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Erosion, Debris Flows and Environment in Mountain Regions

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Proceedings of the International Symposium held at Chengdu, China, 5-9 July 1992. The symposium was organized by the International Commission on Continental Erosion of the International Association of Hydrological Sciences and the Institute of Mountain Disasters and Environment, Academia Sinica. Co-sponsors were the International Union of Forestry Research Organizations, UNESCO, and the Chinese National Foundation of Natural Sciences.
Preface

Over the past decade, the International Commission on Continental Erosion (ICCE) of IAHS has organized and co-sponsored a number of successful international symposia dealing with a variety of topics in the field of erosion and sediment yield. Many of these symposia have focussed on themes which had attracted attention at that time, but one topic has provided the basis for what has become an almost regular series of gatherings. This is the theme of steepland erosion, and more particularly the problems of steepland environments within the Pacific Rim. The International Symposium on Erosion, Debris Flows and Environment in Mountain Regions, which is to be held in Chengdu, China, in July 1992 is the sixth meeting in this informal series which was initiated with the 1981 Symposium on Erosion and Sediment Transport in Pacific Rim Steeplands (IAHS Publication no. 132), held in Christchurch, New Zealand. The meeting in New Zealand was followed by the Symposium on the Effects of Forest Land Use on Erosion and Slope Stability held in Honolulu, Hawaii, in 1984; the Symposium on Erosion, Debris Flow and Disaster Prevention, which took place in Tsukuba, Japan, 1985; the Symposium on Erosion and Sedimentation in the Pacific Rim (IAHS Publication no. 165), held in Corvallis, Oregon, USA, in 1987; and the Symposium on Research Needs and Applications to Reduce Erosion and Sedimentation in Tropical Steeplands (IAHS Publication no. 192), which took place in Suva, Fiji, in 1990.

The Chengdu Symposium, to which this proceedings volume is devoted, again focusses on steepland and mountain environments, but particular emphasis is given to debris flows and to the general problem of environmental degradation in mountain areas, which is now attracting increasing concern. These themes are particularly appropriate for a meeting in Chengdu, in view of the outstanding reputation for research on debris flows and mountain hazards established by the Chengdu Institute of Mountain Disasters and Environment of Academia Sinica, which generously offered to host the Symposium. Although most of the papers deal with studies undertaken within steepland areas bordering the Pacific, there are also contributions reporting the results of investigations in many other mountain areas of the world, including, Greenland, Germany, Poland, Austria, Czechoslovakia, Spain, Italy, Greece, Yugoslavia, several states of the former USSR, and Brazil. The 55 papers provide information from 25 different countries. More familiar
topics such as the measurement and prediction of soil erosion, the
dynamics of debris flows and related phenomena, and slope protection
measures, are joined by papers that deal with the wider field of
environmental degradation and include discussions of vegetation
succession, soil degradation and land restoration.

The broad perspective on erosion problems in mountain areas
provided by the papers in this volume must be seen as a valuable
feature. It reflects the multidisciplinary background of the contributors,
which is in turn a response to the involvement of several organizations
and agencies in organizing and sponsoring the Symposium. The
International Commission on Continental Erosion (ICCE) and the
International Association of Hydrological Sciences (IAHS) have been
joined by UNESCO, the International Union of Forestry Research
Organizations, Academia Sinica and the National Foundation of Natural
Sciences of China, in staging what should prove an extremely interesting
and important symposium that continues the valuable tradition first
established by the meeting in Christchurch, New Zealand, in 1981.

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1 Erosion and Sediment Transport
Threshold of sediment deposition in medium stream power flow

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Abstract The latest results of a research programme on the beginning of sediment deposition in flowing water, and therefore about the maximum bed load capacity, confirm that stream power per unit of bed surface area is the best hydraulic variable to predict deposit formation, and that there are three types of bed load transport with different energy consumption laws. Two simple, equally reliable, criteria are proposed to identify the beginning of deposition, and a single expression for maximum bed load transport capacity is proposed, valid when solid loads and bed lining have the same grain size. These conclusions can be applied to flows in rough fixed bed canals transporting coarse sand with a Froude number ranging from 0.55 to 1.25, and maximum stream power of 0.3 kg m⁻¹ s⁻¹.

NOTATION

\( d \) mean diameter of sediment [L]
\( d_i \) diameter of transported sediment [L]
\( d_f \) diameter of bed material [L]
\( g \) gravitational acceleration [L T⁻²]
\( g_v \) unit solid discharge in volume [L² T⁻¹]
\( k \) ratio between weight of wet material gathered in the slot and the same after drying
\( N \) number of grains deposited on bed
\( q \) unit flow [L² T⁻¹]
\( C_u \) uniformity coefficient of sands
\( F \) Froude number
\( I \) channel slope
\( V \) mean flow velocity [L T⁻¹]
\( X \) dimensionless solid load factor
\( Y \) dimensionless flow factor
\( Z \) dimensionless sediment factor
\( \gamma \) specific weight of fluid [M L⁻² T⁻²]
\( \eta \) rate of deposition
\( \lambda \) length of deposition
\( \mu \) dynamic viscosity [M L⁻¹ T⁻¹]
\( \nu \) kinematic viscosity [L² T⁻¹]
**INTRODUCTION**

In order to understand sediment routing in rivers and mountain streams, and, more generally, to establish sediment budgets in river basins, it is essential to know hydraulic conditions at which the sediment transported by free flow begins to deposit.

When these critical conditions are met, solid particles transported by a flow are deposited and rest on the river bed; this means maximum load capacity has been reached. Bed deposit formation is therefore a safe indication that the bed load transport capacity of flow has been exhausted.

From this standpoint the phenomenon of deposition is no longer seen as a by-product of transport, as is usually the case. It now appears as a distinct phenomenon whose study allows the solution of chronic uncertainties raised by the lack of precision of classic bed load formulae.

This communication presents the latest results of a research programme based on this concept of the phenomenon of deposition and the importance of using independent variables (especially power) in studying river mechanics phenomena (Bordas, 1973). The results were obtained in experiments performed in a rough fixed bed flume for subcritical and supercritical steady uniform flows \(0.55 < F < 1.25\) transporting coarse sand \(0.6 < d < 2\) mm (Silvestrini, 1991).

**BACKGROUND**

The study of the effect of transport capacity on sediment deposition is a less traditional line of research, although it began in 1914 with the studies of Gilbert (1914). In sanitary engineering and hydro-transport this approach is frequently used, mainly because it identifies the conditions under which sediment deposition occurs in canals or pipes (Craven & Ambrose, 1953; Novak & Nalluri, 1978). Pedroli (1963) investigated deposition in smooth fixed bed canals in order to assess bed load transport at gauging stations built in mountain streams.

In 1974 the Institute of Hydraulic Research of the Federal University of Rio Grande do Sul (IPH/UFRGS) began research on this topic. Several series of experiments were carried out in fixed bed (Bordas et al., 1988) and mobile bed (Borges, 1987) flumes. Sands with a mean grain size of 0.77, 1.22 and 1.98 mm were injected at variable rates (up to a maximum of 700 g min\(^{-1}\)) into flumes up to 1 m wide with slopes ranging from 0.004 to 0.010, which
FUNDAMENTALS

If streamflow is defined by an independent variable as unit flow ($q$) we have:
\[ g_v = \phi_1(q, g, I, \mu, \rho, \rho_s, d_i, d_f) \]  

(symbols are as defined in the Notation) for the general case in which the diameter of the injected or transported sediment is different from the diameter of the sediment which constitutes the fixed bed of the stream.

According to normal procedures of dimensional analysis and gathering the three dimensionless terms \( \beta = \rho^{-1} \rho_s; \ I \) and \( q \nu^{-1} \), so as to describe the power flow, the generic expression of transport is obtained:

\[ \frac{g_v}{\nu} = \phi_2 \left( \frac{\gamma q I}{\gamma_s \nu}, \frac{d_i g^{1/3}}{\nu^{2/3}}, \frac{d_f g^{1/3}}{\nu^{2/3}} \right) \]

If streamflow is defined by dependent variables such as shear stress, mean velocity or unit power per unit weight, the above dimensionless flow factor will be substituted by

\[ \frac{\tau g^{1/3}}{(\gamma_s \nu)^{2/3}}, \frac{V}{(\nu g)^{2/3}}, \text{ or } \frac{VI}{(\nu g)^{2/3}} \]

Simplifying the symbols,

\[ X = \phi_2(Y, Z_i, Z_f) \]

In this new formula \( X \) is the dimensionless factor for solid discharge, \( Y \) the dimensionless factor for flow, \( Z_i \) the dimensionless factor which describes the transported sediment and \( Z_f \) expresses bottom roughness. In this communication the results will be presented with reference to the situation where \( d_i = d_f \). Therefore the final relation must be of the following type:

\[ X = \phi_3(Y, Z) \]

**FACILITIES AND EXPERIMENTAL PROCEDURES**

The experiments were performed in a flume 1 m wide and 26 m long, with maximum discharge of 35 l s\(^{-1}\). The flume slope was moulded using levelled transverse profiles. Three slopes were used: 0.006, 0.008 and 0.010. The flume bed was lined with fine, carefully smoothed cement and painted. The sand used for the final lining was applied immediately after the bottom was painted in order to make the sand stick to it. Later, paint was applied lightly with a spray gun to consolidate the lining fixation and provide colour contrast which made it easier to identify the deposits.

The same sands with a density of 2.63 were used both for bottom lining and for the solid load: all of them had a uniform grain size (uniformity
coefficient $C_u = 1.85$) with mean diameters of 0.77, 1.22 and 1.98 mm respectively. The particles were injected in the central part of the flume in a reach with uniform regimen by means of a device which provided an uniform supply to a 70 cm wide strip, at a rate ranging from 300 to 5000 g min$^{-1}$.

Sixty-five critical initial deposition events occurred in a total of 434 experiments performed, during which the Froude number varied from 0.58 to 1.24.

The methodology is basically the same as that used in previous studies. For a given solid discharge a liquid flow which will ensure transport of the injected material is initially discharged into the flume. This flow is then slowly reduced until a continuous deposition with the thickness of the injected sediment diameter is obtained. To help determine the deposit, the flume bed has a transverse retaining slot located a few metres downstream from the sediment injection section, which collects the particles that were not deposited during an experiment.

**IDENTIFICATION OF THE DEPOSITION THRESHOLD**

This task is the most demanding aspect of the research. Two criteria were used to identify the beginning of deposition. The first is basically experimental and attempts to determine the critical flow which generates continuous, uniform deposition with a thickness equal to the diameter of the injected sediment. This criterion is based mainly on the observation of the deposition process. Although this criterion had been easy to use in previous studies, it became a problem at higher powers due to greater irregularity of the deposits.

The second criterion is an effort to eliminate the degree of subjectiveness associated with the first. It is based on the concept of the rate of deposition originally developed by Costa (1974) and improved in this study. Considering the ratio between the area covered by sediment deposited with the thickness of a single grain and the total flume bottom area we have:

$$\eta = \frac{N \pi d^2}{4 B \lambda} \times 100$$

The number of grains deposited during time $\Delta t$ in the bottom reach with length $\lambda$ of the injection section is given by:

$$N = \frac{(PMAS - k PMTM)/\gamma_s}{\pi d^3/6}$$

In this expression, $PMAS$ is the weight of material injected during period $\Delta t$, $PMTM$ is the weight of wet material found downstream from the deposition zone during $\Delta t$ and collected in the slot, and $k$ is the wetness of the sediment collected to the slot, previously determined by weighing the material collected.
in the retention slot before and after drying. Transposing equation (6) to equation (5), the relationship:

\[ \eta = \frac{2.143 (PMAS - kPMTM)}{\gamma s \lambda d} \]  

is obtained, which allows the calculation of the deposition rate after measuring \( PMTM \). The difference between equations (5) and (7) is that expression (7) can be applied to any predetermined value of \( \lambda \) and \( \Delta t \) chosen arbitrarily, thus limiting errors in assessing \( \lambda \) which result from irregular deposition. The procedure used to determine critical flow is simple: the curves \( \eta = \phi(q) \) (see Fig. 2) are traced and the critical flow corresponding to \( \eta = 1 \) is determined. This value is used to calculate the power to be compared to that of the experiment in which deposit formation was observed. The pair of values thus obtained is plotted in Fig. 3.

The coincidence between the criteria is almost perfect although there is a slight tendency for the analytical criterion to indicate initial deposition flows higher than those found according to the visual criterion.

The similarity between results obtained using both criteria allowed the validation of those which were previously obtained using the visual criterion, and consequently increased the data base with the experiments performed by other authors (Almeida, 1980; Costa, 1974; Garcia, 1983).

![Fig. 2](image-url) Identification of the beginning of deposition with analytical criterion.
Threshold of sediment deposition in medium stream power flow

Fig. 3 Comparison of the analytical and visual criterion for the beginning of deposition.

RELIABILITY OF STREAM POWER TO EXPRESS DEPOSIT FORMATION

Experimental ratios were established linking the four dimensionless factors which include the main hydraulic variables

\[ Y_1 = \frac{\gamma q L}{\gamma_s v}, \quad Y_2 = \frac{\tau_s g^{1/3}}{(\gamma_s v)^{2/3}}, \quad Y_3 = \frac{V}{(\nu g)^{2/3}}, \quad Y_4 = \frac{VI}{(\nu g)^{2/3}} \]

with a transport capacity expressed by \( X = q_v v^{1} \).

The fitting of experimental points provided the following results:

(a) \( d = 0.77 \) mm

\[ Y_1 = 4.94 + 1.46 X \quad (R^2 = 99.1\%) \quad (8.1) \]
\[ Y_2 = 0.757 + 0.0391 X \quad (R^2 = 96.6\%) \quad (8.2) \]
\[ Y_3 = 13.4 + 0.386 X \quad (R^2 = 93.4\%) \quad (8.3) \]
\[ Y_4 = 0.106 + 0.00316 X \quad (R^2 = 48.2\%) \quad (8.4) \]

(b) \( d = 1.22 \) mm

\[ Y_1 = 7.14 + 1.57 X \quad (R^2 = 95.4\%) \quad (8.5) \]
\[ Y_2 = 0.914 + 0.039 X \quad (R^2 = 94.9\%) \quad (8.6) \]
\[ Y_3 = 13.1 + 0.405X \quad (R^2 = 93.1\%) \quad (8.7) \]
\[ Y_4 = 0.0985 + 0.0037X \quad (R^2 = 53.0\%) \quad (8.8) \]

(c) \[ d = 1.98 \text{ mm} \]
\[ Y_1 = 15.9 + 1.89X \quad (R^2 = 98.0\%) \quad (8.9) \]
\[ Y_2 = 1.18 + 0.0447X \quad (R^2 = 96.6\%) \quad (8.10) \]
\[ Y_3 = 17.6 + 0.356X \quad (R^2 = 93.8\%) \quad (8.11) \]
\[ Y_4 = 0.127 + 0.0038X \quad (R^2 = 56.2\%) \quad (8.12) \]

In all three cases it is seen that the hydraulic unit stream power per bottom surface unit gives the best fit and also the steepest slope of regression lines. In this way the best functional representation of the experimental points is obtained. These results confirm what was found in previous studies regarding the choice of unit stream power as the best parameter to represent the hydraulic variables and express deposit formation.

**TYPES OF BED LOAD TRANSPORT**

The data obtained in experiments with \( d_i = d_f \) for values of \( 10 < X < 100 \) were plotted in Fig. 1 to verify the different types of bed load transport indicated in previous studies. The result is seen in Fig. 4, which shows that:

(a) the results of present and previous experiments of Costa, Almeida and Garcia agree, proving that the experiments can be repeated, and validating the criteria used for beginning of deposition;
(b) as before, the power consumption due to transport presents three different behaviours, but these do not coincide fully with the previously identified types;

(c) for $X < 1.5$, unit stream power required for transport does not depend on the volume of transported sediment. In this zone, called isolated grain transport, the only variable which governs the phenomenon is particle size (or weight);

(d) for $1.5 < X < 10$, power demand increases with the volume of material to be transported. The fact that the three curves obtained are almost parallel implies that power demand growth is related mainly to the increased volume of material to be transported. In this zone, called "bulk transport", the main intervening factors are therefore the weight of each particle and the total weight of solid load. A preliminary attempt to express the phenomenon as a function of particle size showed that the power consumption law in this zone might take the form:

$$YZ^{-0.73} = \phi(X)$$

and the limits of $X$ between which the equation might be applied vary according to grain size. The limit value 10 is merely indicative: actually, it varies between 5 and 10 according to grain size; and

(e) for $X > 10$ power demand growths are greater than those of the previous zone, but differ according to particle size. They are distinguished by the fact that demand increases faster for smaller grain sizes than for the larger ones. The three curves, therefore, tend to converge and obey the trend which could be perceived in Fig. 1 for flow over a smooth fixed bed (Pedroli, 1963). These facts suggest a new energy dissipation mechanism, probably related to collision among particles during transport, and that a fully established transition zone exists between the "bulk" and "mass" transport zones shown by the research performed by Pedroli.

**MAXIMUM TRANSPORT CAPACITY**

Figure 4 shows that the transition from one type of transport to another occurs more smoothly than expected from previous research. This fact, together with the obvious affinity between the three curves drawn in Fig. 4, makes it acceptable to look for a single expression of maximum transport capacity instead of several different ones, each one appropriate to one bed load type, as was attempted previously.

An attempt to find a single expression, valid when solid load and bottom lining have the same grain size ($d_i = d_j$), was made using the dimensionless term $(X/Z)$ used previously (Bordas et al., 1988). Little is known about it, so far, beyond the fact that it is related to the number of grains transported. The
Fig. 5 Beginning of deposition on a rough fixed bed.

results are shown in Fig. 5: all experimental points surveyed are grouped on a single curve which obeys the equation below (valid for $XZ^{-1} < 3$):

$$YZ^{-1.091} = 1.3404(X/Z)^{0.537}$$

(10)

with $R^2 = 92.7\%$ which corresponds to the following expression of maximum bed load transport capacity:

$$X = 0.676Y^{1.862}Z^{-1.032}$$

(11)

CONCLUSION

The conclusions of previous studies on the relationship between the beginning of deposition and maximum transport capacity of a flow have been confirmed, broadened and improved. The stream power per unit bottom surface is the most appropriate hydraulic variable to predict the beginning of deposit formation. Three types of bed load transport exist, each with its own energy consumption law. The methodology used in determining the beginning of deposition has been improved. Two criteria, visual and analytical, have been defined and validated. Since they are equally reliable, new experiments can be carried out by a person who does not have much previous experience, which was not possible before, since it was difficult to know precisely when deposition would begin. Finally a relation has been established (equation (10)), which allows the calculation of either the minimum power required to transport a given solid load, or the maximum transport capacity of a given flow for a rough fixed bed and solid load with $X/Z$ values below 3. One of the requirements to generalize
these results would be to extend the research to flows with unit stream power greater than 0.3 kg m\(^{-1}\) s\(^{-1}\).

**Acknowledgements** The authors thank the Brazilian National Council of Science and Technology (CNPq) and the Graduate Course in Water Resources and Sanitary Engineering of UFRGS for supporting this study.

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Erosion resistance of cohesive sediments in turbulent flow

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Abstract A series of experiments on erosion resistance of cohesive soil were conducted in a straight flume and in an annular rotating fibreglass channel-ring system (Cao & Fang, 1991). The research confirmed that erosion resistance of cohesive sediments was controlled strongly by the bed shear stress, the consolidation degree, the free pore ratio and the experimental flow volume of bed material for a specific engineering problem. Equations relating erosion rate to the bed shear stress or free pore ratio of bed are derived from the results of experiments. For practical needs, a simple relation between dry density of bed and critical boundary shear stress is recommended.

INTRODUCTION

Erosion of cohesive sediments is common in highways, railways, irrigation channels, navigation rivers and reservoirs. Erosion of cohesive river banks and migration of meandering cohesive channels have received attention in recent years. Forces acting on cohesive sediments are quite complex and relate not only to particle size but also to mineral composition and environment. The erosion behaviour of cohesive sediments in a turbulent flow field plays a dominant role in engineering problems.

Compared with noncohesive sediment, cohesive particles have a large specific surface area which is defined as the surface of the particles per unit weight. The specific surface area of quartz particles with a diameter of 1 mm is 0.0023 m² g⁻¹, but the specific surface area of montmorillonite can be as large as 750 m² g⁻¹. The latter is 3.26 × 10⁵ times the former. Cohesive soil is mainly composed of kaolinite, illite and montmorillonite. The most cohesive particles are slices or needles, so the magnitude of electrochemical cohesive forces are several orders larger than the gravitational forces. The cohesive forces may be 10⁵ times as large as the gravitational forces for cohesive particles with a diameter of 0.001 mm. The rate process theory may be useful for understanding the erosion resistance of cohesive soil. The similarity of erosion at constant stress to creep at constant stress was recognized and the methodology to determine experimental action energies $E_a$ and flow volume $V$ was developed by Kelly & Gularte (1981). In this study, the flow volumes were determined by least squares regression analyses of the data from each run. They are $0.427 \times 10^{13}$ cm³ and $0.158 \times 10^{13}$ cm³ for two types of bed
respectively. The results show that the flow volumes of this study are close to those of other erosion cohesive experiments which were undertaken in straight flumes (Kelly & Gularte, 1981), but the flow volumes from cohesive soil erosion study are several orders of magnitude greater than those for soil creep.

LABORATORY EXPERIMENTS

The experiments on erosion of cohesive sediments were undertaken in an annular rotating fibreglass channel-ring system (Cao & Fang, 1991). Mud from the flood plain of the Yellow River at Huayuanqou was used in the tests. Its $d_{50}$ is 0.0039 mm. The plastic limit of Huayuanqou mud is 24.76%, the liquid limit is 43.39% and the plastic index is 18.63%. The loose pore ratio is 0.81. The mineral composition given by infrared scanner for the portion of $d < 0.002$ mm is: 60-70% of illite and gaolite, 30% of calcite and less than 10% of quartz. The sodium adsorption ratio, SAR, is 5.19.

Twelve runs of erosion resistance experiment used two kinds of consolidating bed with different pore ratio. One bed with a pore ratio of 0.739 and a dry density of 0.70 g mm$^{-3}$ was formed by a mixer. Another with a dry density of 0.55 g cm$^{-3}$ was formed by natural deposition in the annular channel in two months. The free pore ratios are -0.507 and -0.107 respectively. The shear strengths are 10.4 g cm$^{-2}$ and 4.16 g cm$^{-2}$ respectively, which were measured by a fall bore with a bore angler of 60° and a weight of 60 g. The shear strength was calculated by the equation, $\tau' = kQ/h^2$, where $Q$ is the weight of fall bore in gram; $h$, in mm, is the depth of the bed where the apex of the bore reached; $k$ is a constant of 0.3; and $\tau'$ is the shear strength in t m$^{-2}$. During erosion, turbid water samples were taken through sampling taps. The process of concentration increasing with time was recorded in this way. Concentrations under 10 kg m$^{-3}$ were tested by an electro-optical turbidity meter and concentrations above 10 kg m$^{-3}$ by gravimetric analysis.

EROSION RESISTANCE

Figure 1 gives a sample of sediment concentration $c$ varying with time $t$ for erosion experiments. The dry density of the bed is 0.556 g cm$^{-3}$ and bed shear stresses are 28, 46 and 66 N m$^{-2}$ respectively. After about two hours of erosion, the sediment concentrations increase linearly. When shear stress is less than 16.1 N m$^{-2}$, the concentrations remain constant. Those data are not shown in Fig. 1 because the data points of low shear stress are too compact to be shown clearly. In the first two hours, the concentrations increase rapidly since the erosion resistance of the bed surface on which the fresh deposition of the last run has occurred, is quite low. Stable surface erosion occurs after about two hours. The erosion rates, $E$, are constant in the experimental range ($c <$
Erosion resistance of cohesive sediments in turbulent flow

35 kg m\(^{-3}\) and \(\tau < 66.4 \text{ N m}^{-2}\)). The suspended sediments in the water column on a unit bed area can be represented as \(E_0 = V_0c/A\), where the total volume of water in the annular channel \(V_0 = 0.036 \text{ m}^3\) and total bed surface area \(A = 0.565 \text{ m}^2\). \(E_0\) is also the accumulative erosion amount on a unit bed area, in kg m\(^{-2}\).

Figure 2 gives the relation of \(E\) and \(\tau\). When \(\tau > \tau_c\), the equation is

\[
E = (\tau/\tau_c - 1)
\]

where \(\tau_c\) is critical shear stress which can be determined by experiments. Erosion rate can also be expressed by the concentration gradient in volume ratio as \(R = K(\tau/\tau_c - 1)/(\gamma_s h)\) where \(R\) is erosion rate in 1 h\(^{-1}\) and \(\gamma_s\) is the density of sediment in kg m\(^{-3}\). Erosion coefficient \(K\) and critical bed shear stress are very complex and vary with the type of sediments, salinity, ionic species and amount, pH value and temperature of water. A preliminary correlation of \(K\) with free pore ratio \(e_r\) of bed is found in this study: \(e_r\) is a multiple index of sediments, which may reflect the effects of particle size distribution, the content of fine particles and the degree of consolidation. The free pore ratio is defined as \(e_r = N_r/N_m\), where \(N_m\) is the actual pore rate of bed and \(N_r\) is the ratio of the difference between actual pore rate \(N_m\) and the loose pore rate \(N_s\) over the volume concentration in loose condition \(N_s = (N_m - N_s)/(1 - N_s)\). \(N_s\) may reflect the effects of the particle size and the
content of fine particles, which can be determined in laboratory as follows: put a certain amount of sediment with water in a marked vessel, mix well first, then let the sediment fall free, measuring the volume of deposited sediment at the end. The loose pore rate $N_s$ can be calculated. The equation of $e_r$ can be written in terms of the parameters $N_s$ and $N_m$ as
\[
e_r = 1 - N_s(1 - N_m)/[N_m(1 - N_s)].\]
The smaller the value of $e_r$, the lower the free degree of particles, the less the pore between particles, and therefore the larger the cohesion. Based on the data of the experiments, the relationship between $K$ and $e_r$ is given as $K = -13.9 - 133e_r$ where $K$ is in g m$^{-2}$ h$^{-1}$.

APPLICATION OF THE PROCESS THEORY

Some works have shown that the rate process theory may be useful for understanding the erosion resistance of cohesive soil. The similarity of erosion at constant stress to creep at constant stress was recognized and the methodology to determine experimental active energies $E_a$ and flow volume $V$ were developed by Kelly & Gularte (1981). Expressions of $E_a$ (kcal mol$^{-1}$) and $V$ (cm$^3$) can be written as:
\[
E_a = \frac{RT_2T_1}{T_2 - T_1} \ln \left[ \frac{E_2T_1}{E_1T_2} \right]
\]
\[
V = \frac{2kT}{(\tau_2 - \tau_1)} \ln \left[ \frac{E_2}{E_1} \right]
\]

where $\tau$ is the bed shear stress, $E$ is the erosion rate, $R$ is the universal gas constant (1.98 cal K$^{-1}$ mol$^{-1}$), $k$ is Boltzmann's constant ($1.38 \times 10^{-16}$ erg K$^{-1}$ and $T$ is absolute temperature. This study determined the flow volumes only by least squares regression analyses of the data from each run. They are $0.427 \times 10^{13}$ cm$^3$ and $0.158 \times 10^{13}$ cm$^3$ for the two types of bed respectively. Table 1 gives the comparison of this study with other authors. The results show that the flow volumes of this study are close to those of other erosion cohesive experiments.

Table 1 Experimental flow volumes.

<table>
<thead>
<tr>
<th>Material</th>
<th>Type of test</th>
<th>Flow volume (cm$^3$)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Huayuanqou mud</td>
<td>erosion</td>
<td>$0.158-0.427 \times 10^{13}$</td>
<td>This study</td>
</tr>
<tr>
<td>Illite (remodelled)</td>
<td>erosion - water tunnel</td>
<td>$0.154-0.69 \times 10^{13}$</td>
<td>Kelly &amp; Gularte*</td>
</tr>
<tr>
<td>Illite</td>
<td>erosion - pipe</td>
<td>$0.81 \times 10^{13}$</td>
<td>Christensen &amp; Das*</td>
</tr>
<tr>
<td>Kaolinite</td>
<td>erosion - pipe</td>
<td>$0.49 \times 10^{13}$</td>
<td>Christensen &amp; Das*</td>
</tr>
<tr>
<td>San Francisco mud</td>
<td>erosion - flume</td>
<td>$0.13 \times 10^{13}$</td>
<td>Partheniades*</td>
</tr>
<tr>
<td>Kaolinite</td>
<td>erosion - flume</td>
<td>$0.13 \times 10^{13}$</td>
<td>Raudkivi et al.*</td>
</tr>
<tr>
<td>San Francisco mud</td>
<td>creep</td>
<td>$8.7 \times 10^{18}$</td>
<td>Mitchell et al.*</td>
</tr>
<tr>
<td>Illite (remodelled)</td>
<td>creep</td>
<td>$7.3 \times 10^{18}$</td>
<td>Mitchell et al.*</td>
</tr>
<tr>
<td>Illite (remodelled)</td>
<td>creep</td>
<td>$3.0 \times 10^{18}$</td>
<td>Mitchell et al.*</td>
</tr>
<tr>
<td>Sault St Marie clay</td>
<td>creep</td>
<td>$0.6-4.2 \times 10^{18}$</td>
<td>Andersland et al.*</td>
</tr>
</tbody>
</table>

*Kelly & Gularte (1981).
which were undertaken on straight flumes (Kelly & Gularte, 1981), but the flow volumes from cohesive soil erosion studies are several orders of magnitude greater than those for soil creep. After the flow volume has been determined, the erosion rate can be calculated by the following equation:

\[ E_2 = E_1 \exp \left[ \frac{V(\tau_2 - \tau_1)}{2kT} \right] \]

**DRY DENSITY VERSUS CRITICAL SHEAR**

For engineering applications, erosion experiments with three kinds of soil were conducted in a straight flume with a length of 14 m, a bed slope of 0.05 and a width of 0.5 m. The soil samples to be tested were put in test boxes measuring $31 \times 17 \times 2.3$ cm (length, width and depth) which were located in the centre of the flume. The surfaces of soil samples were the same level as the flume bed. The dry densities of soil samples, which were carefully prepared using a mixer, ranged from 0.64 to 0.98 g cm$^{-3}$. The characteristics of the soil samples are shown in Table 2.

<table>
<thead>
<tr>
<th>Material</th>
<th>$d_{50}$ (mm)</th>
<th>Initial dry density (g cm$^{-3}$)</th>
<th>Fluid limit (%)</th>
<th>Plastic limit (%)</th>
<th>Plastic index (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Suicaozi mud</td>
<td>0.0058</td>
<td>0.64</td>
<td>51.95</td>
<td>30.00</td>
<td>21.95</td>
</tr>
<tr>
<td>HYQ mud</td>
<td>0.004</td>
<td>0.505</td>
<td>43.39</td>
<td>24.76</td>
<td>18.63</td>
</tr>
<tr>
<td>Beijing soil</td>
<td>0.023</td>
<td>0.56</td>
<td>30.90</td>
<td>19.70</td>
<td>11.20</td>
</tr>
</tbody>
</table>

During experiments, the discharge was increased step by step until the samples were deformed by fluid erosion. The samples of all runs were eroded as a type of structural deformation. For specific engineering problems, it is practical to develop a relation between dry density of cohesive sediment and the erosion critical bed shear stress. From the data of this study, the relation is:

\[ \tau_0 = \alpha \gamma^\beta \]

in which $\alpha$ and $\beta$ are constants. For this study, $\alpha \approx 0.7$ and $\beta \approx 5$ using the upper limit of the experimental data.

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Soil erosion studies using rainfall simulation on forest harvested areas in British Columbia

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Abstract A field portable rainfall simulator was employed to determine the infiltration and soil erosion response of forest harvested areas (skidroads, fireguards and slash burned sites) in coastal and interior British Columbia. The results indicate that infiltration capacity decreases as total bulk density and coarse fragment content increase. The suspended sediment concentration of the overland flow was found to increase as the volume of the runoff increased. However, variability in suspended sediment concentrations was high for the highest runoff volumes which suggests that other factors such as soil texture, slope gradient and surface armouring also have important effects. Future work will investigate the role of soil texture on infiltration and soil erosion.

INTRODUCTION

Soil erosion on forested land in British Columbia is a serious concern that can have both on-site and off-site detrimental effects. The impacts of erosion on stream resources in the form of fisheries habitat and water quality degradation have received the most attention. Perhaps equally important are the on-site effects including forest road damage, and soil degradation and resulting reduction in site productivity. Surface erosion is mainly associated with bare mineral soil surfaces such as haul roads, skidroads and other bladed or otherwise exposed areas of mineral soil. Numerous studies have made use of artificial rainfall to study infiltration and surface soil erosion on both forest and agricultural land, both in the field and in the laboratory (Luk et al., 1986; Meeuwig, 1971; Munn & Huntington, 1976, Roth et al., 1985; Wilson & Rice, 1990). The major advantage of rainfall simulation research is that it is more rapid, efficient, controlled and adaptable than natural rainfall research (Meyer, 1988). In this study, a field portable rainfall simulator was employed to determine the infiltration capacity and potential surface erosion response of disturbed sites.

METHODS

During the summer of 1990, rainfall simulation experiments were conducted at
a site in coastal British Columbia (Iron River) and at a location in the interior of the province (Cariboo Lake) (Fig. 1). The Iron River site is one of five sites chosen to study the effects of wide-tyre skidder traffic on forest soils (Rollerson 1989). On this site, which was treated during the spring of 1988, five plots each were located on the 5, 20 and 80 skidder pass trails (track and mid-track locations) for a total of 15 rainfall simulation experiments. The Cariboo Lake location consists of two sites: a winter skidder logged area (1987/1988) where experimental biomass harvesting (summer 1988), and prescribed fire (summer 1989) treatments were conducted (Cariboo-1); and a summer logged cutblock (summer 1988) where bladed skidroads were employed (Cariboo-2). At the Cariboo-1 site, two plots were located on skidroads, two on fireguard surfaces (bladed roads designed to contain prescribed fires), and one on an intensively prescribe-burned site (complete consumption of the forest floor). The five plots established at the Cariboo-2 site were located on bladed skidroads (track to mid-track locations). Table 1 lists the characteristics of the plots sampled at each site.

Soil bulk density for the 0-10 cm depth was measured at each plot using either the excavation method (sand cone) or a nuclear densiometer (Troxler model 4311). A sub-sample of 10 soils demonstrated that a high correlation exists between the two measures of soil bulk density at these sites ($r = 0.97$). Soil samples were collected from the 0-10 cm depth and analysed for coarse fragment content and soil texture. The pre-test soil moisture content was determined by collecting gravimetric samples from around the perimeter of the plot prior to the beginning of the test.

A number of simple linear regressions were performed on the combined Cariboo Lake and Iron River data, using infiltration capacity and the suspended sediment concentration of the runoff as dependent variables, and total bulk density, coarse fragment content, pre-test soil moisture content, volume of runoff, and silt content (%) as independent variables. A number of other variables were not significant in explaining variation in infiltration capacity or suspended sediment concentration i.e. fine (<2 mm) soil density, vegetation
Soil erosion studies using rainfall simulation on forest harvested areas

Table 1 Site characteristics.

<table>
<thead>
<tr>
<th>Biogeoclimatic zone/subzone</th>
<th>Iron River</th>
<th>Cariboo-1</th>
<th>Cariboo-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Elevation (m)</td>
<td>CWH</td>
<td>ICHh2</td>
<td>ICHh2</td>
</tr>
<tr>
<td>Slope gradient (%)</td>
<td>820</td>
<td>1200</td>
<td>900</td>
</tr>
<tr>
<td>Soil classification</td>
<td>Podzols</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil texture(^b)</td>
<td>L</td>
<td>SL-SiL</td>
<td>SL-L</td>
</tr>
<tr>
<td>Coarse fragment content (%)</td>
<td>7-25</td>
<td>32-44</td>
<td>18-45</td>
</tr>
<tr>
<td>Vegetation cover (%)</td>
<td>0-20</td>
<td>0-70</td>
<td>0-12</td>
</tr>
</tbody>
</table>

* CWH = Coastal Western Hemlock; ICHh2 = Interior Cedar-Hemlock (Cariboo River variant); ESSFh = Engelmann Spruce-Subalpine Fir (h subzone).
\(^b\) L = loam, SL = sandy loam, SiL = silty loam.
\(^c\) BC Ministry of Forests (1987).

cover, slash cover, slope gradient, and clay content (%).

DESCRIPTION OF THE RAINFALL SIMULATOR

The rainfall simulator was designed and constructed at the Pacific Forestry Centre and field tested during the summer of 1989 (Fig. 2). It consists of an air-tight chamber constructed out of heavy gauge aluminum (76.5 × 76.5 × 5.5 cm) and equipped with 324 Teflon capillary tubes (0.56 mm inside diameter) arranged in a grid pattern. Adjustable telescopic legs are fastened to the chamber by means of U-shaped brackets. With the legs fully extended (2.7 m above the

Fig. 2 Rainfall simulator and overland flow collection system.
The rainfall produced attains 75% of the terminal velocity of similar sized (3 mm) raindrops (Epema & Riezebos, 1983). A 36-litre reservoir tank is held in place above the chamber by means of a frame constructed out of angle iron. Flexible PVC tubing connects the reservoir to an in-line flow meter (1 l min\(^{-1}\) capacity), and connects the flow meter to a valve which is attached to the rainfall chamber. The reservoir tank is fitted with a constant head device (a piece of rigid tubing inserted through the top of the tank and extending almost to the bottom) to maintain the pressure within the tank at a constant level. At a rainfall intensity of 4.5 cm h\(^{-1}\), rain can be applied for up to 75 min.

A set of three overland flow collection troughs (30 cm across) are installed a few centimetres below the first row of drop formers (determined by means of a plumb bob), and placed perpendicular to the slope. Each trough, equipped with a bent lip, is pushed tightly against the soil, held in place with spikes and sealed against the soil with fast setting plaster. A larger trough collects the flow from the three smaller troughs and directs the flow to a collection can. A set of nine cans (12.5 cm diameter) arranged in a grid pattern is used to determine rainfall uniformity and to calibrate the flow meter. An average coefficient of variation (\(C_V = \frac{\text{standard deviation}}{\text{mean}} \times 100\%\)) of 8.9% was calculated for a sample of eight calibration runs.

During a test, the rainfall rate is kept constant and overland flow is collected and the volume measured every 2.5-5 min depending on the volume of runoff. The infiltration rate (cm h\(^{-1}\)) is derived using the difference between the amount applied and the amount collected with the overland flow troughs over the selected time period. These values are plotted as a function of time (Fig. 3). The infiltration capacity for a plot is defined as the equilibrium infiltration rate and is estimated from the graph by projecting the horizontal asymptotic value. Because the tests were designed to provide an index of the potential for soil erosion, the plots were not pre-wetted to attain the saturated soil conditions under which infiltration capacity is normally determined. The objective was to determine the typical soil response to high intensity rainfall events and therefore

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**Fig. 3** Typical soil infiltration curve (Cariboo-1, skidroad surface).
Soil erosion studies using rainfall simulation on forest harvested areas

the soils were tested "as is". Soil moisture contents (by weight) were found to range from 13 to 37% at the Cariboo Lake sites, and from 28 to 56% at Iron River.

RESULTS

The regression of infiltration capacity versus total bulk density is very significant ($p = 0.0001$) and bulk density alone explains 50% of the variability in infiltration capacity (Fig. 4). As expected, the relationship is one of decreasing infiltration with increasing density. However, the regression of infiltration capacity on coarse fragment content, significant at the $\alpha = 0.01$ level ($r^2 = 0.384$), is also one of decreasing infiltration capacity with increasing coarse fragment content. The reason for this unexpected result appears to be the high positive correlation between total bulk density and coarse fragment content ($r = 0.914$), which ensures that the two relationships behave in a similar way. The regression of infiltration capacity versus pre-test soil moisture content, significant at the $\alpha = 0.05$ level, indicates that infiltration capacity increases as moisture content increases. This may be an artifact of combining the data, because the Cariboo Lake plots were drier (late June) and generally denser, whereas the Iron River plots tended to be wetter (late May) and had lower soil densities (and therefore higher infiltration capacities).

Suspended sediment concentration was regressed against total bulk density, coarse fragment content, runoff volume, and vegetation cover. Only runoff volume ($p = 0.0063, r^2 = 0.282$) helped to explain the variation in the observed suspended sediment concentrations (Fig. 5). Not surprisingly, concentrations increase as runoff increases. It should be noted that the variability of the suspended sediment concentrations increases as the runoff volume increases.

![Infiltration vs. Total Bulk Density](image)

**Fig. 4** Infiltration capacity regressed against total soil bulk density for the combined Cariboo Lake and Iron River data.
Although not significant \((p = 0.089)\), the regression of suspended sediment concentration versus vegetation cover suggests a trend of decreasing suspended sediment concentration with increasing vegetation cover which seems intuitively correct.

The Iron River data were analysed separately for suspended sediment concentrations. The regression against runoff volume is more significant than for the combined data \((p = 0.0128)\): runoff volume explains 39% of the variability in the concentration of suspended sediment. To determine if soil texture affected the amount of soil eroded and transported by runoff, the suspended sediment concentrations were regressed against clay content and silt content. The regression versus silt content is significant \((p = 0.0127)\) and explains 39% of the variability in concentration, whereas the regression versus clay content is not significant. The range in silt content for the soils at Iron River is relatively small (33-44%) and these results need to be confirmed for other sites and soil textures.

In addition to the rainfall experiments described above, a test was undertaken on a harvested and slash-burned, but otherwise undisturbed, location at Cariboo-1 (duff layer 6 cm deep, vegetation cover 20%). The results of this test confirmed expectations of high infiltration capacities on such sites \((>7.5 \text{ cm h}^{-1})\).

**DISCUSSION**

The data clearly indicate a strong relationship between the infiltration capacity of exposed mineral soil surfaces and the total bulk density in the 0-10 cm surface layer. In addition, the almost equally strong relationship between infiltration capacity and coarse fragment content, and the high positive
correlation between bulk density and coarse fragment content, suggest that coarse fragment content could be used as a reasonable surrogate property for determining the expected soil infiltration response. This would necessitate the development of site-specific infiltration/coarse fragment content predictive models. The variation in coarse fragment content for the soils in this study was between 5 and 45%. Soils with higher coarse fragment contents could very well behave differently.

The concentration of suspended sediment in the plot runoff was shown to be related to total runoff volume, and both the Iron River and Cariboo Lake sites appear to behave similarly in this respect. However, for the highest runoff volumes, the variability in suspended sediment concentration increases, which suggests that a simple linear model may not be the most appropriate. For the Iron River data, soil texture (in the form of silt content) explained almost 40% of the variation in suspended sediment concentration. Additional field experiments will be required to define the relationship between sediment production and soil texture for British Columbia conditions more fully. Numerous additional parameters would be expected to affect the suspended sediment concentration of runoff, such as slope gradient, surface armouring and organic matter content.

The statistical analyses conducted on the Iron River data alone \((n = 15)\) demonstrated that variability within this site is lower than when the data are grouped with the Cariboo Lake data. This emphasizes that infiltration, runoff and surface soil erosion are processes that can be affected by site-specific factors, which must be clearly identified before predictive models can be developed.

The role of soil moisture content in influencing infiltration capacity is not well defined in this study because of the confounding effects of combining the data from two study locations. However, soil moisture content is expected to affect the infiltration response observed at a site. This is supported by Johnson & Besehta (1981) who reported on variation in infiltration capacities due to seasonal fluctuations in soil temperature and moisture levels.

**FUTURE WORK**

Additional skidroad impacted sites located in southeast British Columbia were examined during the summer of 1991, with the specific objective of sampling a range of soil textures. The experimental technique was changed to accommodate a second hour of rainfall to test the hypothesis that erosion slows down as surface fines are washed away. In addition, a micro-trench experiment was added at each plot to look at the effect of concentrated flow within a rill-sized feature. The data will be analysed and the results will be used to refine a surface erosion hazard key contained in a field handbook (Lewis & Carr, 1989), devised to assist forest managers in minimizing forest site degradation in the interior of British Columbia.
Acknowledgements  The author would like to thank Brian Sieben and Ed Wass for their assistance with the field component of this study, and Steve Glover and Eugene Hetherington for their review comments.

REFERENCES


Erosion and sediment yield in mountain regions of the world

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Abstract. A comparative assessment of erosion intensity in mountain regions of the world has been made based on suspended sediment yield data from 1872 mountain rivers. Quantitative analysis indicates that erosion in mountain regions depends on climate and runoff, landscape character and the degree of economic development, relief, recent tectonic activity, and underlying geology. The most intensive erosion and denudation are found in the glacial and subnival zones, whilst the mountain taiga zone is the least eroded area. A combination of zonal and azonal factors conducive to erosion have given rise to extremely intensive erosion in the mountain areas of east and southeast Asia and New Zealand. In the forest zones, channel erosion dominates slope erosion. Where forest cover is absent due to either natural conditions or anthropogenic activity, slope erosion and denudation become the principal source of sediment in the river network. Depending on the predominance of either channel or slope erosion, two types of mountain erosion systems can be distinguished: in the first sediment yield depends on the basin area directly, in the second the relationship is indirect.

INTRODUCTION

Sediment yield is not an accurate measure of all the products of erosion and mechanical denudation, since a considerable part of the latter is accumulated in the form of slope, proluvial, and alluvial deposits and is not transported to the outlet of the drainage basin. Delivery ratio coefficients, representing the ratio of the material transported out of the basin to all the products of mechanical denudation, may only be used to provide a tentative assessment of the total denudation, since even in mountain basins this ratio does not exceed 0.5. Nevertheless, sediment yield directly depends on the intensity of erosion and the overall mechanical denudation in the river basin and can be used for comparative assessments of the intensity of these processes under different conditions. The aim of this article is to provide a quantitative assessment of the intensity of erosion in mountain regions caused by major natural and anthropogenic factors.

DATA SOURCES

For characterizing erosion, we collected and synthesized data on suspended
sediment yield for 3700 river basins from around the world, including 1872 in mountain regions. Bed load data were obtained for 269 basins, including 158 in mountain areas. For all these basins we collected and presented data on the area, height or type of relief, runoff, underlying geology, landscape patterns, and the degree of anthropogenic modification of the landscape (Dedkov & Moszherin, 1984). Most were from mountain regions of Europe (Caucasus, Carpathians, Alps, Appenines, Scandinavian Mountains, Urals, and Balkan Mountains), North America (the mountainous west of the USA and Canada and the Appalachians), Asia (Pamirs, Tien Shan, the mountains of Siberia, Indo-China, and the Hindu Kush), New Zealand, and southern Africa. There were few data for the mountains of Australia and South America.

The sediment yield per unit area or specific sediment yield \((\text{t km}^{-2} \text{ year}^{-1})\) was taken as the main index of sediment yield. The mean value of this index for a group of basins can be expressed by both the arithmetic mean and a weighted average calculated as the total sediment load from the group of basins divided by the total area of the basins. In this study we used the first value. The variability of sediment yield within the group of basins is expressed by an error term for the mean sediment yield given as a percentage.

Bed load yields from the mountain regions on average comprise 23% of the suspended load.

For the analysis, we used basins with areas ranging from 500 to 100 000 km\(^2\). In order to take account of the dependence of specific sediment yield on the area of the basin, data from small river basins and large river basins were treated separately. The dividing threshold was set at 5000 km\(^2\).

Human activity in mountains results in forest destruction, cultivation of slopes, deterioration of grassland due to overgrazing, and the detrimental consequences of road building and construction of water storage reservoirs. However, the impact of human activity is not so important in influencing erosion as in lowland regions where the natural landscapes have been modified by man to a much greater degree. Nevertheless, the activities of man in mountain areas complicate the relationships between erosion and natural factors and the impact of human activity needs to be evaluated quantitatively. All the basins have been divided into three categories depending on the level of economic development. Characteristic indices for each of these categories (residual woodland and proportion of cultivated land) are given in Table 1.

Table 1 illustrates the strong impact of the economic use of a basin on the intensity of erosion. The analysis of the influence of natural factors on erosion and sediment yield in mountain areas has been undertaken by taking account of this factor. In general terms, we considered basins with a low degree of economic development (Category I) to represent natural landscapes. Based on this reconstruction of natural conditions, the sediment yields from mountain regions of the Earth as a whole can be seen to have increased by 1.4 times.

Reservoirs have been built on 60 of the rivers for which sediment yield data were available. On average, reservoirs cause a 50% reduction in sediment
### Table 1
Suspended sediment yields of mountain rivers (t km\(^{-2}\) year\(^{-1}\)) according to the category of economic development.

<table>
<thead>
<tr>
<th>Degree of development</th>
<th>% Forest cover</th>
<th>% Cultivated land</th>
<th>Small rivers</th>
<th>Large rivers</th>
</tr>
</thead>
<tbody>
<tr>
<td>I low</td>
<td>70</td>
<td>30</td>
<td>744</td>
<td>250</td>
</tr>
<tr>
<td>II medium</td>
<td>30-70</td>
<td>30-70</td>
<td>451</td>
<td>141</td>
</tr>
<tr>
<td>III high</td>
<td>30</td>
<td>70</td>
<td>136</td>
<td>81</td>
</tr>
</tbody>
</table>

\(N\) is the number of basins; \(s\) = mean suspended sediment yield (t km\(^{-2}\) year\(^{-1}\)).

*The error term (%) for the mean value is given in parentheses.

We therefore excluded these rivers from the further analysis of controlling factors.

### EROSION AND RELIEF

Under natural conditions only slightly changed by man (basins of Category I) the intensity of erosion directly depends on the amplitude of relief. The economic activity of man, being different in different altitudinal zones, has, however, modified this simple dependence (Fig. 1).

A particularly marked increase in erosion is observed in low mountains, where economic activity has reached a high level. However, in intensely developed lowland areas erosion has increased to a still greater degree. This

![Diagram](image)

**Fig. 1** The dependence of suspended sediment yield on relief.
has led to a decrease in the difference in erosion intensity between mountains and plains. In landscapes that are completely natural or only slightly changed by man, erosion in mountain areas is 27 times greater than in lowland areas. In the case of anthropogenically changed landscapes, this increase reduces to 3.2 times.

Thanks to a greater preservation of the natural landscape, some mountain areas are less eroded than the adjoining densely populated plains. For example, sediment yields in the South and Middle Urals (up to 30 t km\(^{-2}\) year\(^{-1}\)) are less than in the neighbouring eastern part of the Russian Plain (up to 200 t km\(^{-2}\) year\(^{-1}\)).

Recent tectonic activity also influences the intensity of mountain denudation. Table 2 shows that with an increase of earthquake intensity by 1 point, the intensity of denudation in mountain areas increases almost twofold. However, one should bear in mind that the data presented in this Table also reflect the effect of mountain height, since higher mountains are usually characterized by increased seismic activity.

<table>
<thead>
<tr>
<th>Intensity of earthquakes (points on the Richter scale)</th>
<th>(N)</th>
<th>(s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9</td>
<td>133</td>
<td>540</td>
</tr>
<tr>
<td>8-9</td>
<td>293</td>
<td>360</td>
</tr>
<tr>
<td>7-8</td>
<td>529</td>
<td>180</td>
</tr>
<tr>
<td>6-7</td>
<td>51</td>
<td>47</td>
</tr>
<tr>
<td>6</td>
<td>88</td>
<td>17</td>
</tr>
</tbody>
</table>

\(N\) is the number of basins, \(s = \text{mean suspended sediment yield} (t \text{ km}^{-2} \text{ year}^{-1})\).

**EROSION AND CLIMATE-LANDSCAPE CHARACTERISTICS**

Climate and the resultant landscape characteristics influence erosion processes through two principal factors, namely, runoff and the degree of coverage of the surface by vegetation. Based on sediment yield data, the greatest mechanical denudation occurs in the glacial and subnival zones as well as in the subtropics in the Mediterranean zone (Fig. 2). All these mountain zones are characterized by limited protection of their soils by vegetation due to natural and anthropogenic (Mediterranean) factors. Runoff amounts are considerable here (1.5 or 2 times as great as the mean global value). Surface mechanical denudation is substantial, leading to a predominance of slope erosion over channel erosion.

In forest zones, erosion is weak with slope erosion being generally insignificant. All the processes of mechanical denudation are concentrated in the channels of rivers and on their banks. In forest zones, erosion intensity is closely related to runoff amount. Hence rivers in the tropics and subtropics evidence higher erosion rates than rivers in forest zones located in the
Erosion and sediment yield in mountain regions

Fig. 2 Suspended sediment yield \((s, \text{t km}^{-2} \text{year}^{-1})\) in different landscape zones and belts of mountain regions.

temperate belt. There are only a few data for mountain rainforest areas, and these need to be checked. In the mountainous semi-deserts and steppes, erosion is insignificant due to the aridity of the climate and low runoff amounts.

It should be noted that for both mountains and plains, the rule of Langbein-Schumm (1958) which gives maximum erosion in semi-arid areas is not valid. Other researchers (Walling & Kleo, 1979; Jansson, 1988) have come to the same conclusion. In some measure this rule holds true where basin erosion depends primarily on the protective role of vegetation. For channel erosion, the dependence on vegetation is insignificant. The dependence of sediment yield on runoff in mountain areas is clearly evident, especially in basins with landscapes little changed by man.

EROSION AND ROCK TYPE

In mountain areas, the influence of rock type generates the following sequence of sediment yield: loess \((s = 1800 \text{t km}^{-2} \text{year}^{-1})\), other terrigenic loose rocks (1300), coarse terrigenic rocks (550), metamorphic (420), limestones (310), igneous (100).

The selectivity of erosion in mountain areas is much greater than in plains areas. This feature is particularly prominent in mountain semi-deserts and in the Mediterranean (Dedkov & Moszherin, 1984), where tectonic
structures occur. The weakest selectivity is characteristic of mountain forests in the temperate and tropical belts.

THE COMBINED IMPACT OF DIFFERENT FACTORS

Various combinations of factors controlling erosion in mountain regions create conditions for intense erosion. The Pacific Asiatic-Australian sector evidences the most intensive erosion on the Earth. Based on data from 518 river basins in the mountain areas of this zone, the mean suspended sediment yield is 800 t km\(^{-2}\) year\(^{-1}\). Particularly intensive erosion occurs on the islands separated from the continents of Asia and Australia by ocean basins. In Taiwan a sediment yield of 31 700 t km\(^{-2}\) year\(^{-1}\) has been documented with background values of up to 5000 t km\(^{-2}\) year\(^{-1}\) (Li, 1976). Somewhat lower (11 000 or 12 000 t km\(^{-2}\) year\(^{-1}\)) are the sediment yields reported for Java (Walling & Webb, 1983). In New Zealand, suspended sediment yields also reach high values - between 20 000 and 28 000 t km\(^{-2}\) year\(^{-1}\), with a mean value of about 2000 t km\(^{-2}\) year\(^{-1}\) (Adams, 1979).

The main reasons for such intense erosion throughout the whole sector are as follows: high precipitation amounts and intensities, high and irregular runoff (in New Zealand up to 280 l s\(^{-1}\) km\(^{-2}\)), dissected mountain relief composed mainly of sedimentary rocks, intense recent tectonic activity, and replacement of forests by agricultural land in the low mountain areas.

The least intense erosion is found in the low mountains of the temperate zone that are mainly underlain by crystalline rocks and covered by dense forests (Scandinavia, Urals, the mountains of South Siberia, the Trans-Baikal region, etc.). In some locations within such regions, suspended sediment yields are only 10 or 20 t km\(^{-2}\) year\(^{-1}\).

THE EROSIONAL SYSTEM AND ITS OPERATION

The complex of erosional processes operating in a river basin can be represented as an open dynamic system whose functioning is determined by the movement of water and sediment. Channel and slope erosion are the main elements of the system. Depending on the relative importance of these two elements, we may distinguish two types of system behaviour.

The first type is characterized by a predominance of channel erosion. It is common in forest areas with well preserved forest vegetation (basins of Category I). Slope erosion is minimal here. Erosional processes are mainly concentrated in river channels. The sediment yield of predominantly channel origin is proportional to runoff in basins of Category III (Makkaveev, 1955). That is why there is a downstream increase not only in sediment load but also in specific sediment yield. The specific sediment yield is positively related to basin area (Fig. 3(a)).
Erosion and sediment yield in mountain regions

The second type of system is characterized by a predominance of slope (sheet, gully) erosion. It occurs in natural zones with a poor vegetation cover and with strong surface erosion and mechanical denudation (glacial, subnival, and semiarid zones). It is also common in forest zones with a high degree of economic development (basins of Category III and partly of Category II). Much of the denudation occurs in the upper parts of the river system. When moving down the rivers, part of the transported sediment accumulates in the channels and on the flood plains. This leads to a decrease of specific suspended sediment yield and to an inverse relationship between specific sediment yield and basin area.

The two types of erosional system behaviour are connected by a gradual transition. To produce a tentative quantitative estimate of the ratio of slope to channel erosion, we have developed and used a method based on an analysis of the dependence of suspended sediment transport on water discharge during low-water periods (Dedkov & Moszherin, 1984).

The form of the relationship between specific sediment yield and basin area is also a reliable index of the type of erosional system. A positive relationship is most common for the first type of behaviour and a negative relationship for the second. It is, however, important to take account of other factors influencing the relationship, particularly rock type, relief, and human activity (reservoirs etc.).
REFERENCES


Ten-years of sediment discharge measurement in the Jasenica research drainage basin, Yugoslavia

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Abstract The Jasenica drainage basin is a representative basin for the hilly mountain region in the central part of the Serbian Republic in Yugoslavia. It was selected as a research basin for a variety of research in forestry, hydrology, erosion and sediment transport. Since the 1980s, water discharge and suspended sediment concentration have been measured at the D. Satornja gauging station in the basin. The results of 10 years of data collection for water and sediment discharge make it possible to test the methods most widely applied in Yugoslavia for estimating annual water discharge and sediment yield. The estimates produced by these methods, namely the Poljakov method, the Keller method and the Herheulidze method, are compared with the measured values of water discharge and suspended sediment yields for the period 1980-1989 on the basis of an average year. The results show that these methods have serious limitations for practical use, at least in the central part of the Serbian Republic.

INTRODUCTION

The prediction of total annual sediment yield from river basins in hilly mountain regions is very important for land use planning, for developing erosion control measures on agricultural land, and for the design of reservoirs. It is well known that this information can be obtained by the use of various equations, formulae or models calibrated against long-term measurements. Such measurements are rare in the hilly mountain regions of Yugoslavia and there is therefore always uncertainty as to the applicability of methods or models developed elsewhere. The Jasenica basin has been selected as a representative hilly mountain river basin for the central part of the Serbian Republic, and it has been used as a research basin since the 1980s. Various research projects in forestry, erosion and sediment transport and hydrology have been undertaken. Some of the results of 10-years of suspended sediment yield measurements between 1980 and 1990 are presented in this paper.

METHOD

Water and sediment discharges from the Jasenica basin were measured at the
D. Satornja gauging station. Water discharge was measured using standard equipment comprising a water level recorder with 24-h rotation installed above a broad crested weir. Suspended sediment concentration was measured by manual sampling. One 5 litre sample was collected each day and assumed to be representative of that day. During flood flows, additional manual sampling was undertaken every 2-3 h on the rising stage of the flood. This manual sampling strategy has its limitations, but it is acceptable for stable flows with stable concentration. Sediment concentrations were measured by routine laboratory methods involving filtering, drying and weighing.

RESULTS

The Jasenica River basin is typical of hilly mountain rivers in the Serbian Republic. Its physical characteristics are:

- Area = 95.56 km
- Valley slope = 22.4%
- Length = 12.00 km
- River slope = 3.20%
- Circumference = 47.20 km
- Mean rainfall = 760 mm
- Mean elevation = 507.84 m

The soils and land use are given in Table 1. A summary of the annual water and sediment discharge data for the whole period of investigation is presented in Table 2.

<table>
<thead>
<tr>
<th>Soil type</th>
<th>Area (%)</th>
<th>Land use</th>
<th>Area (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brown acid soils</td>
<td>83.0</td>
<td>Good forest (hardwoods)</td>
<td>42.26</td>
</tr>
<tr>
<td>Brown forest soils</td>
<td>13.3</td>
<td>Cultivated land</td>
<td>33.24</td>
</tr>
<tr>
<td>Smonitzas</td>
<td>2.5</td>
<td>Pasture</td>
<td>22.72</td>
</tr>
<tr>
<td>Alluvial soils</td>
<td>1.2</td>
<td>Orchards</td>
<td>1.78</td>
</tr>
</tbody>
</table>

The procedure commonly employed for estimating sediment yields ($V$, m$^3$ year$^{-1}$) in Yugoslavia is the Polakov method (Jeftic, 1978):

$$V = Q_c (M_o A/10^3 g_v) \times 31.5 \times 10^6 \text{ (m}^3 \text{ year}^{-1})$$

where:

- $Q_c = \text{mean sediment concentration (g m}^{-3}) = a \sqrt{i \times 10^8}$ with $a = \text{erosion index (}= 1.4 \text{ for Jasenica)}$, $i = \text{average river slope (}= 3.2\% \text{ for Jasenica)}$;
- $M_o = \text{mean water discharge (m}^3 \text{ s}^{-1} \text{ km}^2)$;
- $A = \text{area (km}^2)$;
- $g_v = \text{sediment density (kg m}^{-3})$.

The mean water discharge is calculated either by the Keller method ($M_{ol}$)
or by the Herheulidze method ($M_{o2}$). viz.:

$$M_{o1} = (0.942P - 430) \times 10^3/31.5 \times 10^6 \text{ (m}^3\text{s}^{-1}\text{km}^{-2})$$

where $P$ = annual rainfall in mm;

$$M_{o2} = fA^{0.86}/A \text{ (m}^3\text{s}^{-1}\text{km}^{-2})$$

where $f$ = aridity index (for Jasenica $f = 0.13$).

### Table 2

<table>
<thead>
<tr>
<th>Year</th>
<th>$Q$ ($m^3\text{year}^{-1} \times 10^6$)</th>
<th>$Q_s$ (t year$^{-1}$)</th>
<th>Year</th>
<th>$Q$ ($m^3\text{year}^{-1} \times 10^6$)</th>
<th>$Q_s$ (t year$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1980</td>
<td>13.91</td>
<td>1304.0</td>
<td>1985</td>
<td>12.28</td>
<td>2839.19</td>
</tr>
<tr>
<td>1981</td>
<td>29.43</td>
<td>3183.2</td>
<td>1986</td>
<td>18.49</td>
<td>3991.00</td>
</tr>
<tr>
<td>1982</td>
<td>18.29</td>
<td>1940.2</td>
<td>1987</td>
<td>18.03</td>
<td>6123.70</td>
</tr>
<tr>
<td>1983</td>
<td>11.67</td>
<td>1228.8</td>
<td>1988</td>
<td>16.04</td>
<td>2883.30</td>
</tr>
<tr>
<td>1984</td>
<td>32.75</td>
<td>3814.3</td>
<td>1989</td>
<td>19.05</td>
<td>3128.30</td>
</tr>
</tbody>
</table>

Mean annual water yield = $18.99 \times 10^6 \text{m}^3\text{year}^{-1}$
Mean annual sediment yield = $3043.6 \text{t year}^{-1}$
Mean water discharge ($M_o$) = $0.00624 \text{ m}^3\text{s}^{-1}\text{km}^{-2}$
Mean sediment concentration ($Q_c$) = $160.3 \text{ g m}^3$

### Table 3

<table>
<thead>
<tr>
<th>Mean water discharge ($M_o$) ($m^3\text{s}^{-1}\text{km}^{-2}$)</th>
<th>Mean sediment concentration ($Q_c$) (g m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Measured</td>
<td>0.006 24</td>
</tr>
<tr>
<td>Keller estimate</td>
<td>0.009</td>
</tr>
<tr>
<td>Herheulidze estimate</td>
<td>0.068</td>
</tr>
<tr>
<td>Poljakov estimate</td>
<td>2504.0</td>
</tr>
</tbody>
</table>

### Table 4

<table>
<thead>
<tr>
<th>Sediment yield estimates</th>
<th>t year$^{-1}$</th>
<th>t km$^{-2}$ year$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Measured</td>
<td>3 043.60</td>
<td>31.52</td>
</tr>
<tr>
<td>Sediment yield estimates by Poljakov + Keller</td>
<td>68 546.40</td>
<td>709.00</td>
</tr>
<tr>
<td>Sediment yield estimates by Poljakov + Herheulidze</td>
<td>517 906.00</td>
<td>5363.60</td>
</tr>
</tbody>
</table>

A comparison of measured and estimated values of sediment and water yield for the Jasenica River basin is provided in Tables 3 and 4.
CONCLUSIONS

Water and sediment yield measurements in the Jasenica basin during the period 1980-1990 offered the possibility of testing the accuracy of the Poljakov, Keller and Herheulidze equations under Yugoslavian conditions, where they are in common use. The main conclusions are summarized as follows:

(a) The Herheulidze method overestimates the mean water discharge ($m^3 \cdot s^{-1} \cdot km^2$) eleven-fold when compared with the measured value, and therefore cannot be recommended for practical use. For the same calculation, the Keller method is much more appropriate and can be applied with a reasonable degree of confidence (Table 3).

(b) If we accept that the method used for measuring sediment concentrations in the Jasenica basin was adequate and reasonably accurate, the Poljakov method for estimating sediment concentrations (g m$^{-3}$) can be seen to be problematic, since it leads to incorrect estimates of sediment yield, especially when combined with the Herheulidze formula (Table 4).

(c) All the methods tested need further refinement of their parameter values. This can only be achieved by experimental studies under various natural conditions. Until such refinement is achieved, the use of the methods in Yugoslavia, and especially in the Serbian Republic, cannot be recommended.

REFERENCE

Criteria for determining the current activity of torrents in their depositional areas

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Abstract Alluvial fans formed by the perpetual transport and deposition of material by torrential streams, frequently exhibit trenches on their main body. According to some scientists these trenches represent recent torrent activity in the deposition area. The deposits within these trenches can be classified into several categories depending on their residence time. The classification is based on criteria relating to their general appearance and the vegetation cover. It was shown that when a specific width of the trench is occupied by deposits which according to the previous classification are very recent, the torrent should be viewed as active. Where the occupied trench width exceeds a certain limit, the torrent is in a high risk active condition.

INTRODUCTION

The main depositional area of torrential streams is located below the gorge section. The continuous depositional activity of the torrent in this area produces an alluvial fan. If we observe the main body of such features in the field we will see that they are traversed by one or more trenches. According to Bull (1962, 1964) and Wasson (1977) these trenches are formed by the flood waters of the torrent stream after the main body of the fan has been formed. Their width and depth varies considerably. Bull (1962) reports depths from 1 to 15 m and widths from several metres up to several tens of metres. Within the depositional zone, the entire torrent activity is concentrated within the trenches. By the term activity we mean any process which contributes to the formation or destruction of the main body of the landform. The total width which the trenches occupy on the surface of the depositional landform can be considered to represent the zone of recent activity (z.r.a.).

Wasson (1977), considers that if deposition prevails within these trenches, the depositional landform is considered to be active, because its structure is being "supplemented" by new transported material (fan entrenchment-active situation). If erosion phenomena are observed within the trenches and the previously deposited material is being removed, the landform is considered to be in an inactive state which leads to its destruction (fan incision-inactive situation). A sample of torrents and their respective deposition landforms was used to investigate further the suggestions of Wasson (1977).
The following aspects were considered:
(a) the form of the deposits and the fraction of the total width of the trenches (i.e. z.r.a.) occupied by the deposits in the case of active fans;
(b) the form of the deposits and their width in the case of inactive fans.

FIELD INVESTIGATIONS

As noted above, Wasson (1977) considers that the depositional landform should be viewed as active when there is material accumulating within the trenches. However, the presence of debris within the trenches is not in itself sufficient evidence of recent\(^1\) transport. This is because the material may have been there for several decades, in the absence of significant floods capable of remobilizing and transporting the debris. It is therefore useful in the first instance to determine whether the material accumulated within the trenches was deposited recently or whether it is a product of older torrent activity. This has led us to attempt a general classification of the debris found within the trenches and forming the main body of the fan, based on the date of its transport and deposition.

The classification is based on two external morphological features which are easy to identify in the field. The first is the vegetation growing on the surface of the deposits and the second is their surface appearance. This approach was used to investigate 29 alluvial fans scattered throughout the

<table>
<thead>
<tr>
<th>No.</th>
<th>Torrent stream</th>
<th>Drainage area (km(^2))</th>
<th>No.</th>
<th>Torrent stream</th>
<th>Drainage area (km(^2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Vatonias</td>
<td>236.46</td>
<td>16</td>
<td>Moustheni</td>
<td>12.40</td>
</tr>
<tr>
<td>2</td>
<td>Vogeni</td>
<td>2.30</td>
<td>17</td>
<td>Nea Apollonia</td>
<td>232.90</td>
</tr>
<tr>
<td>3</td>
<td>Volinaios</td>
<td>27.00</td>
<td>18</td>
<td>Douvias</td>
<td>8.00</td>
</tr>
<tr>
<td>4</td>
<td>Vyrhon</td>
<td>23.10</td>
<td>19</td>
<td>Xerias</td>
<td>15.30</td>
</tr>
<tr>
<td>5</td>
<td>Zagliveri</td>
<td>126.80</td>
<td>20</td>
<td>Paleovraha</td>
<td>8.50</td>
</tr>
<tr>
<td>6</td>
<td>Therni</td>
<td>52.80</td>
<td>21</td>
<td>Platanias</td>
<td>8.50</td>
</tr>
<tr>
<td>7</td>
<td>Thermopyles</td>
<td>8.80</td>
<td>22</td>
<td>Poroi</td>
<td>96.00</td>
</tr>
<tr>
<td>8</td>
<td>Kamenikia</td>
<td>32.50</td>
<td>23</td>
<td>Portaikos</td>
<td>137.55</td>
</tr>
<tr>
<td>9</td>
<td>Karpenisi</td>
<td>3.00</td>
<td>24</td>
<td>Skalas</td>
<td>33.70</td>
</tr>
<tr>
<td>10</td>
<td>Kerinitis</td>
<td>80.00</td>
<td>25</td>
<td>Ipati</td>
<td>28.40</td>
</tr>
<tr>
<td>11</td>
<td>Kouvelorrema</td>
<td>12.42</td>
<td>26</td>
<td>Phoinikas</td>
<td>27.00</td>
</tr>
<tr>
<td>12</td>
<td>Marathias</td>
<td>33.00</td>
<td>27</td>
<td>Halandritsa</td>
<td>11.70</td>
</tr>
<tr>
<td>13</td>
<td>Meganitis</td>
<td>60.08</td>
<td>28</td>
<td>Haradros</td>
<td>19.50</td>
</tr>
<tr>
<td>14</td>
<td>Megas</td>
<td>4.70</td>
<td>29</td>
<td>Hourou</td>
<td>7.70</td>
</tr>
<tr>
<td>15</td>
<td>Messoropi</td>
<td>15.40</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

\(^1\)The term "recent" is used not in its classic meaning but with the meaning attributed by some researchers (recent (Wasson), Jung (Aulitzky)) in order to characterize torrent activity which is recent in relation to the longer-term activity of a torrent.
Criteria for determining the current activity of torrents

<table>
<thead>
<tr>
<th>Deposit category</th>
<th>Vegetation description</th>
<th>Surface appearance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Category A</td>
<td>No trace of vegetation found</td>
<td>Moveable pebbles without morphologic, chromatic, mechanical, or chemical alteration</td>
</tr>
<tr>
<td>Category B</td>
<td>Very slight traces of vegetation in the form of weak forbs</td>
<td>Superficial pebbles linked together by matrix material (e.g. loam). Less moveable</td>
</tr>
<tr>
<td>Category C</td>
<td>Some forb vegetation. Rarely small trees or shrubs</td>
<td>Pebbles with advanced chromatic and morphologic alteration; more or less joined together</td>
</tr>
<tr>
<td>Category D</td>
<td>Shrubby or woody vegetation, depending on the plant association. Occasionally cultivated</td>
<td>Surface covered by a soil layer. Only a few pebbles visible on the surface</td>
</tr>
<tr>
<td>Category E</td>
<td>Shrubby or woody vegetation sometimes with a coverage of 100%; or systematically cultivated fields</td>
<td>Except for a few gravel particles no pebbles appear on the surface. Difficult to distinguish from soils of non alluvial origin.</td>
</tr>
</tbody>
</table>

Greek mainland and having basin areas ranging between 2.5 and 230 km² (see Table 1). The deposits which make up the main body of the depositional landform and those which fill the trenches can be subdivided into five categories according to their residence time at the location where they were last deposited (see Table 2 and Figs 1-4).

The width of the deposition zone

Our research also confirmed that the width of the deposition zone of category A deposits inside the trenches corresponds to the width occupied by the annual flood associated with the respective torrent. This conclusion was based on the
following evidence.

In every alluvial fan included in the sample, a randomly selected cross section $E_1E_2E_3$ was defined. Within this cross section the width occupied by category $A$ deposits and the height of the respective alluvial terraces up to the point where category $B$ deposits appear were measured (Fig. 5).

Thus, we calculated the area $F$ of the cross section $K_1K_2K_3K_4$ and using the formula $R = F/P$, we calculated its hydraulic radius ($R$). The axial slope ($J$) of the cross section was also measured. By using these values in the Manning-Strickler formula we estimated the peak flow velocity $U_{\text{max}}$ at this point, i.e.

$$U_{\text{max}} = KR^{2/3} J^{1/2}$$

(1)
Criteria for determining the current activity of torrents

Fig. 4 Detail of category C deposits; adequate vegetation, pebbles are linked together.

where:
- $K =$ the roughness coefficient;
- $R =$ the hydraulic radius defined as $R = F/P$ where $F$ is the cross-sectional area and $P$ the wetted perimeter;
- $J =$ the axial slope of the channel measured locally.

The magnitude of the annual flood associated with each study torrent was estimated using the formula of Fuller as given by Kotoulas (1979) which estimates the maximum flood for a specific frequency in years i.e.

$$Q_{\text{max}} = Q_1 (1 + b_{\log_{10} T}) \left[1 + \frac{2.66}{F^{0.30}}\right]$$

(2)

where $Q_1 =$ the mean flood discharge with a frequency of one year (m$^3$ s$^{-1}$).

This is calculated as:

$$Q_1 = 1.80 F^{0.8}$$

where $F =$ the area of basin (km$^2$); and $T =$ frequency or return period, calculated as follows:

<table>
<thead>
<tr>
<th>Frequency (years)</th>
<th>1</th>
<th>5</th>
<th>10</th>
<th>20</th>
<th>50</th>
<th>100</th>
</tr>
</thead>
<tbody>
<tr>
<td>Parameter</td>
<td>1 + $b_{\log_{10} T}$</td>
<td>1.000</td>
<td>1.559</td>
<td>1.800</td>
<td>2.40</td>
<td>2.359</td>
</tr>
</tbody>
</table>

Thus, in this case the formula takes the form:
After the flood discharge were estimated a check was undertaken to determine the size of cross section needed to contain the annual flood discharge. This check indicates that for 27 out of the 29 alluvial fans included in the sample, the required capacity was approximately equal to the area of the cross section $K_1K_2K_3K_4$ (Fig. 5). Table 3 lists the areas of the cross sections as measured in the field, and the capacities required to contain the annual flood.

Table 3 confirms the validity of the above conclusions. In addition, since the flow depth associated with the annual flood does not differ significantly from the height $K_1K_4$ measured in the field, the width of the flood cross section can be taken as equal to the width $K_1K_2$. In other words, the bed width occupied by the annual flood is covered by category $A$ deposits. This is also the reason why for category $A$ deposits there is a complete absence of vegetation. The annual disturbance by the flood prevents the growth of vegetation. Category $A$ deposits thus have a residence time ranging from a few days up to several months.

The structure of the deposits

Category $A$ deposits are sometimes well stratified with a smooth surface and sometimes unstratified with an irregular wave-like surface. These two types of deposit may be found at different positions within the same trench. This phenomenon is obviously related to the shear stress of the water which in turn depends on the bed slope. Wave-like surfaces were observed in reaches with increased slopes, whereas the surfaces were smoother where axial slopes were less and widths greater. Field observations of the beds of the trenches have

\[ Q_{1\text{max}} = Q_1 \left(1 + \frac{2.66}{70.30}\right) \text{ m}^3 \text{ s}^{-1} \]
Table 3 Measured trench cross section areas and required capacities for the annual flood.

<table>
<thead>
<tr>
<th>No.</th>
<th>Torrent stream</th>
<th>Measured trench area (m²)</th>
<th>Required trench capacity (m²)</th>
<th>No.</th>
<th>Torrent stream</th>
<th>Measured trench area (m²)</th>
<th>Required trench capacity (m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Vatonias</td>
<td>141.40</td>
<td>143.6</td>
<td>16</td>
<td>Moustheni</td>
<td>7.40</td>
<td>6.71</td>
</tr>
<tr>
<td>2</td>
<td>Vogeni</td>
<td>2.45</td>
<td>2.15</td>
<td>17</td>
<td>Nea Apollonia</td>
<td>155.50</td>
<td>173.44</td>
</tr>
<tr>
<td>3</td>
<td>Volinaios</td>
<td>13.40</td>
<td>14.21</td>
<td>18</td>
<td>Douvias</td>
<td>7.00</td>
<td>5.10</td>
</tr>
<tr>
<td>4</td>
<td>Vryhon</td>
<td>14.80</td>
<td>12.86</td>
<td>19</td>
<td>Xerias</td>
<td>8.50</td>
<td>8.24</td>
</tr>
<tr>
<td>5</td>
<td>Zagliveri</td>
<td>110.80</td>
<td>116.96</td>
<td>20</td>
<td>Paleovraha</td>
<td>4.60</td>
<td>4.41</td>
</tr>
<tr>
<td>6</td>
<td>Thermi</td>
<td>32.10</td>
<td>55.27</td>
<td>21</td>
<td>Platanias</td>
<td>5.10</td>
<td>5.95</td>
</tr>
<tr>
<td>7</td>
<td>Thermopyles</td>
<td>6.90</td>
<td>5.80</td>
<td>22</td>
<td>Poroi</td>
<td>43.20</td>
<td>48.25</td>
</tr>
<tr>
<td>8</td>
<td>Kamenikia</td>
<td>35.8</td>
<td>31.25</td>
<td>23</td>
<td>Portaikos</td>
<td>49.40</td>
<td>52.84</td>
</tr>
<tr>
<td>9</td>
<td>Karpenisi</td>
<td>2.40</td>
<td>2.21</td>
<td>24</td>
<td>Skalas</td>
<td>18.60</td>
<td>15.57</td>
</tr>
<tr>
<td>10</td>
<td>Kerinitis</td>
<td>39.40</td>
<td>33.06</td>
<td>25</td>
<td>Ipati</td>
<td>13.75</td>
<td>10.74</td>
</tr>
<tr>
<td>11</td>
<td>Kouvelorrema</td>
<td>14.10</td>
<td>8.172</td>
<td>26</td>
<td>Phoinikas</td>
<td>37.50</td>
<td>38.92</td>
</tr>
<tr>
<td>12</td>
<td>Marathias</td>
<td>15.60</td>
<td>13.89</td>
<td>27</td>
<td>Halandritsa</td>
<td>7.10</td>
<td>7.89</td>
</tr>
<tr>
<td>13</td>
<td>Meganitis</td>
<td>38.40</td>
<td>32.45</td>
<td>28</td>
<td>Haradros</td>
<td>14.00</td>
<td>10.38</td>
</tr>
<tr>
<td>14</td>
<td>Megas</td>
<td>2.90</td>
<td>2.71</td>
<td>29</td>
<td>Hourou</td>
<td>2.60</td>
<td>2.70</td>
</tr>
<tr>
<td>15</td>
<td>Messoropi</td>
<td>4.95</td>
<td>6.97</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

shown that for all the alluvial fans included in the sample, wave-like surfaces occurred where the axial inclination exceeded 3% and smooth surfaces occurred where the inclination was less than 3%.

Residence times of the deposits The deposits of categories B, C, D and E upon which various plant species grow must clearly remain undisturbed for substantial periods of time. More specifically, the field observations demonstrated that the longer their residence time at a particular position the more advanced is the colonisation by pioneer species, initially, and woody species subsequently. In the case of categories B, C, D and E deposits, it is not possible to estimate the exact residence time, as in the case of category A deposits. However, we may safely say that category B deposits have residence times in excess of 5-10 years.

In conclusion, it is considered useful to emphasise the following points, in relation to the findings outlined above:

(a) In fully-developed alluvial fans we normally encountered deposits in all five categories. Usually, they are distributed as follows. Category A, B and C deposits are found inside the trenches. In rare cases it is also possible to find category D deposits in such locations. On the contrary, the deposits found outside the trenches (which also constitute the main body of the fan) are principally of categories D and E.

(b) The boundaries between the five categories are not always clearly defined in nature.

(c) Very often, although there is a suitable soil layer (category D or E deposits) woody species may not be present, and only evergreen...
broadleaved shrubs are found due to the local vegetation associations. In very dry-warm climates even evergreen broadleaved shrubs may have a limited presence.

(d) Sometimes, when flood waters overflow onto areas with category D and E deposits, they may leave pockets of category A deposits overlying category D and E deposits. In this case, care is required in applying the classification, because macroscopically the deposits may be classified as D and E whereas at a more detailed level they may be classified as A.

(e) All the above soil-vegetation relationships were developed in Greece. It is believed that the same relationships will apply in other countries with a similar climate. Additional research may be necessary to extend the approach to other regions.

RESULTS

Using the findings outlined above and by undertaking detailed investigations of the depositional landforms included in the sample of alluvial fans, we concluded that only deposits in the first two categories of the classification can be considered to provide evidence of a dynamic condition. Deposits in the next category show features and characteristics which indicate that the materials are not moving.

We have previously classified the depositional landforms in the sample as being active or inactive based on the conclusions of Wasson (1977) (see Table 4). Table 4 also presents the proportions of length and width of the trenches occupied by category A and B deposits. By comparing the two sets of information we find that all the depositional landforms which are in an active condition according to Wasson (1977) have category A and B deposits occupying over 30% of the width of the z.r.a. and its entire length. The remaining examples are all inactive.

Furthermore, based on field observations throughout Greece and on the reports of the Forest Offices and local authorities etc., we found that many alluvial fans had category A and B deposits covering 80% of the width of the z.r.a. These were characterized by frequent flood events and severe disasters during the last decade.

Thus, we come to the general conclusion that in cases where category A and B deposits occupy at least 33% of the average width of the zone of recent activity (z.r.a.) along its entire length, the zone is functioning at the present time and the alluvial fan is in an active condition. Conversely, where the z.r.a. is not functioning at present the alluvial fan is inactive. More specifically, those fans where category A and B deposits cover more than 80% of the width of the z.r.a. should not only be considered as being in an active condition but also in a dangerously active condition.
Table 4  Percentages of the total width and length of the z.r.a. occupied by Category A and B deposits compared to the activity classification based on Wasson (1977).

<table>
<thead>
<tr>
<th>No. Torrent stream</th>
<th>No. Torrent stream</th>
<th>Total percentage of z.r.a. width and length, Deposits A and B:</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vatonias</td>
<td>Mouatheni</td>
<td>Width 40% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 90% per locations only inactive</td>
</tr>
<tr>
<td>Vogeni</td>
<td>Nea Apollonia</td>
<td>Width 95% on entire length active</td>
</tr>
<tr>
<td>Voiniaios</td>
<td>Douvias</td>
<td>Width 60% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 70% on entire length active</td>
</tr>
<tr>
<td>Vryhon</td>
<td>Xerias</td>
<td>Width 55% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 80% on entire length active</td>
</tr>
<tr>
<td>Zagliveri</td>
<td>Paleovraha</td>
<td>Width 30% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 20% per locations only inactive</td>
</tr>
<tr>
<td>Thermi</td>
<td>Flataniyas</td>
<td>Width 20% on entire length inactive</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 90% on entire length active</td>
</tr>
<tr>
<td>Thermopyle</td>
<td>Poroi</td>
<td>Width 10% on entire length inactive</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 70% on entire length active</td>
</tr>
<tr>
<td>Kamenicka</td>
<td>Portaikos</td>
<td>Width 30% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 80% on entire length active</td>
</tr>
<tr>
<td>Karpeniki</td>
<td>Skulas</td>
<td>Width 80% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 80% on entire length active</td>
</tr>
<tr>
<td>Kerinitis</td>
<td>Ipati</td>
<td>Width 75% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 70% on entire length active</td>
</tr>
<tr>
<td>Koulouvelorrum</td>
<td>Floinikas</td>
<td>Width 80% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 80% on entire length active</td>
</tr>
<tr>
<td>Marathias</td>
<td>Halandritsa</td>
<td>Width 90% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 70% on entire length active</td>
</tr>
<tr>
<td>Meganitis</td>
<td>Haradros</td>
<td>Width 95% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 90% on entire length active</td>
</tr>
<tr>
<td>Megas</td>
<td>Hourou</td>
<td>Width 80% on entire length active</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Length 80% on entire length active</td>
</tr>
</tbody>
</table>

*Based on Wasson (1977).

CONCLUSIONS

On the basis of the above work it is possible to classify the deposits of an alluvial fan into categories of various ages based on their major morphological features. From the viewpoint of risk and protection what interests the torrent manager is the presence and mobility of the more recent deposits within the trenches of alluvial fans, since their remobilization can cause further damage. We believe that simply by studying the nature and distribution of the recent deposits it is possible to determine whether a trench is active and the degree of risk associated with this activity. For this reason, it is hoped that the results presented in this paper may be of more general interest.

REFERENCES

A conceptual geomorphological model for the development of a Mediterranean river basin under neotectonic stress (Buonamico basin, Calabria, Italy)

P. ERGENZINGER

Institut für Physische Geographie, Universität Berlin, Grunewaldstr. 35, 1000 Berlin 41, Germany

Abstract

The Buonamico River basin in Calabria, southern Italy is part of the active collision belt between the African and European plates. It drains to the southeast of the Aspromonte Mountains of the Calabrian Massif of basement rocks, from an elevation of 1956 m down to the Ionian Sea. During Pleistocene uplift of the Aspromonte, rivers incised intensively, causing massive amounts of erosion on the corresponding valley slopes. From long-term investigations of active present day fluvial processes, a conceptual model of the interrelationships between mass movements and debris flows created on the slopes, and their effects on the development of the local river systems is proposed. Amongst the main controls of the development of the valley basin are the availability of erodible material from the slopes, and a river system which is capable of adjusting to both violent tectonic impacts and extremely variable situations of sediment input from the adjacent slopes.

INTRODUCTION AND REGIONAL SETTING

Study area

The Buonamico River basin is situated in the Aspromonte Mountains of Calabria, southern Italy (Fig. 1), which are part of the active collision belt between the African and European plates. There is still considerable speculation as to their tectonic setting (cf. Ibekken & Schleyer, 1991, p.11) with theories ranging from inter- or intra plate deformation, to subduction or a deep-seated shear zone. The Aspromonte Mountains are part of the Calabro-Peloritan arc, a region with the highest tectonic activity in Europe. During Pleistocene uplift, the Aspromonte Mountains were raised by 1100-1300 m. Rivers still incise intensively, causing massive amounts of erosion in the valleys and on their slopes. Lembke (1931) first attempted to describe the tectonics and geomorphology of the region.

The Buonamico basin has an area of 139 km², of which approximately 75% lies in the Aspromonte mountains. Monte Alto forms the highest point with an altitude of 1955 m. The basin is underlain by schists (54%), and gneiss (13%) whilst the border and foreland zone consists of sandstone 17%, siltstone, conglomerate 3% and argillite 3%. 

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The Buonamico basin has an area of 139 km², of which approximately 75% lies in the Aspromonte mountains. Monte Alto forms the highest point with an altitude of 1955 m. The basin is underlain by schists (54%), and gneiss (13%) whilst the border and foreland zone consists of sandstone 17%, siltstone, conglomerate 3% and argillite 3%.
The topography was digitized using a 250 × 250 m grid system (Ibekken & Schleyer, 1991). The differentiation of altitude and gradient as a function of distance from the river mouth is displayed in Fig. 2. Maximum differences in altitude of up to 1000 m occur in the middle part of the longitudinal profile which coincides with steep slopes of about 30°. Erosion rates were calculated using the approach of Görler & Uchdorf (1980). The topography before uplift and the volume subsequently eroded from the valleys were determined. In Fig. 2 the maximum erosion rates occur within the valley reach that lies halfway between the Mont Alto and the Ionian Sea. Total amounts of erosion can thus be calculated but not the short term relationship between slope development and the river bed. The short term functional relationship (Ahnert, 1970) between denudation, relief and uplift will be investigated in the area of the Buonamico basin, a small Mediterranean basin under neotectonic stress.

Aims

The aims of the study are to combine the processes of slope and river development under unusual neotectonic conditions within a conceptual model. Long-term steady state conditions cannot exist on the river bed because of episodic delivery of material from the slopes by mass movement and the many decades required to transport this input through the fluvial system into the sea. The different time-scales over which slope development and river development proceeds are investigated. Magnitude-frequency distributions of slope-and-river associated events differ substantially (Ergenzinger, 1988). The formulation of
special feedback mechanisms controlling the interaction of the system is therefore necessary for the model.

Methodology

The distribution and volumes of mass movements were determined by aerial photo interpretation for 1941, 1955 and 1973. River cross profiles were measured annually between 1972 and 1980 and again in 1986. The longitudinal river profile was surveyed in 1980 which allowed comparisons with the topographic maps of the Instituto Geografico Militare at 1:10 000.

On 4 January 1973, a lake was created by the Costantino landslide in the middle reaches of the Buonamico mountain valley. Since this event, the bed load transported from this part of the basin and deposited in the lake delta has been evaluated on a year by year basis. In the period between the winters of 1978 and 1980, hydrology and solid material budgets of the lake drainage basin were measured on an event basis.

Precipitation was measured from 1927 to 1972 at the Santuario di Polsi in the uppermost part of the catchment. On the Ionian side of the Aspromonte there is no other comparable climatological station in the higher parts of the mountain. A ground penetrating shallow seismic technique was used in order to determine the Holocene fluvial infill in the lower valley and the depth of the weathered mantle on the slopes.

Discussion

Slope development in the Buonamico basin is dominated by mass movements. Only a small part of the uppermost catchment is without active mass movements. The distribution of the weathering mantle is very uneven and location specific. Resistant rocks are to be found only along the divides and on
the lower parts of the slopes. As was shown by Mouton in Italpros (1986) there can be up to 40 m of disintegrated material on the middle parts of the slopes (Fig. 3). This material is cohesionless, without interstitial fill, whereas the scree close to the surface consists of a silty matrix with unevenly distributed stones. For this type of scree, layers of stratified stones are typical due to the influence of running water, yet in all other respects they are similar to periglacial screes.

![Fig. 3 Depths and travel times (km⁻¹) across the Amendolea River along a 2 km section.](image)

Deep disintegration has most probably been caused by earthquake activity which is also of fundamental importance in the initiation of mass movements. The disintegration mantle can be found wherever there are traces of former mass movements, but it is uncertain whether this mantle was created by, or was a prerequisite for, the mass movements. Regardless of this question, it is clear that the entire slope dynamics are controlled by the availability of disintegrated slope material.

In contrast to the slope dynamics associated with mass movements, the impact of material eroded by fluvial processes is rather insignificant and will therefore not be treated in this paper.

According to Terzaghi (1960), landslides are controlled by both internal and external factors. External causes, resulting in an increase in shear stress, are created by loading (e.g. due to material loading or the storage of water), earthquake shocks or undercutting of slopes. Internal causes lead to a decrease in shear resistance by pore water pressure or to a decrease in cohesion of the slope material.

During the last half century all large mass movements in the Buonamico basin occurred in response to extreme precipitation events. Caloiero & Mercuri (1980) summarized the influence of large precipitation events occurring from 1921 to 1970, whilst Cotecchia et al. (1969) described the impact of the extreme earthquake of 1783 on the local geomorphology. If all precipitation events with more than 100 mm are added together for an entire winter season, this sum correlates well with the occurrence of mass movements. Thus whenever more than 1200 mm of accumulated precipitation occur, there is a very high probability of the occurrence of a landslide (Fig. 4(a)).

The size and volume of recent landslides were determined by aerial photo
Geomorphological model for a mediterranean basin under neotectonic stress

Fig. 4 (a) Annual extreme precipitation: summary of all precipitation events above 100 mm day$^{-1}$. Volume of landslide material transported (b) to the valley bottom, and (c) by the Buonamico.

interpretation (Fig. 5). The resulting total volumes associated with different time periods are shown in Table 1. The sediment input from mass movements into the river has been estimated as 20% of the total volume. This is a conservative estimate.

The availability of material for transport by the river is governed by the input of material from mass movements. These inputs are of low frequency and high magnitude. In years between the mass movements with low flood discharges (i.e. under high frequency conditions), the transported volumes are of the order of $10^3$ or $10^4$ m$^3$. Figure 6 attempts to demonstrate the distribution of magnitude/frequency for the slope processes (cf. Wolman & Miller, 1960). Magnitudes and frequencies of the related fluvial transport will be discussed later.
River bed development is a reaction to solid material transport and episodic winter floods associated with high precipitation. Since the river is rather steep, suspended load is more or less directly transported into the Ionian Sea, whereas bed load is transmitted step by step through the braided system of the Buonamico foreland. The long term geological development of the river bed can be represented from comparisons for the period 1955-1980 (Fig. 7). More than 4 million m$^3$ of sediment were deposited during this time interval. The Costantino landslide in 1953 caused a large input of coarse material at km 14. Below the landslide the narrow mountain valley was filled with 50 m of sediment. The largest amount of material is nevertheless not to be found in the mountain areas, but in the foreland where average river widths increase up to 800 m (km 7). Between 1955 and 1980 erosion occurred close to the sea where the Buonamico was dammed and widths were reduced to 200 m. Even though seven active winter seasons had passed after the landslide, the former river bed conditions could not be restored. The input of more than 2 million m$^3$ of material from the Costantino landslide by fluvial erosion created a far too great

<table>
<thead>
<tr>
<th>Aerial photo from</th>
<th>Landslides from</th>
<th>Landslide area (km$^2$)</th>
<th>Volume ($\times 10^6$ m$^3$)</th>
<th>Input into the river ($\times 10^6$ m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1941</td>
<td>1933</td>
<td>2.1</td>
<td>21</td>
<td>4</td>
</tr>
<tr>
<td>1955</td>
<td>1951</td>
<td>4.3</td>
<td>43</td>
<td>9</td>
</tr>
<tr>
<td>1973</td>
<td>1958, 1973</td>
<td>3.0</td>
<td>30</td>
<td>6</td>
</tr>
<tr>
<td>last 50 years</td>
<td></td>
<td>9.4</td>
<td>94</td>
<td>19</td>
</tr>
</tbody>
</table>
disturbance of the river bed conditions. The development of the river bed was surveyed on a yearly basis with a set of cross profiles. Between 1972 and 1986, only part of the input of the extreme event of 1973/1973 was eroded. (Fig. 8).

The magnitude/frequency distribution for fluvial transport differs markedly from that of the slopes (Fig. 6). Based on extrapolation of the amount

![Magnitude/frequency diagram for slope and river events in the Buonamico basin.](image)

![The Buonamico long profile: (a) comparison and (b) gradient of the long profiles of 1955 and 1980; (c) net volume differences between 1955 and 1980.](image)
of solid material deposited in the Lago Costantino above the landslide, sediment transport over the last two decades averaged about 10,000 m$^3$ per year. The maximum amount of solid material transported in 1972/1973 is estimated at between 5-9 million m$^3$. The recovery time required to erode and transport material from the last major landslide is at least 20-25 years. This is comparable to the observations of Wolman & Gerson (1978) on semiarid rivers.

As Scheidegger (1991, p.110) stated, endogenic uplift must be compensated by downhill mass movement, but large amounts of mass movement require the evacuation of the infilled material by running water. Although extreme precipitation events cause both extreme landslides and floods, the amount of material transported into the valley bottom according to our observations is several times greater than the transport capacity of the related floods.

The river system therefore requires several decades before it can re-establish the former level of the river bed. During this time interval, the lower slopes along the infilled parts of the valley are protected against undercutting. Only where there is a valley zone with intense incision is there a high probability for new landslides created by extreme precipitation. Thus infilling and undercutting form the main feedback mechanisms between the slope and river bed dynamics.

**CONCLUSIONS**

The development of slopes and fluvial systems proceeds over different time-scales and with different magnitudes and frequencies in the Buonamico basin. Especially under conditions of neotectonic uplift, the coupling of the two systems is dictated by the longitudinal development of the river profile. The development of the longitudinal profile is affected by both the tectonic uplift

![Fig. 8 Changes in cross-section area at three sites between the Aspromonte and the sea.](image-url)
and the local input of large amounts of material by mass movements. The results of the investigations in the Buonamico basin have been drawn together in a conceptual geomorphological model (Fig. 9).

Seismic events and/or precipitation events cause vulnerable material on the steep slopes along the middle part of the valley to move by sliding. Triggered by extreme precipitation, landslides can occur approximately every 20-25 years, whereas earthquake events with a Richter scale magnitude above 8 have a recurrence interval of 100 years. In the Vallone Avrea, the remnants of such a huge event can be seen in a debris flow with a length of more than 2 km, a thickness of 50 m and a volume of $60 \times 10^6 \text{ m}^3$. During this Holocene event, the river was incapable of evacuating all the material.

River bed dynamics operate at different time-scales. Indeed the river needs decades to overcome the local input of coarse material and to transmit this into the Ionian Sea. The difference between the reaction of the river system and the slope system is shown in Fig. 10. The volumes of sediment transport by river and landslides for various magnitudes differ considerably. Whereas the rivers dominate the low magnitude event field, mass movements dominate the high magnitude field.

Fig. 9 Conceptual geomorphological model of the interrelationships between slope and fluvial development.
Fig. 10 Volumes of sediment transported by rivers and landslides in different magnitude classes.

Acknowledgements Field investigations were supported by the DFG and the Free University of Berlin. I am grateful to P. Obenauf, H. Ibbeken, O. Hirsch and a great number of students for assistance in the field. Thanks go to the cartographers J. Schulz and R. Willing. I am also indebted to C. de Jong for translating and typing.

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REFERENCES

The use of fallout radionuclides in investigations of erosion and sediment delivery in the Polish Flysh Carpathians

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Abstract The Polish Flysh Carpathians are a mountain area where climate, relief and the erodible bedrock combine to produce an active erosional system dominated by fluvial processes. Poor land management has further intensified land erosion. An improved knowledge of the erosion and sediment delivery dynamics of the area is required to provide a basis for improved land management and for reducing rates of reservoir siltation. Classical methods for investigating rates of erosion and sediment delivery dynamics possess many limitations in terms of operational problems, the substantial resources required and their limited spatial and temporal coverage. The use of the fallout radionuclide caesium-137 as a sediment tracer appears to offer considerable potential for assembling information on erosion and sediment delivery dynamics in this environment and some preliminary results of radiocaesium-based investigations undertaken in the small (19.6 km$^2$) Homerka drainage basin and the larger (4692 km$^2$) basin of the Dunajec River above Roznowski reservoir are reported. These investigations have focussed on assessment of soil erosion on cultivated fields, identification of the main suspended sediment sources and elucidation of sediment delivery dynamics.

INTRODUCTION

The Polish Flysh Carpathians are a mountain area of moderate altitude and relief where the tree cover extends to the summits. Much of the original forest cover has been removed by deforestation. The headwater zones and steep slopes are characterized by intensively exploited forests accessed by a dense network of unmetalled roads. In the lower parts much of the land is given over to arable cultivation. This land is divided into small plots bounded by terraces and is again characterized by a dense network of unmetalled roads, which frequently extend to the stream channels. The wetter areas at the base of the slopes and on the valley floors are occupied by meadows and pasture. The relatively high annual precipitation of 1000-1500 mm falls primarily during the summer months. During storm events, the dense network of unmetalled roads
and the numerous gullies promote rapid surface runoff which transports large amounts of sediment to the channels (cf. Froehlich, 1991). The critical threshold for the widespread occurrence of surface runoff is a storm rainfall of c. 20 mm (Slupik, 1973).

The area is characterized by highly active erosion, sediment transport and fluvial sedimentation processes, which in turn reflect the climate, the high relief energy, the erodible nature of the soils and rock and the effects of poor land management. Fluvial processes are dominant, and the channel network is being actively deepened in many areas. Locally, mass movements assume considerable importance (cf. Starkel, 1972). Measurements obtained from runoff plots located on cultivated slopes within the area point to high rates of soil erosion under certain crops. Values for potatoes are as high as 22 t ha\(^{-1}\) year\(^{-1}\), whilst typical values for winter crops, meadows and forest are 2.4 t ha\(^{-1}\) year\(^{-1}\), 0.1 t ha\(^{-1}\) year\(^{-1}\) and 0.03 t ha\(^{-1}\) year\(^{-1}\) respectively. Sediment yields from the larger river basins are in the range 90-1000 t km\(^{-2}\) year\(^{-1}\) (cf. Lajczak, 1988). It is, however, difficult to make direct comparisons between estimates of the intensity of erosion processes on the slopes and the sediment yields of Carpathian rivers, because of the wide range of techniques of unknown accuracy and precision which have been used and the different periods of record involved. Further integrated studies are required in order to obtain a clear appreciation and understanding of the erosion and sediment delivery dynamics of this region and of the role of land use and human activity in disturbing the natural conditions. Improved understanding of the spatial variability of soil erosion on slopes, of sediment sources and of sediment supply dynamics is essential for protecting soils against erosion and reducing rates of reservoir siltation.

Classical (standard) methods for investigating erosion and sediment delivery are for the most part expensive to apply, in terms of both equipment and manpower requirements and the timescales involved. As a result, most studies applying such methods have addressed very limited objectives, as well as covering only small areas and involving short periods of record. In consequence, it has proved difficult to produce meaningful assessments of the pattern of spatial variability of the processes involved. The high energy and active morphodynamic environments associated with mountain areas also introduce important technical constraints in the application of traditional classical techniques. It is therefore difficult to relate measurements of soil loss at different slope positions to sediment yield. In addition, it is difficult to use short-term, site-specific measurements in the interpretation of longer-term contemporary relief evolution. Little is currently known about the residence times of sediment particles moving through the fluvial system of drainage basins of different scales.

In recent years, geomorphologists have shown increasing interest in the use of the fallout radionuclide caesium-137 (\(^{137}\)Cs) as a sediment tracer (e.g. Campbell et al., 1982; Walling & Bradley, 1990). Caesium-137 was introduced into the stratosphere by the atmospheric testing of nuclear weapons during the
The use of fallout radionuclides in the Polish Carpathians

middle and late 1950s and the 1960s and the fallout, which was primarily associated with precipitation, occurred globally. In some areas, additional $^{137}$Cs fallout also occurred immediately after the Chernobyl disaster in 1986. In most environments, radiocaesium reaching the land surface is rapidly adsorbed by the upper horizons of the soil. Subsequent movement occurs in association with sediment particles and $^{137}$Cs therefore provides a very effective sediment tracer. Measurements of the redistribution of "bomb-derived" radiocaesium during the period since the fallout occurred offer a means of assessing the movement of sediment within the landscape over a timescale of 25-30 years. Where Chernobyl-derived fallout also occurred, shorter-term assessments may also be obtained.

The use of $^{137}$Cs as a tracer therefore affords a valuable means of investigating the mobilization of sediment and its transfer through the fluvial system over timescales of several decades and over a range of spatial scales, and therefore overcomes many of the limitations of classical monitoring techniques (cf. Walling, 1990). Caesium-137 measurements can be used as a basis for studying both the magnitude and spatial variability of rates of soil loss, identifying sediment sources, establishing sediment delivery ratios and assessing rates and patterns of sedimentation on alluvial fans and flood plains and in reservoirs (cf. Loughran et al., 1982; Sutherland & de Jong, 1990; Walling & Bradley, 1988; Walling & Quine, 1991; Walling et al., 1992). To date, however, the application of $^{137}$Cs measurements to investigations of erosion and sediment delivery has been largely restricted to areas of limited relief in lowland areas. Relatively little work has been undertaken in mountain areas, where altitudinal variations in precipitation, the importance of snow cover and the high energy environment necessitate some modifications to the approach.

This contribution presents a preliminary assessment of the potential for using $^{137}$Cs measurements, as an alternative and complement to classical monitoring techniques, to investigate erosion and sediment delivery processes in the Polish Flysch Carpathians and in mountain areas more generally. The work reported was undertaken in the Homerka experimental catchment where classical monitoring techniques have also been applied over the past 15 years (cf. Froehlich, 1982). The availability of data for this basin obtained using classical methods provides a basis for testing the consistency of the two approaches and for exploring the possibility of using the $^{137}$Cs approach to extrapolate site-specific results obtained by traditional means.

THE STUDY AREA

The 19.6 km$^2$ drainage basin of the Homerka stream (Fig. 1) lies at an altitude of 375-1060 m above sea level and is typical of the largely deforested landscape of this region. Its upper part is characterized by straight or convex slopes and deeply incised V-shaped valleys developed on resistant
Fig. 1 The Homerka basin and the location of the study area in Poland.
Magura sandstones. These areas lie in the lower subalpine zone and exhibit skeletal soils. The forests which occur in the headwaters account for 52% of the total area of the basin and are at present intensively exploited. They are associated with a dense network of unmetalled roads and timber transport trails. The agricultural areas are concentrated in the lower parts of the basin where the relief is more subdued. Here the slopes are mainly convexo-concave, and the cover of loamy regolith gives rise to deeper and less stony soils.

In order to permit detailed investigations of erosion and sediment delivery from a cultivated part of the basin, an area of 26.5 ha located at the boundary of the forest and the agricultural area had been designated as an "experimental slope" (Fig. 1). The slope is 500-700 m long and convexo-

![Diagram](image-url)

**Fig. 2** The two core sampling transects employed for using radiocaesium measurements to investigate soil erosion and redistribution on the experimental slope. The $^{137}\text{Cs}$ profiles associated with four cores collected from the upslope margins of terraces separating the field plots are illustrated.
concave in form. The silty clay soils increase in depth towards the foot of the slope. The slope is subdivided into numerous field plots (Fig. 2), which are tilled across the slope. The plots are separated by terraces and furrows and by the unmetalled roads which traverse the area from the watershed to the stream channel. During times of heavy rainfall these unmetalled roads, which are typical of cultivated slopes in the Carpathians, act as channels for surface runoff and in many places they are deeply incised into the slope, forming ravines up to 7 m deep. Bedrock is exposed along the floors of many of the unmetalled roads. These sunken roads are being continuously deepened by the concentrated runoff and by their increasingly intensive use, and access to the fields often becomes difficult. In such circumstances, the farmers are forced to construct new roads parallel to the old course. The zone of concentrated water flow and accelerated erosion is thereby gradually enlarged at the expense of the cultivated land. The length of unmetalled road traversing the experimental slope is 3.3 km, equivalent to a density of 11.9 km km\(^{-2}\). The density for the overall basin is 5.3 km km\(^{-2}\).

**METHODS OF INVESTIGATION**

The application of radiocaesium measurements to studies of erosion and sediment delivery within the Homerka catchment began in 1984. The Chernobyl disaster in 1986 caused a substantial increase in \(^{137}\)Cs inventories in the area and introduced difficulties in making comparisons between samples analysed before and after the disaster. Measurements of the caesium-134 activity of soils and sediments can be used to apportion the total \(^{137}\)Cs activity between bomb- and Chernobyl-derived fallout, but some of the results presented in this contribution relate only to samples collected between 1984 and 1986, prior to the Chernobyl disaster, in order to simplify the interpretation. This contribution reports the results of investigations aimed at using radiocaesium measurements to assemble information on rates and patterns of soil erosion from the cultivated slopes, the dominant sources of suspended sediment transported by the Homerka Stream, and the delivery of sediment through the basin system.

The use of \(^{137}\)Cs measurements to evaluate rates and patterns of soil erosion is commonly based on measurements of either the total radiocaesium content or inventory (mBq cm\(^{-2}\)) of the soil profile or the distribution of this inventory within the profile. A 75 mm diameter steel corer was used to collect soil cores to depths of 50 cm for the majority of these measurements, and in most cases the cores were sectioned at 5 cm intervals prior to analysis. Where more detailed information of the vertical distribution of \(^{137}\)Cs within a soil profile was required, samples were collected at 2 cm depth increments using a 40 x 20 cm steel frame and scraper (cf. Campbell *et al.*, 1988; Walling & Bradley, 1990) in order to obtain samples of adequate mass for analysis. All samples were dried, disaggregated and sieved to pass a 2 mm mesh prior to
The use of fallout radionuclides in the Polish Carpathians

analysis by gamma spectrometry in the Department of Geography at the University of Exeter.

Information on suspended sediment sources was assembled using the "fingerprinting" approach described by Peart & Walling (1988). Samples of surface material from potential sources (forest, pasture, cultivated areas, unmetalled roads, gully walls and channel banks) were collected from an area of 1 m², using a steel frame. It is not possible to compare directly the ¹³⁷Cs content of these source materials with that of the suspended sediment transported by the stream, because of contrasts in the grain size composition of source materials and suspended sediment and the known enrichment of the finer fractions in ¹³⁷Cs. The < 0.063 mm fraction of the source materials was therefore separated for gamma spectrometry analysis and the resultant values of ¹³⁷Cs content were used for comparisons with those associated with the suspended sediment. Bulk samples of suspended sediment were collected from the main gauging station on the Homerka stream during flood events. The water samples ranged between 200 and 1000 l in volume, depending on the suspended sediment concentration, and were withdrawn from the stream into 120 and 180 l plastic containers using an electromagnetic pump. The suspended sediment was recovered from the water samples by sedimentation and centrifugation, and the < 0.063 and > 0.063 mm fractions were separated by wet sieving. The separated fractions were dried at 60°C and 100 g sub-samples were used for gamma spectrometry.

Some preliminary information on rates of flood plain accretion, and therefore transmission losses of suspended sediment transported through the stream system, was obtained for an area of flood plain at the outlet of the Homerka catchment by collecting sediment cores from the flood plain and an area of undisturbed pasture above the maximum floodwater level. These cores were collected using a 75 mm diameter steel corer to depths of up to 20 cm. The sediment cores were sectioned into 2 cm increments for subsequent gamma spectrometry analysis. The Homerka drainage basin is a tributary of the Dunajec River which flows into the Roznowski reservoir constructed in 1940. The basin of the Dunajec above this reservoir extends to 4692 km² and is representative of a large Carpathian drainage basin. Sediment cores were collected from the delta area of the reservoir in order to provide information on both rates of sedimentation and the dominant source of the sediment entering the reservoir.

INVESTIGATIONS OF SOIL EROSION

The soil erosion study focused on the area of the experimental slope. Two downslope transects crossing the field plots and their associated terraces were sampled (Fig. 2). Transect A comprises several relatively long field plots up to 100 m in length and with gradients between 12° and 18°, separated by terraces. Transect B is steeper and comprises much shorter
field plots, again separated by terraces. Bulk soil cores (75 mm diameter) were collected from the field plots to a depth of 35 cm. On the upslope edges of the terraces, cores were taken to a depth of 50 cm and sectioned in 5 cm increments to a depth of 25 or 30 cm. Information on baseline $^{137}$Cs fallout inventories necessary to interpret the subsequent pattern of radiocaesium redistribution was obtained from a series of three sectioned soil profiles representing areas of undisturbed grassland with minimal slope on the watershed of the experimental slope and from a single soil profile within a forest clearing in the headwater area of the basin. All samples were collected during 1988 and therefore contain both "bomb" and Chernobyl-derived radiocaesium fallout.

The total $^{137}$Cs inventories of the "input" sites ranged from $5302 \pm 114$ to $7226 \pm 134$ Bq m$^{-2}$ at the top of the experimental slope and a value of $7693 \pm 129$ Bq m$^{-2}$ was obtained for the forest clearing. The caesium-134 contents of the profiles ranged from $1597 \pm 76$ to $1916 \pm 96$ Bq m$^{-2}$, providing a mean $^{134}$Cs input (corrected to May 1986, the period of Chernobyl deposition) of $1782$ Bq m$^{-2}$. All four profiles exhibit an exponential decline in $^{137}$Cs activity with depth. Between 60 and 75% of the total $^{137}$Cs inventory is retained within the top 5 cm of the profile, and more than 90% in the upper 10 cm. The depth distribution of caesium-134 in these "input" profiles demonstrated an even stronger exponential decline with depth. No caesium-134 was detected below 8 cm.

In view of the variability of the estimates of $^{137}$Cs reference inventories for the experimental slope noted above, which largely reflects local variability in the receipt of Chernobyl fallout, it is suggested that $^{137}$Cs measurements are more useful for providing information about the general pattern of erosion and soil redistribution operating over the slope as a whole, rather than the detailed rates and patterns of erosion within individual fields. The total $^{137}$Cs inventories of the individual cores collected immediately above the terraces were in nearly all cases substantially greater than those associated with cores collected from within the plots, reflecting significant soil loss within the fields and deposition at the lower boundary of the fields on the terraces. The mean inventory of bomb-derived $^{137}$Cs associated with cores collected from within the field plots is $4080$ Bq m$^{-2}$, whereas that for the terraces is $8484$ Bq m$^{-2}$. The occurrence of deposition on the terraces is further substantiated by the increased depth to which radiocaesium is found at these locations (cf. Fig. 2). The average depth of cultivation in these fields is of the order of 20 cm and, assuming that the associated mixing would distribute radiocaesium to this depth, the four $^{137}$Cs profiles illustrated in Fig. 2 evidence deposition of c. 20 cm. If it is assumed that this deposition has occurred during a period of about 35 years since the first occurrence of significant bomb fallout, rates of deposition may be estimated at c. 5 mm year$^{-1}$. The radiocaesium measurements indicate that appreciable rates of soil erosion are occurring within the fields, but also that a substantial proportion of the eroded soil is redeposited on the terraces and is not transported beyond the field system.
INVESTIGATIONS OF SUSPENDED SEDIMENT SOURCES

Information on the sources of the suspended sediment transported by a river is of fundamental importance in both geomorphological investigations of erosion rates and landform development and in any attempt to develop measures for reducing sediment loads (Froehlich, 1982; Walling, 1983). Such information is difficult to obtain using traditional monitoring techniques, but the fingerprinting technique has been shown by several workers to offer a viable alternative. In this approach, the physico-chemical properties of the suspended sediment transported by a river are compared with those of the equivalent grain-size fraction of material from potential sediment sources (cf. Oldfield et al., 1979; Peart & Walling, 1986, 1988). Evidence for the Homerka basin assembled using traditional long-term monitoring techniques by Froehlich (1982) suggested that unmetalled roads and active gullies provide the major source of suspended sediment. This conclusion was, however, based primarily on measurements undertaken in the area of the experimental slope and it was necessary to test its applicability to the entire basin, particularly since the intensity of erosion occurring on unmetalled roads was known to vary with the age of the road. Furthermore, there was a need to assess the extent to which conclusions based on the Homerka basin were representative of the Flysh Carpathians more generally, and an attempt has been made to identify the dominant sources of the sediment transported by the Dunajec River. In this context $^{137}$Cs has been used as a fingerprint tracer.

Measurements of the $^{137}$Cs content of the <0.063 mm fraction of suspended sediment collected from the Homerka stream at the main gauging station in the pre-Chernobyl period, indicated a range between 6.3 and 22.6 mBq g$^{-1}$, with a mean of 11.9 mBq g$^{-1}$ and a standard deviation of 4.4. Comparison of these values with typical values for potential source materials (Table 1, Fig. 3) suggests that they closely match those associated with material collected from the surface of unmetalled roads. Sediment eroded from the surface of forest and pasture areas is very unlikely to represent an important sediment source, since its $^{137}$Cs content is substantially higher. Material eroded from cultivated areas and from channel and gully banks could represent a source of suspended sediment, but is thought unlikely to constitute a major source because the range of $^{137}$Cs levels associated with these materials extends well above that representative of suspended sediment. The evidence provided by the radiocaesium fingerprints therefore suggests that the major source of the suspended sediment transported by the Homerka stream is the unmetalled roads which occur throughout both the forested and the agricultural zones of the basin. The use of additional fingerprinting properties could provide more conclusive evidence concerning the relative importance of cultivated areas and channel and gully banks as sediment sources, but is likely to confirm the importance of unmetalled roads. This importance is further underscored by the evidence obtained from a small part of the basin using traditional monitoring techniques referred to above and by analysis of available information on the
Table 1  Caesium-137 concentrations associated with suspended sediment, silt deposited in the stilling basin above a drop structure, and potential suspended sediment sources within the Homerka basin.

<table>
<thead>
<tr>
<th>Samples</th>
<th>Mean concentration (mBq g(^{-1}))</th>
<th>Standard deviation (mBq g(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Suspended sediment &lt; 0.063 mm</td>
<td>11.9</td>
<td>4.4</td>
</tr>
<tr>
<td>Suspended sediment &gt; 0.063 mm</td>
<td>0.8</td>
<td>0.8</td>
</tr>
<tr>
<td>Silt from drop structure</td>
<td>3.7</td>
<td>5.3</td>
</tr>
<tr>
<td>Channel bank material</td>
<td>13.5</td>
<td>15.3</td>
</tr>
<tr>
<td>Gully bank material</td>
<td>16.5</td>
<td>17.9</td>
</tr>
<tr>
<td>Surface material from unmetalled roads</td>
<td>3.8</td>
<td>6.3</td>
</tr>
<tr>
<td>Surface material from cultivated fields</td>
<td>21.7</td>
<td>12.0</td>
</tr>
<tr>
<td>Surface material from pasture</td>
<td>49.0</td>
<td>27.6</td>
</tr>
<tr>
<td>Surface material from forest</td>
<td>57.5</td>
<td>38.0</td>
</tr>
</tbody>
</table>

runoff processes operating in the basin which is discussed below.

Existing evidence relating to the generation of storm runoff within the Homerka basin indicates that the frequency of occurrence of surface runoff on unmetalled roads and in gullies and the furrows between fields is considerably greater than on the cultivated areas. This is further emphasized on Fig. 3 where the typical discharge levels at the main gauging station on the Homerka stream associated with initiation of linear flow on the unmetalled roads and in the gullies and furrows and of overland flow on the cultivated plots and within the areas of pasture and forest are shown. The relationship between the \(^{137}\text{Cs}\)
content of suspended sediment and the discharge of the Homerka stream shown in Fig. 3 shows no evidence of shifts associated with the incidence of overland flow contributions from the cultivated plots and pasture and forest areas of the basin and with the initiation of runoff from furrows between the fields. This in turn again strongly suggests that surface runoff from the unmetalled roads and perhaps also gullies represents the major source of sediment transported by the stream.

This conclusion is further supported by other field observations in the drainage basin which suggest that direct delivery of sediment to the channel from cultivated fields is likely to be restricted, since the pasture areas which exist at the foot of the slopes and bordering the stream channel would act as efficient sinks for sediment transported by overland flow (cf. Froehlich, 1982). Furthermore, the evidence provided by the $^{137}$Cs inventories of cores collected from the experimental slope and discussed above indicate that substantial redeposition of eroded sediment occurred on the downslope terraces bounding the individual plots. Any sediment delivered from the cultivated fields to the channels is likely to move via the furrows, which are in turn connected to the unmetalled roads. Concentrations of suspended sediment observed in the furrows are, however, much smaller than those measured on the unmetalled roads and in the stream channels, again emphasizing the limited connection between the cultivated fields and the river channels. Some sediment from the cultivated fields is, however, transported into the unmetalled roads and gullies by deflation processes during the winter and this will be delivered to the streams during the spring floods (cf. Froehlich, 1982; 1991). Therefore, although the unmetalled roads have been identified as the major source of suspended sediment and much of this sediment will be eroded during the incision of the roads, some of the material will represent sediment particles originating from the cultivated fields and transported to the unmetalled roads by deflation. In the absence of such roads, this material would not reach the stream and their existence may therefore be viewed as being of crucial importance to the sediment budget of the basin. The relative unimportance of surface soils in forest and pasture areas as sediment sources, suggested by the $^{137}$Cs fingerprints, is also further underscored by the low frequency of occurrence of overland flow observed in these areas (cf. Gerlach, 1976; Gil, 1976; Froehlich, 1982).

The > 0.063 mm fraction of suspended sediment is characterized by very low $^{137}$Cs activity, ranging from 0.0 to 2.5 mBq g$^{-1}$, with a mean value of 0.8 m Bq g$^{-1}$ and a standard deviation of 0.82. These low levels reflect both the preferential association of radionuclides with the finer fractions (cf. Tamura, 1964; Frissel & Pennders, 1983) and the dominance of unmetalled roads and gullies and channel banks as sediment sources. Some samples of the finer sediment were recovered from sediment basins located above the drop structures constructed along the Homerka stream (Table 1, Fig. 3). These were characterized by a relatively low $^{137}$Cs content ranging from 0.1 to 10.7 m Bq g$^{-1}$, which conforms with the range associated with both the
>0.063 mm and the <0.063 mm fractions of suspended sediment. This provides confirmation of the representativeness of the $^{137}$Cs fingerprints of the suspended sediment samples.

Sediment samples collected from the surface of alluvial deposits within Roznowski reservoir in the period immediately prior to the Chernobyl disaster were characterized by $^{137}$Cs concentrations in the range 9.1-9.8 mBq g$^{-1}$. These conform closely with the concentrations associated with suspended sediment from the Homerka stream and suggest that unmetalled roads and gullies represent the main sediment source throughout the Dunajec basin. The vertical distribution of $^{137}$Cs concentrations within the sediment core collected from Roznowski reservoir in 1990 is also illustrated in Fig. 4. Some indication of rates of sedimentation at this sampling location can be obtained by linking the shape of the $^{137}$Cs profile to the known temporal pattern of $^{137}$Cs fallout. Measurable $^{137}$Cs concentrations first appear in the profile at a depth of 80-90 cm and these probably coincide with the onset of significant $^{137}$Cs fallout in 1954, although some downward migration or diffusion of the radiocaesium within the sediment column can be expected. The distinct peak at 40-42 cm can be closely related to the peak rates of fallout in the years 1963-1964, whilst the peak that occurs at a depth of 6-8 cm undoubtedly reflects the fallout associated with the Chernobyl disaster in mid 1986. Sedimentation rates at this point are therefore of the order of 2 cm year$^{-1}$, although there is some evidence that they have declined over the period involved. Such a decline is consistent with the

![Fig. 4 The vertical distribution of $^{137}$Cs in a sediment core collected from the delta at the head of Roznowski reservoir.](image)
location of the sampling point on a delta at the head of the reservoir, where the area is inundated during the main flood season but the depths of inundation have decreased through time as sedimentation proceeded. Caesium-137 concentrations associated with the levels in the core that immediately predate the Chernobyl incident are of the order of 5-10 mBq g\(^{-1}\). These \(^{137}\text{Cs}\) concentrations will directly reflect the radiocaesium content of sediment eroded from the upstream basin and deposited at the site, since fallout inputs of \(^{137}\text{Cs}\) in the 1970s were extremely low, and they are again consistent with unmetalled roads and gullies and channel banks providing the main sediment sources within the Dunajec basin.

**SUSPENDED SEDIMENT DELIVERY WITHIN CARPATHIAN DRAINAGE BASINS**

The results presented above indicate that unmetalled roads and, to a more limited extent, gullies and channel banks represent the main suspended sediment sources within the study area. Significant rates of soil loss have been documented within cultivated fields, but little of this eroded material leaves the cultivated areas. Most is deposited on the terraces bordering the field plots and sediment delivery ratios associated with such erosion are likely to be very low. In contrast, sediment delivery from erosion occurring on unmetalled roads and in gullies and river channels is likely to be near 100\%, since the erosion sites are directly linked to the channel network and the linear concentrated flow affords an efficient transport agent. It is, however, important to assess the significance of transmission losses associated with the transport of sediment through the main channel network. Overbank deposition on flood plains is likely to represent the major transmission loss and a preliminary attempt has been made to use caesium-134 measurements to document recent rates of deposition on the flood plain bordering the lower reaches of the Homerka stream and the main tributaries of the Dunajec. Figure 5 illustrates typical caesium-134 profiles associated with a flood plain near the junction of the Homerka stream with the Kamienica Nawajowska River. Figure 5(a) depicts the vertical distribution of caesium-134 associated with a core collected from an area of undisturbed pasture above the maximum flood level, whilst Fig. 5(b) depicts the caesium-134 profile measured in a core collected from the flood plain itself. The presence of caesium-134 reflects only inputs of Chernobyl-derived fallout and in the case of the first core most of the caesium-134 activity occurs, as expected, near the surface. Some downward diffusion or migration has obviously occurred (probably along macropores), but the profile shape closely reflects fallout inputs to the surface. The profile shape encountered in core 2 is, however, very different and almost the reverse of that associated with core 1. The total inventory is also substantially greater. These features of core 2 reflect sediment deposition at the sampling point. At the time of Chernobyl fallout in mid 1986, the level which is now 14 cm below the surface would
Fig. 5 Caesium-137 profiles associated with sediment cores collected from a flood plain site in the lower reaches of the Homerka basin. Profile A was collected from an area above the maximum flood level whereas profile B was collected from an area that is inundated during flood events.

have been exposed at the surface and would have received fallout of caesium-134. Subsequent deposition of caesium-134 bearing sediment eroded from the upstream drainage basin has buried the surface exposed in 1986 and the deposited sediment has increased the total caesium-134 inventory. Since these two cores were collected in 1989, annual deposition rates of c. 4 cm year\(^{-1}\) may be estimated. Similar rates of deposition have been estimated for other flood plain locations within the Dunajec river system, indicating that flood plain accretion is widespread and that even in these high energy mountain environments, significant transmission losses may occur in association with flood plain deposition. Further work is, however, required to quantify the magnitude of the transmission losses involved.

PERSPECTIVE

The results presented above indicate that radiocaesium measurements undertaken in the Homerka basin provide evidence relating to soil erosion on cultivated fields, the dominant suspended sediment sources, and sediment delivery which is consistent with the results of long-term labour-intensive classical monitoring techniques. The radiocaesium measurements were, however, undertaken during a short period and provide greater scope for investigating processes at the overall basin-scale. Thus, for example, the fingerprint technique can be used to assemble information on the relative importance of the potential sediment sources within the overall basin, whereas classical monitoring techniques can only document sources within small areas. Furthermore, the use of radiocaesium measurements to interpret sediment collected from the Roznowski reservoir and to estimate contemporary rates of flood plain sedimentation provides a means of extending the scale of investigation from the small Homerka basin to the much larger Dunajec basin. Radiocaesium measurements also afford valuable potential for extending the temporal base of process monitoring since the use of bomb fallout provides a
means of assessing sediment movement over the past 35 years. Further work, exploiting the potential afforded by both "bomb" and Chernobyl fallout, is planned in order to develop an improved understanding of erosion and sediment delivery within the Polish Flysch Carpathians.

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A new look at soil erosion processes on hillslopes in highland Ecuador

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Abstract Field research with portable rainfall simulators in the 5186 km² Paute River basin in highland Ecuador indicates that footpaths generate runoff more rapidly and more often than adjacent fields and that pasture and abandoned crop lands are frequently important runoff source areas. Differential runoff production allows run-on water to play an important, but previously unexamined role in soil erosion. This research demonstrates that the study of soil erosion on mountain hillsides needs to treat the runoff dynamics of the entire hillside, rather than those of only the cultivated plots.

INTRODUCTION

The assessment, modelling and reduction of soil erosion by rainfall are commonly based on the concept that surface runoff occurs when certain threshold conditions are exceeded. Delineating runoff thresholds for different soil, rainfall, slope, and surface treatment conditions thus helps in predicting and avoiding rainfall erosion losses. In the Ecuadorian Andes, I have been surprised to find evidence of soil erosion in places where it would not have been predicted by conventional methods. One example of such unexpected soil erosion is the presence of rills in cultivated fields following relatively insignificant rainfall events.

The heterogeneity of hillslope surface characteristics and the ensuing variability of surface runoff would appear to be key to the understanding of upland soil erosion. In this paper, the results of field experiments with portable rainfall simulators are presented and discussed in order to address the question: How do the runoff and soil detachment characteristics of different hillslope surfaces affect soil erosion dynamics in the highlands of southern Ecuador?

THE STUDY AREA

This study was conducted in the 5186 km² Paute River drainage basin in the Andes Mountains in southern Ecuador (Fig. 1). Elevations range from 4300 a.m.s.l. at the head of the drainage basin, to 1800 a.m.s.l. at the base of the hydroelectric power dam that defines the lower boundary of the study area.
Because of the high elevation and the double rainshadow effect provided by the two cordilleras of the Andes mountains, the climate is driest in the central portion of the drainage basin and is relatively cool for its location at approximately 2 to 3°S. Annual precipitation ranges from 800 mm in the drier central region to 1400 mm in the high paramo and to a maximum of 3500 mm in the vicinity of the dam, where Amazonian air penetrates the eastern Andean cordillera via the Paute River valley (Morris, 1985). The daily mean temperature of 14.6°C at the airport in Cuenca (2527 a.m.s.l.) varies little throughout the year. At the downstream boundary of the drainage basin, the Paute River is a sixth order stream with a mean discharge of 122 m$^3$ s$^{-1}$. Soils in this area are derived primarily from Tertiary marine sediments (Figueroa, 1987).

![Fig. 1 Location of the River Paute study area.](image)

Approximately 500,000 persons live in the study area, 200,000 of whom live in the provincial capital of Cuenca (UMACPA, 1989). Inhabited mountain environments in Ecuador are characterized by a mosaic of small cultivated plots and pasture lands. In the Paute River basin, the average farm size is 1.1 ha and 91.6% of the farms have < 5 ha (Figueroa, 1987).

The landscape of the Paute River basin differs substantially from that of the midwest United States where large numbers of runoff plots have supported the development of empirical soil erosion modelling. In the Paute, farm units are small and frequently on steep (>50%) slopes. The primary crop is corn, often grown in association with beans. Peas are typically rotated with corn so that fields are in use for 12 months. Nearly all agricultural labour is by hand and cattle, sheep and hogs are staked in the fields to graze after the corn is harvested. An extensive network of footpaths links homes, villages, fields,
pasture lands, and primary roads. About one-fourth of the Paute drainage basin is in forest or scrub vegetation.

METHODS

Rainfall experiments were conducted using two portable rainfall simulators based on the design of McQueen (1963). These instruments generate a replicable "rain shower" 15.2 cm in diameter into a ring of the same diameter inserted into the soil. A fixed 5.6 mm drop size is calculated to deliver the kinetic energy representative of natural rain having an intensity of 25 mm h⁻¹, and the experimental rainfall intensity is controlled by maintaining a constant hydraulic head in the rainfall simulator. Thus, experimental rain events may be replicated at the same or different sites. Because the apparatus uses only several litres of water per trial, it may be easily transported and set up at many different locations.

The experiments described in this paper involved 30-min simulated rains with a median intensity of 26 mm h⁻¹. This $I_{30}$ has a frequency of 1.2 year⁻¹ at Cuenca (Ecuador, 1989). Antecedent soil moisture was standardized in 107 trials, including all trials reported in this paper, by pre-wetting the soil. During each trial, simulated rain that ponded at the surface within the ring was drawn off at 5 minute intervals to measure the volume of "runoff". It was filtered at the site and later dried to determine the oven dry mass of detached sediment. Splashed sediment was washed from the ring and weighed with the entrained sediment. Although the simulators may be used on sloping sites, the small size of the base and the microrelief of the land surface do not allow ponded water to obtain any noticeable downslope velocity within the basal ring.

Experimental sites were chosen to represent the range of soil and land use conditions characteristic of the drainage basin. Because little runoff or soil erosion have been previously shown to occur in the high altitude paramos (Harden, 1988), emphasis was placed on rural lands below the paramo. Land uses investigated included road and trail surfaces; actively farmed crop lands, with standing crops or stalks; abandoned crop lands; recently ploughed crop lands, characterized by large clods; grasslands, with the sod removed for the experiment; and tree cover, with either native Polylepis forest or plantations of Eucalyptus or Pinus. It is common for all of these land uses to occur on the same hillside. All of the experiments reported here involved simulated rain falling onto pre-wetted, unvegetated soil surfaces.

RESULTS

The values of runoff and soil detachment measured in the Paute watershed varied spatially and temporally. Spatially, analysis of variance yields no significant relationship ($p > F = 0.65$) between mapped soil type and runoff
volume. Closer examination of runoff volumes and soils within land use groups shows that runoff variations relate to soil type only in recently ploughed fields (significant at the 0.08 level). Runoff volume appears to be more closely related to land use, but its overall relationship to land use is significant at only the 0.19 level. Analysis of variance shows that, within mapped soil units, the relationship between runoff volume and land use is significant at the 0.10 level for three out of nine and at the 0.15 level for five out of nine soil types. Because surface runoff is the primary control of soil erosion in this study and land use appears to be better than soil as a predictor of runoff, results are presented by land use rather than by soil categories.

Variation in runoff volume within land use categories was nonetheless high; in some instances, it was high between adjacent trials in the same field. Table 1 shows the presence and amount of runoff and the quantity of soil detached on bare surfaces in 30-min rainfall simulation experiments. The results shown in Table 1 were obtained from trials involving experimental intensities ranging from the median of 26 mm h⁻¹ to the lower and upper quartiles of 19.8 and 32.6 mm h⁻¹, respectively. Because most road and trail surfaces were tested at intensities below, and most tree covered surfaces at intensities above, this range, those three within the range are included in the table for comparison only. The difference between runoff reported in categories of > 0 or > 1 ml represents the occasional presence of a trace of runoff too small (less than 1 ml) to be measured.

Seven road or trail surfaces were tested in the Paute drainage basin, using rainfall intensities ranging from 5.6 to 11.4 mm h⁻¹. In all cases, more than 1 ml of runoff was produced during the 30-min experiment; in almost all cases, runoff began within the first 5 min. Of the two tree covered sites tested in the median intensity range, one yielded runoff and sediment and one did not. This dichotomy was further explored using higher simulated rainfall intensities. Nine experiments were conducted in forests and tree plantations. In the five trials in native forests and high elevation Pinus plantations with Sphagnum ground cover, no runoff was generated, even at extremely high intensities (I₃₀

<table>
<thead>
<tr>
<th>Use</th>
<th>N</th>
<th>% Trials with runoff</th>
<th>Runoff: mean ± std dev.</th>
<th>Sediment detached: mean ± std dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>0</td>
<td>&gt; 1 ml</td>
<td>ml</td>
</tr>
<tr>
<td>Road/trails</td>
<td>1</td>
<td>100</td>
<td>100</td>
<td>190</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2.6</td>
</tr>
<tr>
<td>Cropland:</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Abandoned</td>
<td>13</td>
<td>100</td>
<td>85</td>
<td>84 ± 81</td>
</tr>
<tr>
<td>Active</td>
<td>20</td>
<td>85</td>
<td>65</td>
<td>50 ± 76</td>
</tr>
<tr>
<td>Recently ploughed</td>
<td>8</td>
<td>75</td>
<td>62</td>
<td>74 ± 95</td>
</tr>
<tr>
<td>Grassland (sod lifted)</td>
<td>6</td>
<td>100</td>
<td>83</td>
<td>106 ± 117</td>
</tr>
<tr>
<td>Tree cover</td>
<td>2</td>
<td>50</td>
<td>50</td>
<td>2 ± 3</td>
</tr>
</tbody>
</table>

*Because all trials on roads and trails yielded runoff at lower intensities, only one was tested in this intensity range.
up to 200 mm h\(^{-1}\)). Runoff occurred and sediment was detached in the remaining four trials in *Eucalyptus* plantations. *Eucalyptus* is commonly planted on degraded soils in highland Ecuador, but it is known to contribute little to the improvement of soil infiltration characteristics (Morris, 1985).

The mean time required to initiate runoff on a pre-wetted surface was calculated for each use category by analysing field data taken at 5-min intervals. Time to runoff, shown in Fig. 2, is generally less for roads, trails and many abandoned surfaces, and particularly high (higher than all experimental intensities) in sites with native forest. Pastures and abandoned lands produce runoff more rapidly than most cultivated fields.

![Fig. 2 Mean time thresholds for runoff on various pre-wetted surfaces.](image)

**DISCUSSION**

Variations in runoff and soil detachment between land use categories demonstrate the importance of land use in determining spatial patterns of soil erosion. These results from Ecuador agree with the work of Dunne in Kenya (Dunne, 1977, 1979) which showed land use to be the dominant control of sediment yield.

The assumption that the erosional behaviour of soil is inherently related to its taxonomic classification is attractive when extrapolating soil erosion rates or determining spatial patterns of soil erosion. This assumption allows models such as the USLE to be used over large areas and often forms the basis for the designation of erosion risk zones in Geographic Information Systems (see, for example, Giordano, 1984). The poor relationship in the Paute River basin between soil and runoff, however, does not support this assumption. Although the distributions of runoff and sediment detachment characteristics had been closely related to the distribution of mapped soil units in a previous study area near Ambato, Ecuador, where soils are derived from volcanic deposits (Harden, 1990), that relationship does not hold for the clay-rich, sediment-derived soils of the Paute River basin.

Spatial variations in runoff and soil detachment at the microscale (within a few square metres) are site-specific and may be due to numerous factors including the presence/absence of macropores, microtopography, compaction by trampling, spatially heterogeneous surface treatments (e.g. furrow
construction or mounding at the base of corn stalks), and spatially heterogeneous soil characteristics. Hills (1971) attributed more of the spatial heterogeneity of infiltration to biotic interference, in the form of compaction, than to natural variations among soils. In the inhabited mountain landscape of the Paute River basin, such biotic interference is widespread. Abandoned fields in the Paute, so called because they are no longer being actively farmed, exhibit a wide variety of surface characteristics, and, are thus highly variable in their runoff and erosion behaviour. In practice, these lands are not truly abandoned - cattle, sheep, and hogs graze in them, compacting the soil and limiting the growth of a protective vegetative cover. Vegetation is also slow to re-establish where crop lands were abandoned for reasons of soil degradation. Because grazing is a year-round activity and because the tropical environment of highland Ecuador lacks a freeze-thaw season that would reduce compaction, trampling impacts are cumulative. The high degree of microsite variability evident in this study and the importance of soil compaction to runoff production suggest that the use of generalized rather than site-specific soil parameters in soil erosion studies may be misleading, thus supporting the view of Govers (1989) that soil erodibility is dynamic and strongly influenced by antecedent conditions.

The existence of partial and variable runoff source areas is well-established in the literature of hydrology. Betson (1964) reported storm runoff to occur in only a small part of the watershed area in pasture watersheds in North Carolina; and subsequent papers by Betson & Marius (1969), Dunne & Black (1970), and Hewlett & Troendle (1975) further developed the partial- and variable source area concepts. It is important that this concept be extended to

![Fig. 3 Rainfall frequency for Cuenca, Ecuador 1972-1984, with runoff thresholds for trails and agricultural lands.](image-url)
Soil erosion processes on hillslopes in highland Ecuador

the heterogeneous landscapes of inhabited mountains. Simulated rainfall experiments in the River Paute drainage basin demonstrate that roads and trails play a key role as partial area runoff contributing networks, and that abandoned and pasture lands generate runoff more frequently than actively cultivated and recently ploughed crop lands or forest.

While the spatial heterogeneity of inhabited mountain environments helps to account for the variety of sources of surface runoff, it also provides runoff sinks. These may be high infiltration capacity surfaces, such as recently ploughed fields, or topographic barriers that block runoff and sediment transport and prevent downslope soil losses. The concept of sediment and runoff sinks forms the basis for many conservation practices. Cultivation of the soil may legitimately be the primary cause of soil loss in areas where uncultivated lands have good vegetative cover and high infiltration capacities, but the present study indicates that ploughing the clayey soil of Paute hillslopes improves infiltration and, relative to abandoned and compacted surfaces, retards runoff and sediment transport.

Differences in the time between onset of simulated rainfall and the initiation of surface runoff within a short duration rain event demonstrate an important temporal aspect of partial area runoff. The rapid onset of runoff on roads and trails makes them a primary target for soil loss and (along with muddy feet) helps to explain the greater cumulative losses found from entrenched trails on mountain hillsides compared to adjacent cultivated lands. Although wholesale soil loss in the Andean region is probably rightly attributed post-colonial activities (de Noni et al., 1986), extensive pre-Colombian trail systems may also have played an important runoff contributing role. Temporal differences in rainfall runoff become more noticeable when the soil is not already wet at the onset of rain, since the time needed to generate surface flow is far less on compacted than on uncompacted dry surfaces (Harden, unpublished data).

Differences in time to runoff between land uses appear during single rain events, when trail surfaces begin to initiate runoff prior to crop lands (Fig. 2). These differences in runoff initiation are amplified over the course of a year, since the frequency of rain events exceeding the runoff thresholds of road and trail surfaces is greater than of those yielding runoff on ploughed fields (Fig. 3). Some forest surfaces may never yield surface runoff.

The existence of partial areas of runoff production and sediment transport leads to the presence of run-on water on surfaces that would otherwise not experience runoff. It thus has extremely important implications for the study of soil losses at the hillslope or watershed scale. Particularly in inhabited mountain watersheds, where plots tend to be small and trails tend to be frequent and steep, the study of erosion needs to treat the entire hillside rather than be based on homogenous cultivated plots. "Anomalous" rill erosion that I had observed in highland Ecuador can be explained by looking at site-specific factors, especially the presence of upslope runoff sources. The presence of a trail, pasture, abandoned field, or other low infiltration surface upslope from a cultivated field can (a) cause surface flow to run onto the cultivated field
during a rain event that would otherwise not initiate surface flow there and (b) cause surface flow to be present in the field sooner during a rain event, setting the stage for increased erosion. Such interaction of landscape units occurs where barriers are not present and surface flow is able to pass between units.

Dunne & Dietrich (1982) estimated the sediment contributions from roads in densely settled subsistence agricultural areas in both wet and drier regions of Kenya to be 25 to 50% of basin yields. As in Kenyan regions, roads and trails are estimated to occupy about 1% of the area of the Paute River basin in Ecuador. Preliminary observations in the Ecuadorian drainage basin indicate that, although interaction with less efficient transport surfaces leads to generally poor delivery of runoff and sediment from footpaths to stream channels, the interaction itself is an aspect of upland erosion that merits further study. Dunne & Dietrich (1982) did not link road erosion with a lowering of the productivity in agricultural land in Kenya, but the present work in Ecuador suggests that run-on water from road and trail runoff, as well as that from pastures and abandoned lands, may be responsible for a significant portion of annual soil losses from crop lands.

This study demonstrates the need for soil erosion research and models to extend beyond crop lands and to include potential upslope contributing areas. Since the distribution of overland flow-producing areas can be expected to influence the composition of compound overland flow downslope (Hills, 1971), identifying surface runoff patterns on entire hillslopes will contribute to hydrogeochemical modelling. A further implication of this study is that conservation efforts should look beyond the small farmers, whose activities may cause their lands to play greater roles in the upland erosion system as runoff sinks than as sediment sources.

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