = *intensity of precipitation*

- 1) probable max. 24-hour prec. in inches
- 2) max. 24-hour prec. of the year, inches
- 3) 5-year frequency, 1-hour duration rainfall in inches
- 4) mean storm precipitation intensity (> 0.05 inch/h) in inches per hour, using hourly data

Q = *discharge or runoff*

- 1) average annual runoff, inches
- 2) m³/s
- 3) instantaneous discharge rate, cfs/sq mi
- 4) annual discharge in 10³m³/km²
- 5) liter/sec/km²
- 6) mean annual runoff, mm
- 7) mean annual discharge cfs/sq mi
- 8) the sum of mean daily discharges per second minus ground water runoff = net surface runoff in cfs-days
- 9) ground water runoff at the beginning of the storm event in cfs

q = *peak discharge*

- 1) max. yearly peak discharge, cfs/sq mi
- 2) peak discharge of total flow in cfs
- 3) flood intensity index
- 4) discharge peakedness (regression of peak discharge on geologic rock types, area receiving rain and relative snow melt frequency)
- 5) peak flow = peak rate of water discharge less the base flow, in cfs
- 6) peakedness index = q_5/Q_8

M = *month*

- 1) the consecutive number of months during the period of record beginning with the first month of record
- 2) time by months (January 1944 = 0.01, February 1944 = 0.02 etc).
- 3) season by months, evaluated by arbitrary setting January = 30°, February = 60° etc.

Climatic variables

P = *precipitation*

- 1) number of rainfall events, average annual number of events equal to or exceeding one inch per day, during the growing season (April to Oct. 15).
- 2) effective precipitation (Langbein et al. 1949) in inches

T = temperature

- 1) average temperature in °F for 10 days prior to the storm
- 2) average annual temperature in degrees Fahrenheit
- 3) average annual temperature, °C
- 4) long term mean air temperature for the given time of year, °F

Relief and similar variables

R = relief

- 1) mean slope of third order streams, feet/mile
- 2) mean land slope from grid squares, degrees
- 3) average slope of catchment, %
- 4) slope of the highest order streams, ft/foot
- 5) rate of channel slope change 1st to 2nd order, ft/mile
- 6) relief/length, ft/foot
- 7) weighted average catchment slope between contours in %
- 8) main channel slope, m/km = %
- 9) relief/length, feet/mile
- 10) relative relief (= lowest to highest point) /length of perimeter
- 11) ruggedness number = drainage density × lowest to highest/5280
- 12) relief/length, %
- 13) average slope of the stream order of one mile average length, ft/mile

H = altitude

- 1) meters above sea level

L = channel length

- 1) main channel length, miles

A = area

- 1) catchment area, sq mi
- 2) catchment area, acres
- 3) catchment area, km²
- 4) mean direct tributary drainage area to reservoir, acres
- 5) ratio of mean direct tributary area to net drainage area, acres/acre
- 6) sediment contributing area = catchment area minus portions upstream of lakes or dams, in sq mi

F = shape

- 1) principle inflow channel length/diameter of circle equal to the catchment area, miles/mile

D = drainage density

- 1) density of nonincized channels, ft/acre
- 2) area of main channel/catchment area, acres/sq mi
- 3) km/km²
- 4) feet/sq mi

B = geometric mean bifurcation ratio

Soil and geology

K = soil

- 1) surface-aggregation ratio
- 2) Roselawn-Emmet-Rubicon Soil, %
- 3) Grayling-Rubicon Soil, %
- 4) Ogemaw-Rubicon-Nester Soil, %
- 5) Moraine+outwash, %
- 6) Emmet-Roselawn Soil in % +K₂+K₃
- 7) percent of soil particles coarser than 1.0 mm (by weight), in the surface 2 inches of the soil profile
- 8) soluble salts in the representative soil sample, %
- 9) soil aggregation variable of the pH of the soil and the amount of clay present in 2 inches surface soil
- 10) silt and clay (<0.05 mm in 6 inches top soil) %

G = geology

- 1) rock hardness. (Paleozoic G=3, Mesozoic G=5, Cainozoic G=6, Quaternary G=2)
- 2) % of catchment area on igneous rock

Land use and vegetation

C = land use or vegetation

- 1) relative erosion under different land use and farming practices (row crop and fallow = 100)
- 3) bare soil, %
- 4) cultivated + range land, acres/sq mi
- 5) cover density of forest litter, %
- 6) cultivated land
- 7) natural vegetation (forest C=4, grass C=3, steppe C=2, desert C=1)
- 8) area covered by lakes and bogs in basin, %
- 9) % of catchment area covered by rain forest
- 10) bare cultivation (row crops, small grains), % of catchment
- 11) other cultivation than bare cultivation, %
- 12) roads, % of catchment
- 13) area cutover in last ten years, %
- 14) cultivation, % (C₁₀+C₁₁)

Other variables

EB = eroding banks

- 1) total length of eroding banks, 100's of feet
- 2) length of eroding main channel banks, feet

QS = rising-falling stage

rising = 1, falling = 2

(TR) = trap efficiency

- 1) %
- 2) capacity/catchment area acft/sq mi
- 3) capacity/inflow, ratio

- 4) reservoir shape factor = ratio of reservoir area at spillway elevation to the area of a circle of the same perimeter
- 5) direct tributary average annual inflow/reservoir surface area $\times 1$ ft, acft/acft
- 6) reservoir capacity, acft
- 7) capacity/catchment, tenths acft/sq mi

- 8) capacity/catchment ratio divided by original max. reservoir depth, acre feet per sq mi feet

AGE = *age in years*

- 1) age of reservoir
- 2) time by water years since tree plantation

References

Abbreviations according to International List of Periodical Title Word Abbreviations. Moreover, AISH = Association Internationale des Sciences Hydrologiques, IAHS = International Association of Hydrological Sciences, IASH = International association of Scientific Hydrology, SIES = Secretariat for International Ecology, Sweden, UNGI = Uppsala Universitet Naturgeografiska Institutionen.

- Ackermann, C. W. & Corinth, R. L., 1962, Empirical equation for reservoir sedimentation. *Int. Assoc. Sci. Hydrol. Publ.* 59.
- Adu, S. V., 1972, Eroded savanna soils of the Navrongo-Bawku area, northern Ghana. *Ghana J. Agric. Sci.* 5(1), p. 3-12.
- Ahmad, N., 1960, Soil erosion by the Indus and its tributaries. *Pak. Geogr. Rev.* 15(2), p. 5-17.
- Ahmad, N. & Breckner, E., 1974, Soil erosion on three Tobago soils. *Trop. Agric. (Trin.)* 51, p. 313-324.
- Ahn, P. M., 1977, Erosion Hazard and Farming Systems in East Africa. In *Soil conservation and management in the humid tropics*. (Ed.) Greenland, D. J. & Lal, R.
- Akallal, M., 1966, L'érosion dans le bassin du S'ra Quergha /Sebou/ Maroc. (Stenciled report).
- Allen, J. R. L., 1974, Reaction, Relaxation and Lag in Natural Sedimentary Systems: General Principles, Examples and Lessons. *Earth-Sci. Rev.* 10(4), p. 263-342.
- Allis, J. A., 1962, Comparison of storm runoff volumes from small single-crop watersheds and from a larger mixed-crop watershed. *Agric. Eng. April* 1962.
- Anderson, C. H. & Wenhardt, A., 1966, Soil erodibility, fall and spring. *Can. J. Soil Sci.* Vol. 46, p. 255-259.
- Anderson, H. W., 1949, Flood Frequencies and Sedimentation from Forest Watersheds. *Trans. Am. Geophys. Union* 30 (40).
- 1954, Suspended sediment discharge as related to streamflow, topography, soil and land use. *Trans. Am. Geophys. Union* 35 (2), p. 268-281.
- 1975, Relative contributions of sediment from source areas and transport processes. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Anderson, H. W. & Wallis, J. R., 1963, Some interpretations of sediment sources and causes. Pacific coast basins in Oregon and California. *US Dep. Agric. Misc. Publ.* 970.
- André, J. E. & Anderson, H. W., 1961, Variation of soil erodibility with geology, geographic zone, elevation and vegetation type in north California wildlands. *J. Geophys. Res.*, Vol. 66, p. 3351-3358.
- Anttila, P. W. & Tobin, R. L., 1978, Fluvial Sediment in Ohio. *US Geol. Surv. Water-Supply Pap.* 2045.
- Arnett, R. R., 1976, Some Pedological Features Affecting the Permeability of Hillside Soils in Caydale, Yorkshire. *Earth Surf. Processes* 1(1).
- Arnoldus, H. M. J., 1974, Soil Erosion. A review of processes assessment techniques, UNESCO/FAO. Manuscript.
- 1977a, Predicting soil losses due to sheet and rill erosion. In *FAO Conservation Guide* 1.
- 1977 b, Methodology used to determine the maximum potential average annual soil loss due to sheet and rill erosion in Morocco. In *FAO Soils Bulletin* 34, p. 39-48.
- Arulanandan, K. & Heinzen, R. T., 1977, Factors influencing erosion in dispersive clays and methods of identification. *Proc. Paris symp. July 1977. IAHS-AISH Publ.* 122.
- Ateshian, K. H., 1976, Comparative costs of erosion and sedimentation control measures. *Proc. Third Fed. Inter-agency Sediment. Conf.* 1976, Denver, Colo.
- Atlas of American Agriculture, 1936. (Prepared under the supervision of O. E. Baker.) Wash.
- Atlas Mira, 1964, *Fisiko-Geografičeskij Atlas Mira*. Moscow.
- Atlas of world water balance, 1974, USSR Committee for the International Hydrological Decade. Hydrometeorological Publishing House, Moscow-Leningrad.
- Australian Water Resources Council, 1970, Sediment Sampling in Australia. *Hydrol. Ser. No. 3*.
- Ayres, Q. C., 1936, *Soil Erosion and Its Control*. New York.
- Babu, R., Tejwani, K. G., Agarwal, M. C. & Bhushan, L. S., 1978, Distribution of erosion index and isoerodent map of India. *Indian J. of Soil Conserv.* 6(1).
- Barendregt, R. W. & Ongley, E. D., 1977, Piping in the Milk River Canyon, southeastern Alberta — a contemporary dryland geomorphic process. *Proc. Paris symp., July 1977, IAHS-AISH Publ.* 122 p. 233-243.
- Barnett, A. P., Carreker, J. R. & Abruna, F., 1972, Soil and nutrient losses in runoff with selected cropping treatments in tropical soils. *Agron. J.* 64, p. 391-395.
- Barnett, A. P. & Rogers, J. S., 1966, Soil Physical Properties Related to Runoff and Erosion from Artificial Rainfall. *Am. Soc. Agric. Eng. Trans.* 9(1), p. 123-125.
- Barry, R. G., 1969, The World Hydrological Cycle. In *Water, Earth and Man*. (Ed.) Chorley, R. J.
- Basu, J. K. & Puranik, N. B., 1951, Land erosion

- with particular reference to agricultural lands in the dry areas of the Bombay state. *Assem. Gen. de Brux. 1951. Assoc. Int. d'Hydrol. Sci. Publ.* 33.
- Beckinsale, R. P., 1969, River Regimes. In *Water, Earth and Man* (Ed.) Chorley, R. J.
- Belaid, R., 1970, Carte de l'érosion de la Tunisie. Sols de Tunisie. *Bull. de la Div. des sols* 2. Tunis.
- Bennett, H. H., 1939, *Soil Conservation*. Mc Graw-Hill, 993 pp.
- Bergsma, E., 1980, Provisional rain-erosivity map of the Netherlands. In *Assessment of erosion*. (Ed.) De Boodt, M. & Gabriels, D.
- 1981, Indices of rain erosivity. *ITC J.* 1981-4, p. 460-484.
- Betson, R. P., 1964, What is watershed runoff? *J. Geophys. Res.* 69(8), p. 1541-1552.
- Beven, K. J. & Kirkby, M. J., 1979, A physically based variable contributing area model of basin hydrology. *Hydrol. sci. Bull.* 24(1)
- Blake, G. J., Cook, A. D. & Greenall, D. H., 1970, The use of principal component factor analysis to establish the uniformity of a hydrological region in Northland, New Zealand. *IASH Publ.* 96.
- Blom, H., 1962, Laboratorieförsök över rinnande vattens rörelsetillstånd. (Stenciled report.)
- Bobrovitskaya, N. N., Bogolubova, I. N. & Tumanovskaya, S. N., 1977, Methods for computation of runoff and sediment yield from slopes used for agricultural needs. *Proc. Paris symp., July 1977, IAHS-AISH Publ.* 122.
- Bogolyubova, I. V. & Karaushev, A. V., 1974, Water erosion and Sediment Discharge. *Sov. Hydrol.: Sel. Pap. No. 3.*
- Bollinne, A. & Hanotiaux, G., 1975, Mesures de la stabilité structurale dans les sols limoneux: étude critique de la méthode Henin. *Bull. Rech. Agr. Genbloux* 10(3), p. 247-258.
- Bollinne, A., Laurant, A. & Bonn, W., 1979, L'érosivité des précipitations à Florennes. Révision de la carte des isohyètes et de la carte d'érosivité de la Belgique. *Bull. Soc. Géogr. Liège* 15, p. 77-99.
- Bowie, A. J., Bolton, G. C. & Spraberry, J. A., 1975, Sediment yields related to characteristics of two adjacent watersheds. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40).*
- Brański, J., 1975, Rozkład wskaźników Zmaccania i transportu rumowiska unoszonego oraz wskaźników denudacji na obszarze Polski.
- 1980, Sediment runoff from the catchment area of the Carpathian rivers. *Acta Hydrobiol.* 22(1). Kraków, p. 89-100
- Branson, F. A. & Owen, J. B., 1970, Plant cover, runoff and sediment yield relationships in Mancos Shale in western Colorado. *Water Resour. Res.* 6, p. 783-790.
- Bremer, H., 1973, Der Formungsmechanismus im tropischen Regenwald Amazoniens. *Z. Geomorph. N. F. Suppl. Bd. 17*, p. 195-222.
- Bridges, E. M., 1970, *World soils*. Cambridge University Press.
- Browning, G. M., 1938, Changes in the Erodibility of Soils Brought about by the Addition of Organic Matter, *Soil Sci. Soc. Am. Proc. Vol. 2.*
- Brune, G., 1948, Rates of sediment production in midwestern United States. *Soil Conserv. Serv. Tech. Publ.* 65, 40 pp.
- Bryan, R. B., 1968, The development, use and efficiency of indices of soil erodibility. *Geoderma* 2.
- 1977, Assessment of soil erodibility: new approaches and directions. In *Erosion. Research techniques, erodibility and sediment delivery*. (Ed.) Toy, T. J.
- 1979, The influence of slope angle on soil entrainment by sheet wash and rainsplash. *Earth Surf. Processes* 4(1).
- Bryan, R. B., Yair, A. & Hodges, W. K., 1978, Factors controlling the initiation of runoff and piping in Dinosaur Provincial Park badlands, Alberta Canada. *Z. Geomorph. NF Suppl. Bd* 29.
- Bucko, S., Holy, J. & Stehlic, O., 1964, Soil erosion in Czechoslovakia. *J. of the Czechoslovak Geogr. Soc. Suppl. for the XXth Int. Geogr. Congr., Lond.* p. 37-45.
- Bue, C. D., 1970, Streamflow from the United States into the Atlantic Ocean during 1931-1960. *U.S. Geol. Surv. Water-Supply Pap.* 1899-I, 36 pp.
- Carson, M. A. & Kirkby, M. J., 1972, *Hillslope form and process*, The Univ. Press, Camb.
- Chakela, Q. K., 1981, Soil Erosion and Reservoir Sedimentation in Lesotho. *UNGI Rapp. Nr 54.* Upps.
- Chepil, W. S., 1953, Factors that influence clod structure and erodibility of soil by wind. II Water stable structure. *Soil Sci.* 76(5), p. 389-399.
- 1955a, Factors that influence clod structure and erodibility of soil by wind. IV Sand, silt and clay. *Soil Sci.* 80(2), p. 155-162.
- 1955b, Factors that influence clod structure and erodibility of soil by wind. V Organic matter at various stages of decomposition. *Soil Sci.* 80(5), p. 413-421.
- Chorley, R. J., 1969, The Drainage Basin as the Fundamental Geomorphic Unit. In *Introduction to fluvial processes* (Ed.) Chorley, R. J., Methuen & Co Ltd, Lond.
- Chorley, R. J. & Kennedy, B. A., 1971, *Physical Geography. A Systems Approach*. PrenticeHall Int. Inc., Lond.
- Colby, B. R., 1963, Fluvial sediments, a summary of source, transportation, deposition and measurement of sediment discharge. *US Geol.*

- Surv. Bull. 1181-A.*
- Coleman, N. L., 1970, Flume Studies of the Sediment Transfer Coefficient. *Water Resour. Res.* 6(6), p. 801-809.
- Common, R., 1966, Slope failure and morphogenetic regions. In *Essays in Geomorphology*. (Ed.) Dury, G. H.
- Conservation Foundation & FAO, 1954, Soil erosion survey of Latin America. *J. Soil Water Conserv.*, 9. I 158-168. II 214-219, 223-229, 237, III 275-280.
- Cook, F. A., 1967, Fluvial processes in the high Arctic. *Geogr. Bull. Vol. 9*, p. 262-268.
- Cooke, R. U. & Doornkamp, J. C., 1974, *Geomorphology in environmental management. An Introduction*. Clarendon Press, Oxf.
- Cooper, A. J. & Snyder, W. M., 1956, Evaluating the effects of landuse changes on sediment load. *Am. Soc. Civ. Eng. Proc. Vol. 82 (HYI) Pap.* 883, 14 pp.
- Corbel, J., 1959, Vitesse de l'érosion. *Z. Geomorph. N. F. 3*, p. 1-28.
- Craddock, G. W. & Pearce, C. K., 1938, Surface runoff and erosion on granitic mountain soils of Idaho as influenced by range cover, soil disturbance, slope and precipitation intensity. *US Dep. Agric. Circ.* 482, 24 pp.
- Crawford, N. H. & Linsley, R. K., 1966, Digital Simulation in Hydrology: Stanford Watershed Model IV. *Tech. Rep. No. 39 Dep. of Civ. Eng., Stanford Univ.*
- Davis, J. C., 1973, *Statistics and Data. Analysis in Geology*. Wiley & Sons, New York.
- Debski, K., 1957, Essai d'estimation de l'érosion du sol en Pologne. *IASH Publ.* 43.
- De Ploey, J., 1972, A quantitative comparison between rainfall erosion capacity in a tropical and a middle-latitude region. *Geogr. Pol.* 23.
- Diaconu, C., 1969, Resultats de l'étude de l'écoulement des alluvions en suspension des rivières de la Roumanie. *Bull. Int. Assoc. Sci. Hydrol.* 14(3), p. 51-89.
- 1971, Probleme ale scurgerii de aluviuni a riurilor României. *Stud. de Hidrol. Vol. 31*.
- 1974, Phénomènes d'érosion des bassins versants reflétés par les transports solides des cours d'eau. *Proc. Paris symp., September 1974. IAHS-AISH Publ.* 113.
- Dickinson, W. T. & Whiteley, H., 1970, Watershed areas contributing to runoff. *IASH Publ.* 96, p. 12-26.
- Diseker, E. G. & Richardson, E. C., 1962, Erosion Rates and Control Methods on Highway Cuts. *Trans. Am. Soc. Agric. Eng.* 5(2), p. 153-155.
- Donà, F., 1962, Considerazioni generali ed aspetti particolari dell'erosione del suolo in valle Aurina. In *L'erosione del suolo in Italia II*, Padova.
- Donigian, A. S. Jr & Crawford, N. H., 1976, Modeling nonpoint pollution from the land surface. *Ecol. Res. Ser. EPA-600/3-76-083*. US Environ. Prot. Agency, Athens, Ga. 30601
- Douglas, I., 1967, Man, vegetation and the sediment yield of rivers. *Nat.* 215, p. 925-928.
- 1968, Erosion in the Sungei Gombak catchment, Selangor, Malaysia. *J. Trop. Geogr. Vol.* 26.
- 1969, Sediment sources and causes in the humid tropics of northeast Queensland, Australia. *Br. Geomorph. Res. Group, Occas. Pap.* 5, p. 27-39.
- 1973, Rates of denudation in selected small catchments in eastern Australia. *Univ. of Hull Occas. Pap. in Geogr.* 21. 128 pp.
- 1975, Erosion rates and climate: geomorphological implications, and lithology, landforms and climate. In *Geomorphology and Climate* (Ed.) Derbyshire, E., Wiley, Chichester.
- Dragoun, F. J., 1962, Rainfall energy as related to sediment yield. *J. Geophys. Res.* 64(4), p. 1495-1505.
- Drozd, N. I., Goretskaya, Z. A. & Shkryabiy, P. A., 1974, Formation of water and sediment discharge during the spring flood on a slope and in a small gully-ravine drainage basin. *Sov. Hydrol.: Sel. Pap. No. 3*
- Dumas, J., 1965, Relation entre l'érodibilité des sols et leurs caractéristiques analytiques. *Cah. ORSTOM, Sér. Pédol.* 3(4), p. 307-333.
- Dunne, T., 1973, *Studies of soil erosion and soil degradation in Kenya*. Mimeo. Univ. of Wash., Seattle.
- 1978, Field studies of hillslope flow processes. In *Hillslope Hydrology* (Ed.) Kirkby, M. J.
- Dunne, T., Aubry, B. & Wahome, E. K., 1981, *Effect of woodfuel harvest on soil erosion in Kenya*.
- Dunne, T. & Black, R. D., 1970, Partial area contributions to storm runoff in a small New England watershed. *Water Resour. Res.* 6(5), p. 1296-1311.
- Dunne, T. & Leopold, L. B., 1978, *Water in environmental planning*. Freeman, San Franc., 818 pp.
- Dunne, T., Moore, T. R. & Taylor, C. H., 1975, Recognition and prediction of runoff producing zones in humid regions. *Hydrol. Sci. Bull.* 20(3), p. 305-327.
- Ekern, P. S., 1953, Problems of raindrop impact erosion. *Agric. Eng.* p. 23-25.
- Elwell, H. A., 1979, Modelling soil losses in Zimbabwe Rhodesia. *Summer Meet. Am. Soc. Agric. Eng. and Can. Soc. Agric. Eng. June 24-27, 1979. Winnipeg, Can. Pap. No.* 79-2051.
- 1981, A Soil Loss Estimation Technique for Southern Africa. In *Soil Conservation. Problems and Prospects*. (Ed.) Morgan, R. P. C., John Wiley & sons.

- Elwell, H. A. & Stocking, M. A., 1973, Rainfall parameters for soil loss estimation in a subtropical climate. *J. Agric. Eng. Res.* 18, p. 169-177.
- Ellison, W. D., 1947, Soil Erosion Studies. Parts I to VII; *Agric. Eng.* 28. p. 145, 197, 245, 297, 349, 402, 442.
- El-Swaify, S. A., 1977, Susceptibilities of Certain Tropical Soils to Erosion by Water. In *Soil conservation and management in the humid tropics*. (Ed.) Greenland, D. J. & Lal, R., Wiley, Chichester.
- Emerson, W. W., 1967, A classification of soil aggregates based on their coherence in water. *Aust. J. Soil Res.* Melb., p. 47.
- Emmett, W. W., 1978, Overland flow. In *Hillslope Hydrology*. (Ed.) Kirkby, M. J., Wiley, Chichester.
- Engman, E. T. & Rogowski, A. S., 1974, A partial area model for storm flow synthesis. *Water Resour. Res.* 10(3), p. 464-472.
- Epstein, E., Grant, W. J. & Struchtmeyer, R. A., 1966, Effects of stones on runoff, erosion and soil moisture. *Soil Sci. Soc. Am. Proc.* 30(5), p. 638-640.
- Eyre, S. R., 1968, *Vegetation and Soils. A World Picture*. Lond.
- FAO, 1960, Soil Erosion by Wind and Measures for its Control on Agricultural Lands. *FAO Agricultural Development Paper 71*.
- 1965, Soil erosion by water. Some measures for its control on cultivated lands. *FAO Agricultural Development Paper No. 81*.
- 1975, Lutte contre l'érosion et conservation des sols Maroc. Analyse des problèmes de l'érosion à l'échelle nationale. *Rapport technique 1. AG:DP/MOR/71/536*.
- 1976a, *Soils Bulletin 30*. Soil conservation for developing countries. (Ed.) Constantinesco, I.
- 1976b, *Conservation Guide 2*. Hydrological techniques for upstream conservation. (Ed.) Kunkle, S. H. & Thames, J. L.
- 1977, *Soils Bulletin 33*.
- FAO, UNEP & UNESCO, 1979, *A Provisional methodology for soil degradation assessment*.
- 1980, *Provisional map of present degradation rate and present state of soil*. Based on an interpretation of major environmental parameters.
- Farnham, C. W., Beer, C. E. & Heinemann, H. G., 1966, Evaluation of factors affecting reservoir sediment deposition. *Int. Assoc. Sci. Hydrol., Publ.* 71, p. 747-758.
- Finch, V. C., Trewartha, G. T., Robinson, A. H. & Hammond, E. H., 1957, *Elements of Geography*. Mc Graw-Hill, New York-Toronto-Lond.
- Flaxman, E. M., 1966, Some variables which influence rates of reservoir sedimentation in western United States. *Int. Assoc. Sci. Hydrol., Publ.* 71, p. 824-838.
- 1972, Predicting sediment yield in Western United States. *Proc. Am. Soc. Civ. Eng., J. Hydraul. Div.* 98 (HY12), p.2073-2085.
- 1975, The use of suspended-sediment load measurements and equations for evaluation of sediment yield in the West. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*
- Flaxman, E. M. & Hobbs, R. L., 1955, Some factors affecting rates of sedimentation in the Columbia river basin. *Trans. Am. Geophys. Union* 36, p. 293-303.
- Flegel, R., 1958, Die Verbreitung der Bodenerosion in der Deutschen Demokratischen Republik. *Bodenkd und Bodenkultur, Heft 6*, Leipzig.
- Fleming, G., 1969, Design curves for suspended load estimation. *Proc. Inst. of Civ. Eng.* 43, p. 1-9.
- 1975, *Computer simulation techniques in hydrology*. Elsevier, New York.
- Fleming, G. & Walker, R. A., 1977, Digital simulation of soil erosion from the land surface. *Int. Assoc. for Hydraul. Res. 17th Kongress Baden Baden 15-19 Aug. 1977*.
- Forest Service & Soil Conservation Service, 1940, Influence of vegetation and watershed treatment on runoff, silting, and stream flow. A progress report of research. *US Dep. Agric. Misc. Publ.* 397. 80 pp.
- Forsling, C. L., 1931, A study of the influence of herbaceous plant cover on surface runoff and soil erosion in relation to grazing on the Wasatch plateau in Utah. *US Dep. Agric. Tech. Bull.* 220, 72 pp.
- Foster, G. R. & Meyer, L. D., 1972, A closed-form soil erosion equation for upland areas. In *Sedimentation. Symposium to honor Professor H. A. Einstein*. Fort Collins.
- 1975, Mathematical simulation of upland erosion by fundamental erosion mechanics. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*, p. 190-207.
- Foster, G. R., Meyer, L. D. & Onstad, C. A., 1977a, An erosion equation derived from basic erosion principles. *Trans. Am. Soc. Agric. Eng.* 20(4), p. 678-682.
- 1977b, A runoff erosivity factor and variable slope length exponents for soil loss estimates. *Trans. Am. Soc. Agric. Eng.* 20(4), p. 683-87.
- Fournier, M. F., 1949, Les facteurs climatiques de l'érosion du sol. *Bull. de l'assoc. de géogr. fr.* No 202-203.
- 1960, *Climat et Érosion*. Presses Universitaires de France, Paris.
- 1967, La recherche en érosion et conservation des sols sur le continent Africain. *Sols Afr.* 12(1), p. 5-53.
- Free, G. R., 1960, Erosion Characteristics of Rainfall. *Agric. Eng.* 41(7), p. 447-449, 455.
- Gagoshidze, M. S. & Natishvili, O. G., 1974, Na-

- ture of Alma-Ata mudflow. *Sov. Hydrol.: Sel. Pap. No. 5*.
- Galli-Olivier, C., 1969, Climate: a primary control of sedimentation in the Peru-Chile trench. *Bull. Geol. Soc. Am.*, Vol. 80, p. 1849-1852.
- Gazzolo, T. & Bassi, G., 1960, Contribution a l'étude du degre d'érosion des sols constituant les bassins versants des cours d'eau italiens. *Assem. Gen. de Helsinki. Assoc. Int. d'Hydrol. Sci. Publ. 53*.
- Geiger, A. F., 1957, Sediment yields from small watersheds in the United States. *AIHS Publ. 43*.
- Glass, L. J. & Smerdon, E. T., 1967, Effect of Rainfall on the Velocity Profile in Shallow-Channel Flow. *Trans. Am. Soc. Agric. Eng. 10*, p. 330-332, 336.
- Glymph, L. M. Jr, 1954a, Studies of sediment yields from watersheds. *Assem. Gén. de Rome, Assoc. Int. Hydrol. Sci. Publ. 36*, p. 178-191.
- 1954b, Water erosion problems and control on non-irrigated agricultural lands. *Trans. Am. Geophys. Union 35(2)*, p. 246-252.
- Góczán, L., 1972, New methods of mapping the water budget of sloping areas. *Geogr. Pol. 23*
- Gong, S. & Xiong, G., 1980, The origin and transport of sediment of the Yellow River. *Int. symp. on river sediment. Beijing*.
- Gorrie, R. M., 1938, Soil Erosion in India. *Nat. Vol. 142, Lond.*, p. 560-561.
- Greenland, D. J., 1977, Soil Structure and Erosion Hazard. In *Soil conservation and management in the humid tropics*. (Ed.) Greenland, D. J., & Lal, R., Wiley, Chichester.
- Grissinger, E. H., 1966, Resistance of Selected Clay Systems to Erosion by Water. *Water Resour. Res. 2(1)*
- 1972, Laboratory studies of the erodibility of cohesive materials. In *Proc. Miss. Water Resour. Conf., Miss. State Univ. Water Resour. Inst.* p. 19-36.
- Guy, H. P., 1964, An analysis of some stormperiod variables affecting stream sediment transport. *US Geol. Surv. Prof. Pap. 462-E*, p. E1-E46.
- Hadley, R. F. & Lusby, G. C., 1967, Runoff and hillslope erosion from a high-intensity thunderstorm. *Water Resour. Res. 3(1)*, p. 139-43.
- Hadley, R. F. & Schumm, S. A., 1961, Sediment Sources and Drainage Basin Characteristics in Upper Cheyenne River Basin. *Geol. Surv. Water-Supply Pap. 1531-B*.
- Hempel, L., 1963, Bodenerosion in Nordwestdeutschland. Erläuterungen zu Karten von Schleswig-Holstein, Hamburg, Niedersachsen, Bremen und Nordrhein-Westfalen. *Forsch. zur dtsh. Landeskdt., Band 144*. Bad Godesberg.
- 1968, Bodenerosion in Süddeutschland. *Forsch. zur dtsh. Landeskdt., Band 179*, Bad Godesberg.
- Henin, S., 1976, Cours de physique du sol T. I. Initiations, documentations, techniques *ORSTOM, Editest (29)*.
- Higginson, F. R., 1974, Soil Erosion and Siltation within the Murray Valley. *J. Soil Conserv. Serv. N S W.*, 30(3).
- Hindall, S. M., 1976, Measurement and prediction of sediment yields in Wisconsin streams. *US Geol. Surv. Madison, Wis., Water Resour. Div., Water-Resour. Invest. WRD-76/012, WRI-54-75*.
- Hjulström, F., 1935, Studies of the morphological activity of rivers as illustrated by the river Fyris. *Bull. Geol. Inst. Upsala, Vol. 25*, Upps.
- 1939, Transportation of detritus by moving water. In *Recent Marine Sediments. A Symposium*. (Ed.) Trash, P. D.
- Holtan, H. N., 1970, Representative and experimental basins as dispersed systems. *IAASH Publ. 96*.
- Holtan, H. N. & Overton, D. E., 1964, Numerical Experiments in Generating and Routing Watershed Runoff. *Trans. Am. Soc. Agric. Eng.*
- Horton, R. E., 1945, Erosional development of streams and their drainage basins; Hydrophysiscal approach to quantitative morphology. *Bull. Geol. Soc. Am 56(1)*, p. 275-370.
- Hudson, N. W., 1958, Erosion Control Research-Progress Report on Experiments at Henderson Research Station 1953-56. *Rhod. Agric. J. 54(4)*, p. 297-323.
- 1971, *Soil Conservation*. Bratsford Ltd, Lond.
- Instituto Nacional de Tecnologia Agropecuaria, *Degradación de los suelos*.
- Ives, N. C., 1951, Soil and water runoff studies in a tropical region. *Turrialba 1*, p. 240-244.
- Imeson, A. C., 1971, Hydrological factors influencing sediment concentration fluctuations in small drainage basins. *Earth Sci. J. 5*, p. 71-78.
- Jacobi, S., 1977, *Sediment load estimates of rivers in Lesotho*. Dep. of Hydrol. and Meteorol. Serv., Minist. of Works.
- Jamieson, D. G. & Amerman, C. R., 1969, Quick return subsurface flow. *J. Hydrol. 8(2)*, p. 122-136.
- Jansen, J. M. L. & Painter, R. B., 1974, Predicting sediment yield from climate and topography. *J. Hydrol. 21*, p. 371-380.
- Jennings, J. N., 1967, Two maps of rainfall intensity in Australia. *Aust. Geogr. 10*, p. 256-62.
- Jiang, D. & Wu, Y., 1980, Sediment yield and utilization. General Report, Theme A(1). *Int. Symp. on River Sediment. Beijing*.
- Joglekar, D. V., 1960, Silting of reservoirs. *J. of the Inst. of Eng., India, Vol. 40*, p. 505-534.
- Joshua, W. D., 1977, Soil erosive power of rainfall in the different climatic zones of Sri Lanka. *Proc. Paris symp., July 1977. IAHS-AISH Publ. 122*.

- Jovanovic, S. & Vukcevic, M., 1957, Suspended regimes on some watercourses in Yugoslavia and analysis of erosion processes. *IASH Publ.* 43, p. 337-359.
- Judson, S. & Ritter, D. F., 1964, Rates of regional denudation in the United States. *J. Geophys. Res.* 69, p. 3395-3401.
- Kadomura, H., 1980, Erosion by human activities in Japan. *Geo J.* 4(2), p. 133-144.
- Kazó, B., 1969, Determination of waterhousehold parameters of sloping soil surfaces measured by the aid of a rainfall simulator. *Research problems in Hungarian Applied Geography*. Budap. p. 111-123.
- Kemper, W. D. & Koch, E. J., 1966, Aggregate stability of soils from western United States and Canada: measurement procedure; correlations with soil constituents. *US Dep. Agric. Tech. Bull. No 1355*, 52 pp.
- Khosla, A. N., 1953, Silting of Reservoirs. *Central Board of Irrigation and Power, India. Publ.* 51, 203 pp.
- Kihlbom, U., 1970, *Flygbildstolkning för jordartsbestämning*. Stockh. 189 pp.
- Kilinc, M. & Richardson, E. V., 1973, Mechanics of soil erosion from overland flow generated by simulated rainfall. *Colo. State Univ. Hydrol. Pap. No 63*. Ft Collins, Colo.
- Kinawy, I. Z., Wafa, T. A., Labib, A. H. & Shenouda, W. K., 1973, Effect of sedimentation in High Aswan dam reservoir. *Comm. Int. des Grands Barrages. Onzième Congrès des Grands Barrages. Madrid 1973*. Q40 R. 7.
- Kirkby, M. J., 1969, Infiltration, Throughflow, and Overland Flow. In *Introduction to fluvial processes* (Ed.) Chorley, R. J.. Methuen & Co Ltd, Lond. p. 85-97.
- 1975, Hydrological Slope Models: The Influence of Climate. In *Geomorphology and Climate*. (Ed.) Derbyshire, E.. Wiley, Chichester, p. 247-267.
- 1978, Implications for sediment transport. In *Hillslope Hydrology*. (Ed.) Kirkby, M. J.. Wiley, Chichester, p. 325-363.
- Kirkby, M. J. & Chorley, R. J., 1967, Throughflow, overland flow and erosion. *Bull. IASH* 12, p. 5-21.
- Kittredge, J., 1948, *Forest Influences*. The Effects of Woody Vegetation on Climate, Water and Soil, with Applications to the Conservation of Water and the Control of Floods and Erosion. McGraw-Hill.
- Knighton, A. D., 1973, Riverbank erosion in relation to streamflow conditions, River Bollin-Dean, Cheshire. *East Midl. Geogr.* 5, p. 416-426.
- Konteatis, C. A. C., 1974, *Dams of Cyprus*. Zavallis Press Ltd., Nicosia-Cyprus.
- Kowal, J. M. & Kassam, A. H., 1977, Energy Load and Instantaneous Intensity of Rainstorms at Samaru, Northern Nigeria. In *Soil conservation and management in the humid tropics*. (Ed.) Greenland, D. J. & Lal, R., Wiley, Chichester.
- Kugler, H., 1969, Die Entwicklung der geomorphologischen Karten des Atlas DDR als Beitrag zur Strukturforchung und Ressourcenerkundung in der Deutschen Demokratischen Republik. *Geogr. Ber.* 52/53, Heft 3/4, Jahrg. 14, Gotha/Leipzig, p. 242-251.
- 1976, Geomorphologische Erkundung und agrarische Landnutzung. *Geo. Ber.* 80, Heft 3, Jahrg. 21. Gotha/Leipzig, p. 190-204.
- Kung, S. & Chiang, T., 1977, Soil erosion and its control in small gully watersheds in the rolling loess area on the middle reaches of the Yellow river. Peking, China.
- Lal, R., 1976, Soil erosion on Alfisols in Western Nigeria: Effects on rainfall characteristics. *Geoderma* 16, p. 389-401.
- 1977, sections 3.1, 4.1, 4.3, 5.1, 6.1, 7.1 in *Soil conservation and management in the humid tropics*. (Ed.) Greenland, D. J. & Lal, R.. Wiley, Chichester.
- Lal, V. B., Banerji, S. & Narayanan, J., 1977, Sediment yield in relation to drainage basin characteristics in some Indian river valley projects. *Proc. Paris Symp., July 1977. IAHS-AISH Publ.* 122.
- Lam, K-C. & Cheng, S-M., 1977, Prediction of Soil Loss by Rainfall Parameters in a Subtropical Climate. *Hydrol. Symp. Brisbane, 28-30 June 1977. Inst. Eng., Canberra, Aust., Natl. Conf. Publ.* 77/5, p. 40-44.
- Lambert, A. O., 1969, A comprehensive rainfall-runoff model for an upland catchment. *J. Inst. Water Eng.* 23, p. 231-238.
- Lane, E. W. & Borland, W. M., 1951, Estimating bed load. *Trans. Am. Geophys. Union* 32(1)
- Langbein, W. B. & Schumm, S. A., 1958, Yield of Sediment in Relation to Mean Annual Precipitation. *Trans. Am. Geophys. Union* 39(6).
- Leopold, L. B., Wolman, M. G. & Miller, J. P., 1964, *Fluvial processes in geomorphology*. W. H. Freeman and Co, San Franc. 522 pp.
- Li, R.-M., Simons, R. K. & Shiao, L. Y., 1977, Mathematical modeling of on-site soil erosion. *Int. Symp. on Urban Hydrol., Hydraul. and Sediment Control*, Univ. of Ky., Lexington, Ky., July 18-21, 1977.
- Lisitsyna, K. N. & Alexandrova, V. I., 1972, Sediment discharge of the rivers of the European USSR. *Sov. Hydrol.: Sel. Pap. No. 2*.
- Livesay, R. H., 1975, Corps of Engineers methods for predicting sediment yields. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Lugo-Lopez, M. A., 1969, Prediction of the erosiveness of Puerto Rican soils on a basis of the

- percentage of silt and clay when aggregated. *J. Agric. Univ. Puerto Rico*, 53, p. 187-190.
- Lundin, L., 1982, Soil moisture and groundwater in till soil and the significance of soil type for runoff. *UNGI Rapp. Nr 56*, 216 pp.
- Lusby, G. C., 1965, Causes of variations in runoff and sediment yield from small drainage basins in Western Colorado. *US Dep. Agric. Misc. Publ. No. 970*, p. 94-98.
- Maner, S. B., 1958, Factors Affecting Sediment Delivery Rates in the Red Hills Physiographic Area. *Trans. Am. Geophys. Union* 39 (4), p. 669-675.
- Mapes, B. E., 1969, Sediment transport by streams in the Walla Walla River basin. *US Geol. Surv. Water-Supply Pap. 1868*, 32 pp.
- McGinnies, W. G., McComb, A. L. & Fletcher, J. E., 1963, Role of Watersheds and Forests in the Arid West. In *Aridity and Man* (Ed.) Hodge, C. & Duisberg, P. C., Horn-Shafer, Baltimore, Md.
- McGuinness, J. L., Harrold, L. L. & Edwards, W. M., 1971, Relation of rainfall energy and stream flow to sediment yield from small and large watersheds. *J. Soil and Water Conserv.* 26, p. 233-235.
- McPherson, H. J., 1975, Sediment yields from intermediate sized stream basins in southern Alberta. *J. Hydrol.* 25, p. 243-257.
- Megahan, W. F., 1975, Sedimentation in relation to logging activities in the mountains of central Idaho. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Melton, M. A., 1957, An analysis of the relations among elements of climate, surface properties and geomorphology. *Tech. Rep. No. 11. Office of Nav. Res.*, Dep. Geol. Columbia Univ. New York. 102 pp.
- Meyer, L. D., Foster, G. R. & Römken, M. J. M., 1975, Source of soil eroded by water from upland slopes. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Meyer, L. D. & Monke, E. J., 1965, Mechanics of soil erosion by rainfall and overland flow. *Trans. Am. Soc. Agric. Eng.* 8, p. 572-577.
- Meyer, L. D. & Wischmeier, W. H., 1969, Mathematical Simulation of the Process of Soil Erosion by Water. *Trans. Am. Soc. Agric. Eng.* 12, p. 754-758, 762.
- Moore, T. R., 1979, Rainfall erosivity in East Africa. *Geogr. Ann.* 61A, 3-4 p. 147-156.
- Morgan, M. A., 1969, Overland flow and man. In *Water, Earth and Man*. (Ed.) Chorley, R. J., Methuen & Co Ltd, Lond.
- Morgan, R. P. C., 1973, The influence of scale in climatic geomorphology: A case study of drainage density in West Malaysia. *Geogr. Ann.* 55 A (2), p. 107-115.
- 1979, *Soil erosion*. Longman.
- 1980, Field studies of Sediment transport by overland flow. *Earth Surf. Processes Vol. 5*, p. 307-316.
- Murray-Rust, H., 1972, Soil erosion and reservoir sedimentation in a grazing area west of Arusha, northern Tanzania. *Geogr. Ann.* 54A, 3-4.
- Musgrave, G. W., 1947, The quantitative evaluation of water erosion - A first approximation. *J. Soil Water Conserv.* 2, p. 133-138.
- Mutchler, C. K. & Young, R. A., 1975, Soil detachment by raindrops. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Negev, M., 1967, A sediment model on a digital computer. *Tech. Rep. 76, Dep. of Civ. Eng., Stanford Univ.*
- 1972, Suspended sediment discharge in western watersheds of Israel. *State of Isr. Hydrol. Serv. Hydrol. Pap. 14*, 73 pp.
- Newton, R. G. & Spurrell, D. J., 1967, A development of multiple regression for the analysis of routine data. *Applied Statistics*, 16, p. 51-64.
- Nie, N. H., Hull, C. H., Jenkins, J. G., Steinbrenner, K. & Bent, D. H., 1975, *Statistical Package for the Social Sciences*. McGraw Hill.
- Nikolayeva, G. M. & Chernogayeva, G. M., 1974, Maps of the elements of the hydrologic budget of Asia. *Sov. Hydrol.: Sel. Pap. No. 2*.
- Nilsson, B., 1971, Sediment Transport in Swedish Rivers. An IHD-Project, *UNGI Rapp. 4*.
- Noble, E. L., 1963, Sediment Reduction through Watershed Rehabilitation. *US Dep. Agric. Misc. Publ. 970, Pap. No. 18*, p. 114-123.
- Ollier, C. D., 1969, *Weathering. Geomorphology text 2*, Edinburgh.
- Olivry, J. C., 1977, Transports solides en suspension au Cameroun. *Proc. Paris. symp., July 1977. IAHS-AISH Publ. 122*.
- Onstad, C. A. & Bowie, A. J., 1977, Basin sediment yield modelling using hydrological variables. *Proc. Paris. symp. July 1977, IAHS-AISH Publ. 122*, p. 191-202.
- Onstad, C. A. & Foster, G. R., 1975, Erosion modelling on a watershed. *Trans. Am. Soc. Agric. Eng.* 18 (2), p. 288-292.
- Pardé, M., 1961, *Sur la puissance des crues en diverses parties du monde*. Zaragoza, 293 pp.
- Parker, G. G., 1963, Piping, a geomorphic agent in landform development in the drylands. *Int. Assoc. Sci. Hydrol. Publ.* 65 p. 103-113.
- Phanartzis, C., 1971, *Cyprus, Water Resources*. UNDP/FAO, Rome.
- Phelps, H. O., 1970, The Friction Coefficient for Shallow Flows over a Simulated Turf Surface. *Water Resour. Res.* 6 (4).
- Piest, R. F., 1963, The role of the large storms as a sediment contributor. *US Dep. Agric., Misc. Publ. 970*.
- Piest, R. F., Kramer, L. A. & Heinemann, H. G., 1975, Sediment movement from loessial water-

- sheds. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Podmore, T. H. & Merva, G. E., 1971, Silt transport by thin film flow. *Trans. Am. Soc. Agric. Eng.* 14 (6), p. 1065-1067, 1072.
- Proceedings of the Third Federal Inter-agency Sedimentation Conference 1976, Denver, Colo., Sediment. Comm., Water Resour. Counc.
- Ramaiah, R. & Sreenivas, G. N., 1975, Rainfall erosion index of Bangalore agro-climatic area. *Mysore J. of Agric. Sci.* 9 (3) p. 448-453.
- Rango, A., 1970, Possible effects of precipitation modification on stream channel geometry and sediment yield. *Water Resour. Res.* 6, p. 1765-1770.
- Rapp, A., 1974, Physiographical research and teaching on developing countries. *Symp. in appl. phys. geogr., UNGI Rapp.* 34, p. 435-446.
- Rapp, A., Axelsson, V., Berry, L. & Murray-Rust, H., 1972, Soil erosion and sediment transport in the Morogoro river catchment, Tanzania. *Geogr. Ann.* 54A, 3-4.
- Rapp, A. & Strömquist, L., 1979, Field experiments on mass movements in the Scandinavian mountains with special reference to Kärkevage, Swedish Lappland. *Stud. Geomorphol. Carpatho-Balcanica, Vol 13*, Kraków.
- Renfro, G. W., 1975, Use of erosion equations and sediment delivery ratios for predicting sediment yield. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Reniger, A., 1951, Soil erosion investigation in Poland. *Assem. Gen. de Brux. 1951, Assoc. Int. d'Hydrol. Sci. Publ.* 33.
- Rice, R. M., 1970, Factor analyses for the interpretation of basin physiography. *IASH Publ.* 96, p. 253-268.
- 1977, Forest management to minimize landslide risk. In *FAO Conservation Guide 1*.
- Richter, G., 1965, Bodenerosion. Schäden und gefährdete Gebiete in der Bundesrepublik Deutschland. *Forsch. zur dtsh. Landeskd., Band 152*, Bad Godesberg.
- 1978, Soil erosion in central Europe. *Pedol.* 28, Ghent, p. 145-160.
- 1980, Soil erosion mapping in Germany and in Czechoslovakia. In *Assessment of erosion*. (Ed.) de Boedt, M & Gabriels, D.
- Riquier, J., 1977, Soil degradation map of Morocco. In *FAO Soils Bulletin 34*.
- Robinson, A. R., 1977, Relationship between soil erosion and sediment delivery. *Proc. Paris symp. July 1977. IAHS-AISH Publ.* 122, p. 159-167.
- Roehl, J. W., 1962, Sediment Source Areas, Delivery Ratios and Influencing Morphological Factors. *Int. Assoc. Sci. Hydrol. Publ.* 59, p. 202-213.
- Roose, E., 1967, Etudes et Travaux. Dix années de mesure de l'érosion et du ruissellement au Sénégal. *Agron. Trop.* 22, p. 128-153.
- 1977, Application of the Universal Soil Loss Equation of Wischmeier and Smith in West Africa. In *Soil Conservation and mangement in the humid tropics*. (Ed.) Greenland, D. J. & Lal, R. Wiley, Chichester.
- Rose, C. W., 1960, Soil detachment caused by rainfall. *Soil Sci.* 89 (1), p. 28-35.
- Rubin, J., 1966, Theory of Rainfall Uptake by Soil Initially Drier than their Field Capacity and its Applications. *Water Resour. Res.* 2 (4).
- Ruxton, B. P., 1967, Slopewash under Mature Primary Rainforest in Northern Papua. In *Landform Studies from Australia and New Guinea*. (Ed.) Jennings, J. N. & Mabbutt, J. A. The Univ. Press, Camb., p. 85-94.
- Sartz, R. S., 1970, Effect of Land Use on the Hydrology of Small Watersheds in South-western Wisconsin. *Symp. of Wellington, IASH Publ.* 96.
- Schattner, I., 1962, The Lower Jordan Valley: a study in the fluviomorphology of an arid region. *Heb. Univ., Jerus., Scr. Hierosolymitana* 11.
- Schumm, S. A., 1965, Quaternary paleohydrology. In *The Quaternary of the United States*. (Ed.) Wright, H. E. & Frey, D. G.
- Schumm, S. A. & Hadley, R. F., 1961 Progress in the application of landform analysis in studies of semi-arid erosion. *Geol. Surv. Circ.* 437. Wash.
- Schumm, S. A. & Lichty, R. W., 1965, Time, space and causality in geomorphology. *Am. J. Sci.*, 263, p. 110-119.
- Schwartz, H. I. & Pullen, R. A., 1966, A guide to the estimation of sediment yield in South Africa. *The Civil Eng. in South Afr.* Dec. 1966.
- Scott, K. M., Ritter, J. R. & Knott, J. M., 1968, Sedimentation in the Piru Creek watershed, southern California. *US Geol. Surv. Water Supply Pap.* 1798 E, 48 pp.
- Sedimentation Engineering, 1975, (Ed.) Vanoni, V. A.. Am. Soc. Civ. Eng., Sediment. Comm. Hydraul. Div.
- SIES Report No. 8, 1975, Can Desertization Be Stopped? (Ed.) Rapp, A., Stockh.
- Silleos, N. G., 1981, The effect of calcium carbonate on soil erodibility in a survey area in northern Greece. *ITC J.* 1981-4, p. 418-434.
- Simons, D. B., 1969, Open Channel Flow.. In *Introduction to fluvial processes* (Ed.) Chorley, R. J., Methuen & Co Ltd, Lond.
- Simons, D. B., Li, R-M. & Ward, T. J., 1977, A simple procedure for estimating on-site soil erosion. *Int. Symp. on Urban Hydrol., Hydraul. and Sediment Control*. Univ. of Ky., Lexington, Ky. July 18-21 1977.

- Simons, D. B., Li, R.-M., Ward, T. J. & Shiao, L. Y., 1979, *Modeling of Water and Sediment Yields from Forest Catchments*. Eng. Res. Center, Colo. State Univ., Fort Collins, Colo. 67 pp.
- Sims, A. L. & Jones, D. M. A., 1973, *Climatology of instantaneous precipitation rates*. Ill. State Water Surv.
- Singh, G., Dayal, R. & Bhola, S. N., 1967, Soil and water loss studies under different vegetative covers on 0.5 and 1 % slope at Kota. *J. Soil Water Conserv. India* 15 (3/4). p. 17-23.
- Skrodzki, M., 1972, Present-day water and wind erosion of soils in N.E. Poland, *Geogr. Pol.* 23.
- Slaymaker, H. O. & McPherson, H. J., 1973, Effects of landuse on sediment production. *Proc. 9th Hydrol. Symp. Edmonton, May 8-9, 1973, Fluvial Processes and Sedimentation*.
- Smerdon, E. T., 1964, Effect of Critical Tractive Forces in Channels with Shallow Flow. *Trans. Am. Soc. Agric. Eng.* 7, p. 287-290.
- Smith, D. D., 1962, Discussion of Paper by A. R. Croft, "Some Sedimentation Phenomena along the Wasatch Mountain Front". *J. Geophys. Res.* 69(4).
- Smith, D. D. & Wischmeier, W. H., 1962, Rainfall erosion. *Adv. in Agron.* 14, p. 109-148.
- Sorozan, R. & Lawrence, J. W., 1974, A study of factors that influence soil aggregate stability. *Proc. 29th Annu. Meet. Aug. 11-14 1974, Soil Conserv. Soc. Am.*
- Stall, J. B., 1962, Soil conservation can reduce reservoir sedimentation. *Public Works Mag.*
- Stall, J. B. & Bartelli, L. J. 1959, Correlation of reservoir sedimentation and watershed factors. *Ill. State Water Surv. Rep. No. 37*.
- Starkel, L., 1972, The modeling of monsoon areas of India as related to catastrophic rainfall. *Geogr. Pol.* 23.
- 1975, The role of extreme (catastrophic) meteorological events in contemporary evolution of slopes. In *Geomorphology and Climate*. (Ed.) Derbyshire, E., Wiley, Chichester, p. 202-246.
- Stafanovits, P., 1964, Talajpusztulás Magyarországon. *Genetikus Talajterhések Ser. 1 No. 7*. OMMI, Budapest.
- Stehlik, O., 1975, Potencialni eroze pudy proudici vodou na uzemi CSR. *Stud. Geogr.* 42, Brno.
- Stichling, W., 1973, Sediment loads in Canadian rivers. *Proc. Ninth Hydrol. Symp. Edmonton May 8-9, 1973, Fluvial Processes and Sedimentation*.
- Stocking, M. A., 1972, Relief analysis and soil erosion in Rhodesia using multi-variate techniques. *Z. Geomorph. N.F.* 16(4), p. 432-443.
- 1981, A model of piping in soils. *Proc. Third Meet. Int. Geogr. Union Comm. on Field Exp. in Geomorph.*, Kyoto, Aug. 1980.
- Stocking, M. A. & Elwell, H. A., 1973, Soil erosion hazard in Rhodesia. *Rhod. Agric. J.* 70(4).
- Stoddard, D. R., 1969, World Erosion and Sedimentation. In *Water, Earth and Man* (Ed.) Chorley, R. J., Methuen and Co Ltd, Lond.
- St-Onge, D. & Lengellé, J. 1974, Erosion susceptibility maps - Swan Hills, Alberta. *Open file 240, Geol. Surv. Can., Ottawa*.
- Strakhov, N. M., 1967, *Principles of Lithogenesis Vol. 1*, Oliver & Boyd, Edinburgh and Lond.
- Strand, R. I., 1975, Bureau of Reclamation procedures for predicting sediment yield. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*.
- Stratham, I., 1977, *Earth surface sediment transport*. Clarendon, Oxf., 184 pp.
- Striffler, W. D., 1963, Suspended Sediment Concentrations in the Michigan Trout Stream as Related to Watershed Characteristics. *US Dep. Agric. Misc. Publ.* 970. p. 144-151.
- Strömquist, L. & Johansson, D., 1978, Soil erosion and sediment transport in the Mtera Reservoir Region. *Z. Geomorph. N.F. Suppl. Bd. 29*, p. 43-51.
- Sundborg, Å., 1956, The river Klarälven. A study of fluvial processes: *Medd. från Uppsala Univ. Geogr. Inst. Ser. A, No. 115*.
- 1978, River Processes in Different Climates. *Geol. Mag. Sep. Copy. Vol. 87, No. 3* (825).
- Sundborg, Å. & Norrman, J., 1963, Göta älv. *Sver. Geol. Unders. Ser. Ca, Nr 43*.
- Taylor, B. D. & Vanoni, V. A., 1972, Temperature effects in hightransport, flat-bed flows. *J. of the Hydraul. Div. Proc. of the Am. Soc. Civ. Eng.* 98(HY 12), p. 2191-2206.
- Teakle, L. J. H., 1939, Notes on soil erosion in the northern agricultural areas. *J. Agric. West. Aust.*, 16.
- Temple, P. H., 1972, Measurements of runoff and soil erosion at an erosion plot scale with particular reference to Tanzania. *Geogr. Ann.* 54 A 3-4.
- Temple, P. H. & Rapp, A., 1972, Landslides in the Mgeta area, western Uluguru mountains, Tanzania. *Geogr. Ann.* 54 A 3-4.
- Temple, P. H. & Sundborg, Å., 1972, The Rufiji River, Tanzania, hydrology and sediment transport. *Geogr. Ann.* 54 A 3-4.
- Thornthwaite, C. W., Sharpe, C. F. S. & Dosch, E. F., 1942, Climate and accelerated erosion in the arid and semi-arid Southwest, with special reference to the Palacca Wash drainage basin, Ariz. *US Dep. Agric. Tech. Bull.* 808, 134 pp.
- Tienseemuang, S. & Ponsana, P., 1977, Hydrologic Properties of Soils in Relation to Runoff and Water Control. In *Soil conservation and management in the humid tropics*. (Ed.) Greenland, D. J. & Lal, R., Wiley, Chichester.
- Tinney, R. E., 1962, The process of channel degradation. *J. Geophys. Res.* 67(4), p. 1465-1480.
- Tkacheva, L. G., 1974, Suspended Sediment Dis-

- charge in the Rivers of Soviet Central Asia. *Sov. Hydrol.: Sel. Pap. No. 3.*
- Trimble, S. W., 1974, *Man-induced soil erosion on the southern piedmont 1700-1970*. Soil Conserv. Soc. Am.
- 1975, A volumetric estimate of man-induced soil erosion on the southern Piedmont Plateau. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40).*
- UNESCO, 1972, Status and trends of research in hydrology. *Studies and reports in hydrology 10.*
- UNESCO & AIHS, 1974, *Gross Sediment Transport into the Oceans*. First preliminary edition with preface and introduction in English and French.
- United Nation, ECAFE, 1951, *Proc. Regional Technical Conference on Flood Control in Asia and the Far East.*
- Ursic, S. J. & Dendy, F. E., 1963, Sediment yields from small watersheds under various land uses and forest covers. *US Dep. Agric. Misc. Publ. 970.*
- US Army, 1958, Atlas of Mountain Glaciers in the Northern Hemisphere, *US Army, Environ. Res. Div., Tech. Rep. EP-92.*
- US Dep. Agric., Agric. Res. Serv., 1975, Control of Water Pollution from Cropland. Vol. I. A manual for guideline development. *USDA-ARS-H-5-1.*
- US Dep. Agric., Soil Conserv. Serv. 1948, *Generalized map of soil erosion in the United States.*
- 1975, Procedure for Computing Sheet and Rill Erosion on Project Areas. *Techn. Release No. 51, Geol.*
- 1977, Cropland Erosion. In *The Second Nat. Water Assess., Water Resour. Council, Wash. D.C.*
- US Dep. Army, 1963, *Technical Manual 5-330*. Planning, site selection, and design of roads, airfields, and heliports in the theater of operations.
- Walker, P. H., 1980, Differentiation of sediments in overland flow. In *Int. Symp. on River Sediment. Beijing.*
- Walling, D. E., 1977, Limitations of the rating curve technique for estimating suspended sediment loads, with particular reference to British rivers. *Proc. Paris symp., July 1977. IAHS-AISH Publ. 122*, p. 34-48.
- Vanoni, V. A., 1962, Discussion of Paper by Frank J. Dragoun, "Rainfall energy as related to sediment yield". *J. Geophys. Res. 67(4).*
- Vanoni, V. A., Brooks, N. H. & Kennedy, J. F., 1961, Lecture notes on sediment transportation and channel stability. W. M. Keck Laboratory of Hydraulics and Water Resources, California Institute of Technology, Pasadena, Calif. *Rep. No. KH-R-1.*
- van Rensburg, H. J., 1955, Runoff and soil erosion tests, Mpwapwa, central Tanganyika. *East. Afr. Agric. J., 20(4)*, p. 228-231.
- Varshney, R. S., 1970, Regional sedimentation curves for India. *Indian J. Power and River Val. Dev. 20(9)*, p. 375-378, 364.
- Weber, J. E., Fogel, M. M. & Duckstein, L., 1976, The use of multiple regression models in predicting sediment yield. *Water Resour. Bull., Am. Water Resour. Assoc. 12(1)*, p. 1-17.
- Wigham, J. M. & Stolte, W. J., 1973, Sediment yield estimates from an index of potential sediment production. *Proc. ninth Hydrol. Symp., Edmonton May 8-9, 1973. Fluvial Processes and Sedimentation.*
- Wilkinson, G. E., 1975, Rainfall characteristics and soil erosion in the rainforest area of Western Nigeria. *Exp. Agric. 11(4)*, p. 247-255.
- Williams, A. R. & Morgan, R. P. C., 1976, Geomorphological mapping applied to soil erosion evaluation. *J. Soil Water Conserv. USA. 31(4)*, p. 164-168.
- Williams, J. R., 1974, Predicting sediment yield frequency for rural basins to determine man's effect on long-term sedimentation. *Proc. Paris. symp., September 1974. IAHS-AISH Publ. No. 113*, p. 105-108.
- 1975, Sediment-yield prediction with universal equation using runoff energy factor. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40)*
- 1977, Sediment delivery ratios determined with sediment and runoff models. *Proc. Paris symp., July 1977. IAHS-AISH Publ. 122.*
- Williams, J. R. & Berndt, H. D., 1972, Sediment yield computed with universal equation. *Proc. Am. Soc. Civ. Eng., J. Hydraul. Div. 98 (HY 12)*, p. 2087-2098.
- 1976, Sediment yield prediction based on watershed hydrology. *1976 Winter Meet. Am. Soc. Agric. Eng., Chicago. Pap. No. 76-2535.*
- Williams, M. A. J., 1969, Prediction of rainsplash erosion in the seasonally wet tropics. *Nat. 222*, p. 763-764.
- Wilson, L., 1973a, Relationships between geomorphic processes and modern climates as a method in paleoclimatology. In *Climatic Geomorphology*. (Ed.) Derbyshire, E., p. 269-284.
- 1973b, Variations in mean annual sediment yield as a function of mean annual precipitation. *Am. J. Sci. 273 B*, p. 335-349.
- 1977, Sediment yield as a function of climate in United States rivers. *Proc. Paris symp., July 1977, IAHS-AISH Publ. 122.*
- Virgo, K. J. & Munro, R. N., 1978, Soil and erosion features of the Central Plateau region of Tigray, Ethiopia. *Geoderma 20(2)*, p. 131-158.
- Visher, S. S., 1945, Climatic maps of geologic interest. *Bull. Geol. Soc. Am. 56*, (2).
- Wischmeier, W. H., 1962, Rainfall erosion potential. *Agric. Eng. Vol. 43*, p. 212-215, 225.

- 1974, New developments in estimating water erosion. *Proc. 29th Annu. Meet. Aug. 11-14 1974, Soil Conserv. Soc. Am.*
- 1975, Estimating the soil loss equation's cover and management factor for undisturbed areas. *US Dep. Agric., Agric. Res. Serv. (ARS-S-40).*
- 1976, Use and misuse of the universal soil loss equation. *J. Soil and Water Conserv. 31(1).*
- Wischmeier, W. H., Johnson, L. B. & Cross, B. V., 1971, A Soil Erodibility Nomograph for Farmland and Construction Sites. *J. Soil Water Conserv.*, 26, p. 189-193.
- Wischmeier, W. H. & Smith, D. D., 1958, Rainfall Energy and Its Relationship to Soil Loss. *Trans. Am. Geophys. Union 39(2).*
- 1965, Predicting Rainfall-Erosion Losses from Cropland East of the Rocky Mountains-guide for selection of practices for soil and water conservation. *US Dep. Agric., Agric. Res. Serv. Agric. Handb. No. 282, Wash. D.C. 47 pp.*
- Wischmeier, W. H., Smith, D. D. & Uhland, R. E., 1958, Evaluation of Factors in the Soil-Loss Equation. *Agric. Eng. 39(8)*, p. 458-462, 474.
- Wolman, M. G., 1967, A cycle of sedimentation and erosion in urban river channels. *Geogr. Ann. 49A. Stockh.*
- Wood, P. A., 1977, Controls of variation in suspended sediment concentration in the River Rother, West Sussex, England. *Sedimentol. 24*, p. 437-445.
- Wood, W. F. & Snell, J. B., 1960, A quantitative system for classifying landforms. *US Army Techn. Rep. EP-124, Natick. Mass.*
- Woolhiser, D. A., Hanson, C. L. & Kuhlman, A. R., 1972, Overland flow and rangeland watersheds. *IASH. Publ. 97*, p. 23-39.
- Yair, A., Sharon, D. & Lavee, H., 1978, An instrumented watershed for the study of partial area contribution of runoff in the arid zone. *Z. Geomorph. N. F. Suppl. Bd 29.*
- Yamamoto, T. & Anderson, H. W., 1973, Splash Erosion Related to Soil Erodibility Indexes and Other Forest Soil Properties in Hawaii. *Water Resour. Res. 9(2).*
- Yevjevich, V., 1972, *Probability and Statistics in Hydrology.* Water Resour. Publ., Fort Collins, Colo.
- Young, A., 1972, *Slopes. Geomorphology text 3.* Oliver & Boyd, Edinburgh, 288 pp.
- Young, K. K., 1976, Erosion potential of soils. *Proc. third Fed. Inter-agency Sediment. Conf. 1976, Denver, Colo.*
- Young, R. A. & Wiersma, J. L., 1973, The Role of Rainfall Impact in Soil Detachment and Transport. *Water Resour. Res. 9(6).*
- Ziemnicki, S. & Repelewska-Pekalowa, J., 1980, The intensity of erosion in Poland and the criterion of determination of erosion regions. In *Assessment of erosion* (Ed.) De Boodt, M & Gabriels, D.
- Zingg, A. W., 1940, Degree and Length of Land Slope as it Affects Soil Loss in Runoff. *Agric. Eng. 21*, p. 59.
- Zumer-Linder M., 1975, Environmental changes in the Kalahari in relation to the expansion of livestock grazing. *SIES Rep. No. 8. Stockh.*
- Özyuvaci, N., 1971, Hydrologic characteristics of the Arnavutköy creek watershed as influenced by some plant-soil-water relations.
- Ågren, A., 1981, Ett oväntat fenomen vid multipel regressionsanalys. (Stenciled report).

