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Erosion and sediment dynamics from catchment to coast

A NORTHERN PERSPECTIVE

CONTENTS

1. BASIC FORMS OF SEDIMENT MOTION

Removal of sediment from the watershed slopes (erosion) and the subsequent discontinuous motion (dynamics) to the ocean, involve a variety of processes that may be analyzed and classified under different view points, as described in the following.

A middle size watershed of temperate zones (but in fact applying to other climates as well) is schematically depicted in Fig. 1. To give an idea of what usually takes place at different elevations and distances along the watercourse, the longitudinal dimension is approximately indicated by a logarithmic scale, in such a way as to emphasize the complexity of the problems occurring at the smaller scales (farther and higher areas of the watershed).

Under the action of water (direct: rainfall, overland flow, channeled flow; and indirect: freezing and melting, infiltration, etc.), sediments are removed from the surface of the watershed and conveyed downstream. Depending upon the prevalent extension of the process in three, two or one spatial dimensions, sediment motion assumes three basic forms (mass, surface and linear), more or less corresponding, respectively, to (i) landslides, occasionally produced in the steepest slopes of the watershed, even if protected by vegetation; (ii) distributed soil erosion mainly occurring in undulated, scantily vegetated surfaces; and (iii) bedload and suspended transport by waterflow in the stream network. There are also a number of intermediate forms which share some characteristics with the basic ones, as for example: gully development (mass/surface/linear motion) rills erosion (surface/linear movement), debris flow (mass/linear motion). Where rainfall is extremely scarce, as in the desert or in arid zones, wind is often the most effective cause of surface erosion.

Physical phenomena related to sediment motion are therefore extremely numerous and strictly connected with the morphoclimatic conditions under consideration. Moreover, they are traditionally dealt with by different disciplines and professions, very often under a quite “parochial” perspective.

Mass movement, characterized by quick and short displacements of large portions of soil, represent sometimes a risk for human settlements and infrastructures, but also a physiological source of sediments to the rivers in several natural watersheds (e.g. in alpine and humid tropical regions). Investigation on mass movement is generally carried on by applied geologists and, for the structural aspects, by soil mechanics engineers. Mass movement specialists are often barely interested in the final destination of the removed material as sediment yield.

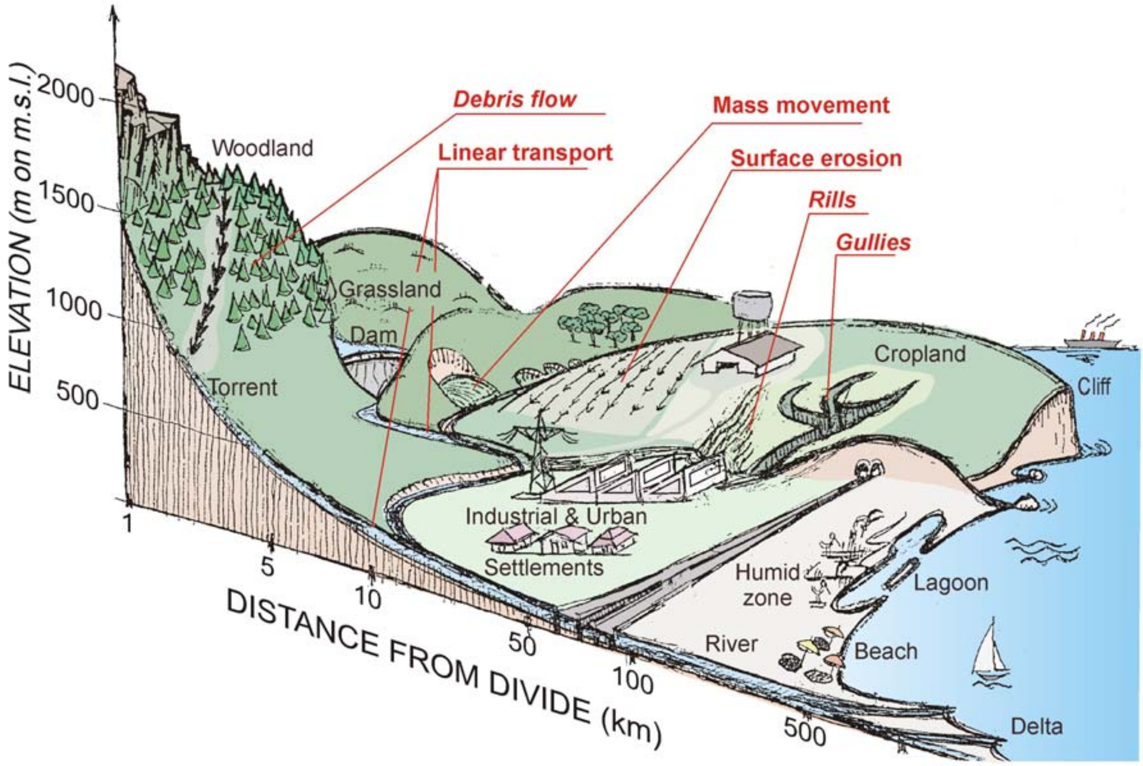

Fig. 1 Sketch of a watershed in temperate zones: Basic forms of sediment motion

Surface erosion, usually as *sheet erosion* but including also intermediate forms like *rill* and *gully erosion*, because of its strict implications with land use and agricultural practices, usually belongs to the province of agronomists and agricultural engineers. It is also investigated however by various scholars of earth science. These forms of erosion constitute a natural source of sediments both in arid tropical and temperate regions where rainfall is generally the dominant mechanism of sediment production. On the other hand, surface erosion tends sometimes to be overestimated as a component of sediment yield even in the cases where mass movement prevails.

Finally, *linear transport* is traditionally in the competences of hydrologists and rivers engineers. Bedload and suspended sediment transport convey coarse and fine particles over extremely long distances along the river, down to the estuary, the sea and the adjacent beaches, where they usually pass under the "jurisdiction" of coastal engineers and oceanographers. While solid transport in the river includes material produced by the entire watershed, little attention is generally paid by fluvial and maritime specialists to the sediments' sources.

A specific disciplinary approach is almost invariably assumed, successfully, to solve most of the engineering problems. However, to understand the behaviour of sedimentary systems at relatively large space– and time scales (see Sect. 5), knowledge and experiences from different branches of science and professions should be brought together. This operation is not at all easy, not only between academic disciplines but also between separate ministries and agencies which in each country have competence on sediments.

A short review of the three basic forms of sediment motion mentioned above is given in the following Sections 2, 3 and 4.

2. MASS MOVEMENT

Mass movement corresponds to the detachment of sediments as a bulk from their original position (landslides), when the resisting forces (friction and cohesion) become lesser than the acting force (gravity). Mass movement is an important source of material for many rivers and in some cases the most important one. In *humid tropical forests* as well as in *alpine climates*, for example, the natural thick vegetation cover is such that the direct effect of rainfall (kinetic energy) on the soil is negligible and the sediment production by surface erosion is practically zero. Yet the sediment transport by mountain rivers may be substantial and even extremely large (up to 10^4 t/km^2/year), due to the contribution of repeated *slope collapses* and occasional big *landslides*. Small and large mass movements from the watershed slopes typically occur during large floods and intense storms and are often associated with *mud*- and *debris flows* in the upper branches of the hydrographic network.

Mud- and debris flows (including ash flow or "lahars", taking place along the steepest channels of volcanoes) are intermediate forms of sediment motion, between mass movement and linear transport, which require a relatively small minimum steepness to be initiated. While their motion depends on particle- and fluid dynamics (similarly to linear transport), their triggering is controlled by static forces, basically depending upon friction, cohesion, slope and the degree of saturation of permeable material (as for mass movement). For this reason attempts have been made to model the triggering of both shallow landslides and debris flows by simulating the saturation process of the surface layers of watershed slopes and steep channels (see for instance Dietrich, W.E. and Montgomery D. R., SHALSTAB - a digital terrain model for mapping shallow landslide potential -. to be published as a technical report by NCASI (Nat. Counc. For Air and Stream Improvement). Also available on Internet Baum et. al, 2o2002;

Collapse of river banks is also controlled, like landslides, by friction, cohesion, slope and saturation, but its triggering is often determined by foot erosion produced by water flow. For this reason bank collapse material is considered part of linear transport (see Sect. 4) and its simulation usually included in morphodynamic modelling of rivers (see Sect. 6).

3. SURFACE EROSION

Surface erosion, prevalently developing over two dimensions, is definitely the most important source of sediment production wherever vegetation does not provide a sufficient cover of the soil from the rainfall impact, and morphological conditions are such as to foster the removal of particles by overland flow. This means that surface erosion is particularly active in cropland areas, especially where the type of soil is more vulnerable, yet erosion-control measures and correct cultivation practices have not been applied. In many temperate countries, extremely high rate of surface erosion took place in historical times, following the rapid expansion of cultivated areas and before sustainable land management was adopted. The most recent episodes of this type occurred about hundred years ago in the U.S.A., where extensive areas of the Midwest were rapidly transformed from natural grassland into cropland. For this reason soil erosion was first investigated at scientific and technical levels in this country, with special reference to the undulated landscape and climatic conditions typical for these areas.

The most active institution in this field was certainly the U.S. Department of Agriculture, where the renowned U.S.L.E. model has been proposed. The U.S.L.E. (Universal Soil Loss

Equation) was developed since several decades (Wishmeyer and Smith) by using the U.S.D.A. data base containing a very large number of results. The multiplication structure of the formula tries to put into account all the following factors: kinetic energy impact of rainfall combined with the intensity of rainfall, this last proportional to overland flow discharge (*erosivity factor*, *R*); resistance of the soil, quantified by means of descriptive tables (*erodibility factor*, *K*); slope length, also proportional to overland flow discharge (*length factor, L*); slope steepness, related to overland flow velocity (*steepness factor*, *S*); protection by vegetation depending on plants, crop and vegetative phase (*cover factor, C*); and management practices (*practice factor*, *P*).

The U.S.L.E. has been thoroughly criticized and defended in literature, but also extensively applied even outside the U.S.A., although very often with various "adaptations". The formula provides, in principle, the values of sediment production at the "plot- or field scale" for a given period of time. For obtaining the corresponding data at catchment scale, the sediment production should be "routed" downhill to the hydrographic network and, eventually, downstream along the river to the closure section of the basin. The routing process that transforms the local *sediment production* into the integral *sediment yield* of the entire watershed is a rather delicate matter (see Sect. 8).

Besides the U.S.L.E. equation, more sophisticated models as ANSWERS, WEPP, SHE-SED, EUROSEM etc. have been recently developed for simulating, at catchment level, the detachment of soil particles by rainfall and their subsequent transport by overland flow and by river flow over the entire catchment (Beasley at al. 1980, Nearing et al. 1984, Wicks and Bathurst, 1996, Morgan et al, 1996 etc.). In contrast with the so-called "empirical" models (like USLE), the last models are usually called "physically based", since they are constituted by theoretical differential equations (expressing the mass balance of water and sediments) and by appropriate algebraic equations (describing each of the physical processes involved).

Physically based models resemble somehow the erosion- transport- deposition models employed in river morphodynamics (see Sect. 6). The physical processes involved in both *water flow* and *sediment motion*, however, are much more complicated on the watershed slopes than in rivers, and therefore much more difficult to be realistically simulated (see Sect. 6 and 8). For this reason, empirical models controlled by few overall coefficients (scarcely recognizable from the physical point of view but quite consistant and confirmed by many and many experiments) frequently give much better results than physically based models controlled by a large number of coefficients (generally unknown and based on hardly plausible physical and geometrical schematizations) which ignore in any case relevant existing interactions.

4. LINEAR TRANSPORT

Linear transport, namely taking place along one prevailing (longitudinal) direction, is the motion of sediments produced by persistent, channelized water flow. It is mainly responsible for river processes in the hydrographic network.

4.1 Modes and rate of transport

Linear transport assumes various modes (*bedload*, *suspension* and *intermediate forms*), but attempts have been made towards a conceptual unification of these forms, through the notion of *adaptation length*. The adaptation length expresses the distance required by clear water entering a uniform flow stream flowing over a uniform grainsize bottom to reach the uniform sediment transport conditions. The adaptation length depends on the particle grain size and on the characteristics of the water flow, i.e. more precisely on the ratio between friction velocity u^* and particle settling velocity w_s. When the ratio (u^*/w_s) is very small, the adaptation length has the order of magnitude of 10^2 *grain diameters* and the particles move by sliding and rolling as bedload. When this ratio increases, also the adaptation length correspondingly increases and the motion passes from saltation to suspension. Adaptation length is practically zero for coarse material moving as bedload, while for fine particles moving in suspension it may reach the value of tens of kilometers.

The *solid discharge* of a natural stream (expressed by the mass or volume of sediments conveyed per unit time through a given cross section) may be somehow evaluated by the so-called *sediment transport formulas*. As it will be seen better in Section 7, most of the formulas have been obtained in laboratory under *uniform conditions (uniform transport by uniform plane flow* and *uniform grainsize* material). In these conditions, the *total solid discharge* can be expressed as a function of the water flow characteristics and the particle diameter, but the total amount may be somehow splitted between bedload and suspended transport. In fact, the distance covered by the particles under the action of the water flow does have a statistical distribution, depending on the ratio (u^*/w_s). This ratio therefore defines the ratio between the number of particles instantaneously subject to different modes of transport, as well as their adaptation length.

When the adaptation length is quite long, the sediment transport rate does not depend solely on the local hydrodynamic and sedimentological characteristics, but also on the conditions upstream. This circumstance in part explains why the suspended transport in a given cross section of a river is often scarcely correlated with the local water flow .
The adaptation length can be evaluated by different approaches (Galappatti, 1985, Armanini and Di Silvio, 1988, Bolla Pittaluga and Seminara, 2003) and its effect should be taken into account, when necessary, in sediment transport computations (see Sects. 6 and 7).

4.2 Sorted material

In real rivers, particle grainsizes are more or less non-uniformly distributed, with markedly different statistical distributions for *bed material* and *transported material*. In general, bed material appears to be coarser than transported material, and the two distributions can be mutually related by considering the transport of each grainsize class (see Sect. 7).

When treating different grain size classes, due attention should be paid to the interference of particles of different diameter. In sediment mixtures, in fact, the intrinsic larger mobility of finer particles is somewhat diminished by the presence of the coarser ones (“hiding” effect) while the intrinsic smaller mobility of coarser particles is augmented by their protrusion (“exposure” effect). With very strong water flow in flood periods, the hiding-and-exposure effect may even lead to an “almost equal mobility”. In low flow periods, by contrast, the different intrinsic mobility of various diameters strongly prevails over the hiding-and-exposure effect (indeed, the coarser particles may even not move at all). In any case, over a long period of time, the transported material (e.g. the material intercepted by a reservoir) appears to be definitely finer than average composition of the river bed.

The “hiding-and-exposure” effect may be taken into account by various empirical coefficients to be introduced in the formulas developed for uniform material. The time evolution of bed- and transport composition is usually modelled by resorting to the *active layer* concept, first proposed by Hirano and subsequently incorporated in many morphodynamic models. More sophisticated approaches have been developed more recently, either by disaggregating the bottom *active layer* into a *mixing*- and an *intrusion layer* (Di Silvio, 1991), or by considering the bottom a continuous, indefinitely deep layer, statistically described in terms of entrainment capacity (Armanini, 1995, Parker et al, 2000)

4.3 Cohesive material

In some circumstances (e.g. estuaries, flood plains, deep reservoirs) sediments can hardly be considered as non cohesive. The role of cohesion is quite important both in the deposition phase (flocculation) and in the re-entrainment process (compaction). The pioneering work of Partheniades and Krones in the sixties of last century and, in the seventies, of Methas and Partheniades is still the foundation of many models for cohesive materials. Some of the models developed from their basic concepts, however, do not appear completely satisfactory and are unable to explain a number of phenomena observed in nature. It is therefore of much interest the attempt by Winterverp (2001) to bring together the behaviour of cohesive and non-cohesive material within a unified physical framework with specific definitions of vertical fluxes for each type of sediment.

5. TIME- AND SPACE SCALES OF SEDIMENTARY SYSTEMS

Morphological processes may be seen as the product of repeated succession of three phases of sediment motion : erosion, transport and deposition. In some cases, one of the three phases is definitely dominant. For example, soil removed from short watershed slopes, either by surface erosion or mass movement, may be never replaced by other soil. Conversely, sediment trapped by a deep lake or sea are not entrained and put in motion anymore. In these cases the erosion or deposition process is time-depending but monotone (namely producing either a progressive degradation or a progressive aggradation). In many other cases, by contrast, subsequent phases of erosion, transport and deposition take sequentially place on the same location, giving origin to complicated alternating morphological processes. In this last cases one can only speak of *net* degradation or aggradation of a certain *sedimentary system* over a prescribed *period of time*.

When considering morphological processes, it is important to have in mind the time- and space scales under consideration. The repeated succession of erosion, transport and deposition, may concern for example: (i) the sliding, rolling and saltation of sediment particles over bed ripples (space scale: boundary layer, say millimeters); (ii) the propagation of dunes (space-scale: river depth, say meters); (iii) the formation of bars and meanders (space scale: river width, say hundreds of meters); (iv) the general aggradation or degradation of a river (space scale: watershed, say up to thousand kilometers). The time-scale of each system may be associated to the corresponding space-scale, *via* a typical process velocity.

It is important to note, in any case, that each system at a given scale may be considered a component (or sub-system) of the system at the larger scale. The morphodynamics of the component does in principle intereact with the morphodynamics of the system at larger scale. However, to describe the behaviour of a component (e.g. the propagation of dunes along a river reach) it is usually assumed that the system at larger scale (e.g. bars and meanders) remains *stationary* at the time-scale of interest for the component (dunes). At this time-scale, conversely, one assumes that the subsystem at an even smaller scale (e.g. bed ripples), although non-stationary, is in *equilibrium conditions* with the larger system (dunes). This simply means that, during the propagation of the dune, single ripples may appear or disappear, but their statistical distribution (and consequent hydraulic roughness of the dune surface) depends exclusively on the dune configuration. This assumption is only valid, in principle, when the relevant systems and sub-systems have markedly different scales, yet it is implicitly assumed in most morphological models (see Sect. 6).

The scales of morphological processes extend over several orders of magnitudes ranging from microns to continental sizes (in space) and from seconds to millions of years (in time). The graph of Fig. 2 indicates the range of interest for various disciplines interested in sedimentary systems. For *Hydraulic Structures* (*construction prototype)* engineers are generally interested in problems defined (in space) by the "size of the structure" and (in time) by the "event duration", or, at most, by the "project life" of the structure. For *Hydraulic Laboratories (laboratory experiments)* the range of interest is defined by the facility's size and the process' velocity. In basic research (e.g. for analyzing the behaviour of individual sediment particles) the relevant sizes may be extremely small, while for physical models they are generally larger, although obviously much smaller than the size of the corresponding prototype structure (we may say, in Froude similitude, 100 times less in space and 10 times less in time). However, if the engineer wants to assess the morphological effects of the structure he has designed on the entire river system, he should take into account much larger scales. For example, the presence of the Aswan Dam is already perceived, after several decades since its construction, in the Nile's Delta (subject to erosion) which is thousands of kilometers downstream. Yet, the adaptation process of the entire river Nile system will take an extremely longer time to attain a new quasi-equilibrium configuration. In other words, as shown in Fig. 2, the time- and space-scales for *Environmental Engineering (protection)* tend to be much larger than for hydraulic structures and be closer to those of the *Geological Sciences*, namely "geological times" and "continental sizes".

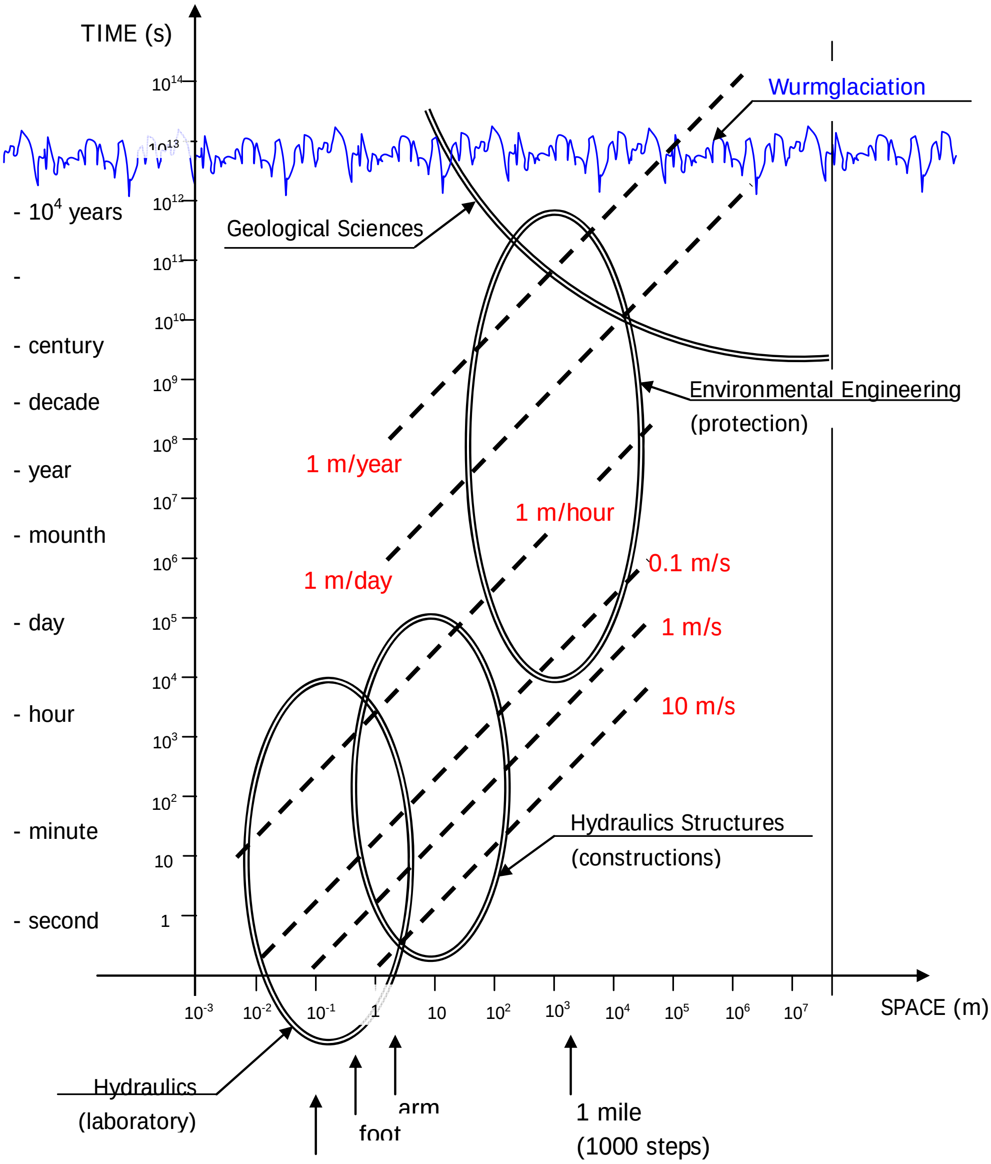

Fig. 2 Time- and space scale of sedimentary systems

As a pure indication, the time axis of the figure is bounded by the end of Würm glaciation, as it is called in Europe the last large climatic change before present, which has interested, with some local variation, both emispheres and, in terms of sea level change, the entire planet. We may assume, in fact, that at geological scale the climatic forcing after the Würm glaciation was reasonably stationary. By contrast, should we consider a longer period of time, several processes would appear to be controlled by non-stationary climatic conditions (sequence of glaciations and consequent sea-level changes), as well as by variable phases of tectonic uplift and subsidence.

However, even by limiting the analysis to the last ten thousand years, a quite large number of time- and space scales controlling the behaviour of sedimentary systems should be considered when developing morphological models (see Sect. 6).

6. MORPHOLOGICAL MODELS

A large number of morphological models developed at different time- and space scales and with various degrees of detail and approximation are available in literature. In this section attention will be especially concentrated on the modelling of linear transport phenomena (see Sect. 4). Models of mass movement and surface erosion are briefly mentioned in Section 2 and 3.

6.1 Small scale models

Detailed small-scale models have especially been developed for research purposes. Many of these models (still in their infancy and needing improvements) have the scope to reproduce the movement of individual particles under the action of other particles and water flow and are usually based on a lagrangian approach. They should be able, in principle, to reproduce the behaviour of small scale systems (microforms up to the river depth-scale) and may be extremely useful to explain the hydraulic resistance mechanisms (grain and form roughness), to show the validity and limitations of transport formulae, to investigate the dynamics of movable bottom and to describe the motion of hyper-concentrated liquid-solid mixtures.

6.2 Intermediate scale models

These models are the most commonly used for practical applications. They are typically extended to the size of a river reach and applied for relatively short time durations (from one single flood event to a few years). As mentioned in Section 5, during this time all the processes at subsystem scale (microforms, hydraulic resistance, sediment transport rates etc.) are incorporated *via* simple predictors, usually “equilibrium” algebraic equations, as a function of the water discharge. Conversely, all the processes at larger scale (climate, watershed configuration etc.) are supposed to be stationary.

Intermediate scale models are obtained by averaging convection-diffusion equations for sediments are the Reynolds equations for water (in their turn obtained by averaging the water continuity and Navier-Stokes equations over turbulence) over appropriate space dimensions. The most common commercial models are *1-D* (*one-dimensional),* i.e. averaged over the river cross-section (but possibly disaggregated in a number of sub-sections). One-dimensional models can simulate bottom erosion and deposition along the river (generally the most relevant requested information), somehow “re-distributed” over the cross-section. 1-D models can easily be applied to relatively large portions of the hydrographic network.

Rather common, however, are nowadays becoming *2-D (two-dimensional)* models, i.e. averaged over the river depth, also available as commercial codes developed by several laboratories. Two dimensional models can in principle simulate all the process at the width-scale (migrating and stationary bars, braiding and bifurcations, sediment exchange with flood plains etc.). Bank collapse and reconstruction can also be incorporated in a 2-D model, which therefore will be able to reproduce meander formation and propagation.

Secondary currents over the cross section are important for localized scouring (piles, groynes etc.) and in meander morphodynamics. Their reproduction require in principle a *3-D* (*three-dimensional)* model but, at bar/meander scale, their local effect can be approximately accounted for by a 2-D model. Reproduction of density currents, often important in certain reservoirs, also requires a 3-D model. In some cases, however, a *vertical 2-D* model (i.e. averaged over the reservoir width) can also be considered.

Intermediate scale models, either 1-D, 2D or 3-D, are extremely sensitive to the *boundary conditions* to be prescribed at the upstream and downstream ends of the river reach under investigation. Correct boundary conditions for morphological models (de Vries, 1993) should be given in terms of sediment input of each grainsize fraction (at the upstream end) and in terms of either water-level or bottom-elevation, respectively for sub-critical and super-critical water flows (at the downstream end). Note that boundary conditions depend in principle on what is going on respectively upstream and downstream of the considered reach. For relatively short simulations (years), sediment input upstream can be evaluated by reasonable hypothesis based on "local" quasi-equilibrium conditions (see Sect. 7); the same can be made for water level or bottom elevation downstream. For longer simulations (centuries), however, the behaviour of the entire river system should be explicitly accounted for (Sect. 8).

6.3 Large scale models

Although 1-D models have been sometimes applied to relatively large real watersheds for specific flood events, not many examples are available in the literature of morphodynamic modelling at very long (historical or geological) time-scale, except in a few very schematized situations (simple geometry, constant waterflow, uniform grainsize). The effects of geometrical, hydrological and sedimentological *non-uniformities,* invariably present in real systems, have been only in part investigated for long-term, large-scale simulations of actual rivers and relevant watersheds. In fact, averaging "non-uniformities" of any type in non-linear equations produces "residual terms" which should be properly assessed and eventually modelled with appropriate sub-models (Di Silvio and Marin, 1996).

It may be of interest, in this respect, to explore the possibilities offered by *long-term, large-scale morphological models* where averaging is performed on time (year or number of years) and/or space (river reaches of various length). In practice, these models filter the shorter morphological fluctuations and compute only the long-term, large-scale evolution.

Long-term models have been especially developed for estuaries, but could in principle be applied also to river systems.

7. SEDIMENT TRANSPORT FORMULAS

A fundamental component of any morphological mathematical model is the predictor of the sediment transport rate as a function of sediment grainsize and flow characteristics. Sediment movement produced by channelized water flow (the so-called *linear transport*, introduced in Section 4), may be modelled in detail, from the particle scale up to the depth scale (ripples, dunes, hiding-and-exposure etc.), as mentioned in Section 6.1. In practice, what can be utilized for larger scale modelling (Sections 6.2 and 6.3) is a series of algebric equations (the so-called “transport formulas”), sometimes associated to other formulas providing the size of bedforms and the corresponding hydraulic resistance.

The available transport formulas have been obtained, above all, from laboratory flume experiments carried out in uniform conditions (uniform flow and transport and uniform grainsize). Since the early work of Du Boys (1879), a large number of transport formulas have been proposed by different authors. As all these formulas have been obtained in specific experimental ranges of flow and sediment characteristics, it is no wonder that they appear inaccurate when applied to other situations.

Assessing the prediction capability of different formulas, or even recognizing the validity limits of each one, is not an easy task. At first glance all the formulae seem to be hardly comparable. Yet it may be interesting to perform a dimensional analysis of their structure, assuming that the process is controlled by 6 *independent* variables (e.g. shear stress u_*, grainsize d, density of grain ρ_s, density of fluid ρ, viscosity ν and gravity g).

Let us define (see for example Yalin, 1977) the non-dimensional sediment transport rate:

$$T_* = \frac{q_s}{d\sqrt{g\Delta d}} \quad (1)$$

where q_s is the solid discharge in volume per unit width and $\Delta = \frac{(\rho_s - \rho)}{\rho}$ is the relative density of sediments. It follows that T_* should be a function of 3 independent non-dimensional morphological parameters; for example :

- the particle Froude number (or mobility index, or Shields parameter):

$$F_* = \frac{u_*^2}{g\Delta d} \quad (2)$$

- the particle Reynolds number :

$$\mathrm{Re}_* = \frac{u_* d}{\upsilon} \quad (3)$$

- the relative depth :

$$\left(\frac{h}{d}\right) \quad (4)$$

The particle's Reynolds number plays an important role for fine particles transported in suspension, as it controls the settling process. The relative depth, by contrast, is important for very coarse particles moving as bed load, as they can affect the free surface of shallow flows. Conversely, as for several hydraulic phenomena, the influence of both the Reynolds number and the relative depth tends to disappear when these quantities become very large. This occurs respectively for mountain rivers (high flow velocity and coarse material) and for large plain rivers (high depth and fine material).

In any case, the parameter F_* is invariably the most important one as it represents the ratio between the mobilizing effect of the water drag on the particle and the stabilizing effect of the particle's immersed weight. Most of the available transport formulae, in fact, can be approximately plotted on a graph T_* vs. F_* , either in the form

$$T_* = aF_*^{\beta} \quad (5)$$

or in the form :

$$T_* = b\left(F_* - F_{*cr}\right)^{\gamma} \quad (6)$$

assuming that the coefficients a and b and the exponents β and γ are not constant, but functions of other quantities besides F_*.

The *monomial structure* of eq. (5) is typical of many formulae like those of Kalinske (1947, Brown (1950), Engelund and Hansen (1966) etc. The *binomial structure* of eq. 6 implies that no movement occurs if the mobility (or Shield's) parameter F_* is smaller than a critical value F_{*cr} (in principle, function of Re_* and h/d). The binomial structure is typical of several popular formulae like those of Meyer Peter and Mueller (1948), Ackers and White (1973), van Rijn (1975) etc. Note that also other formulae in literature, having an apparently different theoretical background (like the ones based on minimal stream power, for instance Chang, 1977, Yang, 1976), can approximately be written as eqs. 5 and 6.

It is matter of philosophical discussion whether in principle a critical value F_{*cr} for incipient sediment transport should exist at all. Indeed, due to the stochastic character of turbulence, one may think that an (occasional) transport would even take place with extremely small (average) values of F_*. In any case, since the transport rate should rapidly decrease for very low values of F_*, the exponent β in monomial formulae needs to be rather large (and in fact it ranges between 2.5 and 3), while the exponent γ in binomial formulae is much smaller (it ranges between 1.2 and 1.5).

For both types of formulae, however, the numerical value of the exponents (β and γ) and of the coefficients (a and b) should depend, explicitly or implicitly, on the other non-dimensional parameters mentioned before, that is $Re*$ and (h/d). In fact, many of the experimental formulae contain other quantities besides F_* that affect the non dimensional sediment transport. Although these quantities are not explicit functions of $Re*$ and (h/d), they are very likely somehow related, depending upon the range of flow- and sediment characteristics in which the experiments have been carried out.

At this point, for practical applications, instead of selecting a certain available formula, it is perhaps better resorting to an expression like (5) or (6). In this case, of course, the values of exponents and coefficients should be properly chosen for the river configuration one is interested in (ranging from steep alpine torrents conveying gravel and boulders, to slow lowland rivers conveying silt and sand). This choice is in fact the transport formula calibration.

For calibrating the transport formula for a given river configuration, the simpler monomial equation (5) is preferable to the binomial equation (6), even if one has to expect for the values of a and β a larger variability than for the values of b and γ. The calibration procedure of the transport formula (Di Silvio, 1996) consists in associating to eq. (5) a uniform flow formula, either Chézy or Manning-Gauckler-Strickler, and in introducing the grainsize distribution of the bed material together with an appropriate "hiding-and-exposure" coefficient (Sect. 4.2). The hiding and exposure coefficient, multiplying the value of q_{si} (solid discharge of the i-th grainsize) in eq. (1), is assumed here as $(d_i/d_m)^s$, where $d_m = \sum \beta_i d_i$ represents the mean grainsize of the bottom material, with β_i the percentage of the i-th grainsize class present in the bottom. With respect to the uniform grainsize material, the "hiding-and-exposure" coefficient slightly augments the "intrinsic mobility" of coarse particles ($d_i>d_m$) and diminishes that of the finer ones ($d_i>d_m$).

The final expression for the total sediment discharge (sums of all grainsize classes, $i=1,2,\ldots.N$) is :

$$Q_s = \sum Q_{si} = \alpha \frac{\sum \beta_i d_i^{\,s-q}}{(\sum \beta_i d_i)^s} \cdot \left(\frac{Q^m I^n}{b^p} \right) \qquad (7)$$

where Q, I and b are respectively the waterflow discharge, the energy slope and the river width, d_i is the diameter of the i-th grainsze class and β_i is the percentage of the i-th grainsize class present in the bottom. The coefficient α incorporates all the quantities assumed as "constant" in the above mentioned procedure. The value of the exponents m, n, p and q depends on the exponent β in eq. (5) and on the selected uniform flow formula, according to the following expressions :

Chézy

$$\begin{aligned} m &= \frac{2}{3}\beta \\ n &= m \\ p &= m-1 \\ q &= \frac{3}{2}(m-1) \end{aligned} \qquad (8)$$

Manning-Gauckler-Strickler

$$\begin{aligned} m &= \frac{3}{5}\beta \\ n &= \frac{35}{30}m \\ p &= m-1 \\ q &= \frac{3}{2}(m-1) \end{aligned} \qquad (9)$$

In general, assuming a constant Chézy coefficient (i.e. a constant *relative* roughness) is more appropriate for lowland rivers (dominant dune resistance); while a constant Manning coefficient (i.e. a constant *absolute* roughness) is more appropriate for mountain rivers (flat bed and dominant grain resistance). The exponent s of the hiding-and-exposure coefficients tends to increase for strongly sorted material (mountain rivers) and to become equal to q for extremely high values of Q (equal mobility). It may be taken equal s = 0.8 in torrents and much less (down to almost zero) in many plain rivers.

Note that eq. (7) is just another form of eq. (5), in which the transport of each grainsize class present in the bottom has been considered. Equation (7) indicates that, being the other quantities constant, a biunivocal relation should exist between Q_s and Q. This is true, however, only for an experimental flume in *equilibrium conditions* (uniform flow for water and sediments): indeed, for a re-circulating flume with prescribed values of I, b and bottom composition, the transport rate Q_s is a unique function of the water flow Q.

For a real river, by contrast, different values of Q_s are measured for the same value of Q. This is basically due to the fact that the local energy slope I and the local bottom composition β_i may vary during the hydrological cycle, as fluctuating erosions and depositions invariably occur. As seen before, moreover, also the exponent s may not be constant. Finally, if the material is very fine, the material transported in suspension may be not solely controlled by the local conditions, but also by the conditions upstream (see Sect.

4.2). The last circumstance, however, is not so dramatic if the "adaptation length" is shorter than the river reach under investigation.

Eq.(7) indicates that the composition of transported material (*Qsi/Qs)* is much finer than the local bottom composition (β_i). This means that the total transport formula gives reason for a relevant transport of very fine particles, even if their presence in the bottom is extremely scarce. As it appears from eq. (7), in fact, due to their much larger mobility, the particles belonging to a very fine faction (say d_i = 50 microns) may have a very small value of β_i, but a very large value of $\left(Q_{si}/\sum Q_{si}\right)$. In other words, the notion may be misleading that only the transport of the material abundantly represented in the river bed (the relatively coarse, so-called "bed material") depends on the local conditions, while the fine material should be considered "wash-load". By contrast, even the so called "wash-load" leaves a trace in the bottom composition that can be used to compute the total transport.

In conclusion, for relatively large watersheds, the scattering of short-term measurement Q_s vs. Q is generally due to short-term fluctuation of I, β; and (probably) exponent s, rather than to the time-dependent input of fine sediments from the watershed slopes.

Indeed, if one supposes that fluctuations of the above mentioned quantities are mutually independent and assumes an exponential duration curve for $Q(t)$, the integration of (7) over one year provides:

$$V_s = \frac{\alpha}{m}\frac{I^n}{b^p}\frac{\sum \beta_i d_i{}^{s-q}}{(\sum \beta_i d_i)^s}\cdot\left[Q_o^{m-1}V_o\right] \qquad (10)$$

where V_s is the total annual transport of sediments (all classes), V_o is the annual runoff volume and Q_o the annual flood peak.

Although the hypothesis on the statistical independence of the fluctuations of I and β may be questionable, experimental applications to real measurements show a very good correlation between hydrological parameters Q_o and V_o and the annual sediment yield V_s, with an exponent m between 1.5 and 2.5 (depending on the type of river). The structure of eq. (10) is particularly convenient for calibration against sedimentation data in reservoirs.

8. SEDIMENT YIELD AND SEDIMENT PRODUCTION

One of the most difficult problems in establishing a sediment balance at watershed scale is the relationship between the sediment removed from the watershed slopes (soil production) and the soil transported by the river (sediment yield). The very same definition of those quantities may present in fact some ambiguities.

A possible, rather unambiguous, definition of *sediment yield is the total amount (mass or volume) of sediments, of any size and origin, transported as bedload or in suspension through a given cross-section during a certain period of time* (year, day, flood event, etc..). Very often, however, sediment transport is disaggregated in two parts: the so-called "bed-material transport" (typically coarser than than a conventional grain-size limit, say between 20 and 80 microns) and the so-called "washload" (below that limit). While the transport of bed-material is supposed to be a function of riverbed composition and flow characteristics, washload is assumed to be fed into the river from the watershed slopes and conveyed downstream by the river flow, with the same velocity as that of the water, i.e. without any interaction with the bottom. In many instances, washload (defined in this way) results to be a very large portion of the total transport, so that "sediment yield" it is assumed to be practically coincident with the corresponding "sediment production" during the same period of time. The distinction between bedload material and washload is obviously made for sake of simplification but, as mentioned in the previous section, it does not have a solid physical foundation. Indeed, even the finer particles have multiple phases of transport, deposition and resuspension and their average motion is by far much slower than the water's. Consequently the sediment yield of the river may be much lesser or larger than the sediment production during the same period of time.

Let us now consider the definition of sediment production. On the analogy of sediment yield, a straightforward definition of *sediment production is the total amount of sediments, of any size and origin, detached by surface erosion and mass movement, from a given location of watershed and transported downhill during a certain period of time (year, mounth, storm event etc..).* It is apparent that, in this way, sediment production is expressed as entrainment per unit surface but, in practice, it can only be measured as a transport per unit width of the watershed slope at a distance more or less remote from the closest "divide". In fact, although a number of small scale models (see Sect. 2 and 3) are available for a theoretical evaluation, it is apparent that a “punctual” measurement of sediment production does not have much sense and that some space-averaging operation should be performed over the slope surface. Experimental data, indeed, are never available point by point, but at “plot” or “field” scale (for cropland) or at “slope” scale (for natural watersheds).

As already observed, however, space-averaging is not at all a banal operation. First of all, except for extremely tiny pieces of slope surface, different transport processes occur at different scales. At a *small scale* (overland flow depth) we may observe that a thin overland flow can not maintain a stable fully two-dimensional aspect but invariably tends to concentrate into a channelized flow. This is very apparent for rills and gullies, but even diffused sheet erosion actually occurs though embryonic and intermittent micronetworks, basically controlled by vegetation. For larger and larger sizes, as it conveys larger and larger concentrated waterflow, the micronetwork tends to became more stable and well defined and to evolve towards the permanent, morphologically controlled hydrographic network. At an *intermediate scale* (experimental plot, field or natural slope) both the runoff and the sediment transport, actually concentrated along the micronetwork, are somehow integrated (i.e. averaged) over the relevant surface. A complete and reliable set of data on sediment production by surface erosion has been formed, over decades, and decades by agricultural engineers on experimental plots in many countries of the world with different soils and different crops. Experimental plots have in general a narrow rectangular surface with no transversal elevation gradient and a uniform longitudinal steepness. These data have been employed in the USA to develop the celebrated Universal Soil Loss Equation (USLE) and elsewhere around the world to adapt this formula to different agricultural and climatic conditions.

As anticipated in Section 3, the USLE estimates the sediment production, in mass per unit surface, as the product of six factors which include the length of the plot L and the steepness S. While for an *experimental plot* or even for a regular *cropland field*, ditches clearly show where they initiate and terminate, for a *natural slope* the only apparent boundary is represented by the channels and the divides of the hydrographic network. It is more practical, in this case, to define *the sediment production in a given (preferably small) hydrographic watershed as the portion of sediments, of any size and origin, detached by surface erosion and mass movement, which reaches the hydrographic network during a certain period of time (year, month, storm event, etc..).*Sediment production of the watershed can be computed by applying the same formulae (e.g. U.S.L.E.) calibrated at field scale from experimental plot data, where one assumes for the *length* and *steepness* of the natural slopes respectively the inverse of the basin's *drainage density* and *relief*. In this computation a certain reduction coefficient (*slope delivery ratio*) should be applied for taking into account the trapping effect along the natural slope, especially when the slope is quite long and its profile is undulated.

To transform sediment production into sediment yield, it would now be necessary to route the input of sediment all along the hydrographic network, down to the closure section of the watershed. With the previous definition, a distinction has been made between the intermediate scale (field or slope length) where sediment production takes place and the large scale (watershed or river length) where river processes take place. An even more aggregate definition of *sediment production is the portion of sediments … which reaches the closure of the watershed.* In this case, the computation at river scale should be affected by an even smaller reduction coefficient (*overall delivery ratio*), which should take into account also the river processes along the entire hydrographic network.

The concept of "overall delivery ratio" for sediments is somehow analogous to the concept of "runoff coefficient" for water. Yet it is much more elusive to be defined and difficult to be predicted, due to its variability in space and time along the sediment route. In fact the very notion of overall delivery ratio is not much utilized in recent literature. From the early data (Gottschalk and Brun, Shumm, Mauer, Roehl, Williams and Berndt,) it appears that delivery ratio decreases from 1 to a few percents, more or less proportionally to the inverse of the stream length (or square root of the watershed area) but scattering of data appears to be extremely high. Several attempts to have a more accurate prediction of delivery ratio as a function of the watershed and river morphology (see for instance Walling and Webb, 1996) did not provide generally valid results.

Similarly to the “runoff coefficient”, the concept of “delivery ratio” is hardly useful when it becomes much smaller than 1 (namely for watersheds larger than 50-100 km^2). The notion of delivery ratio is in fact probably acceptable exclusively at intermediate scale, namely for an overall description of the “monotone” trapping effect the watershed slopes, where very few localized permanent can only give rise to (averaged) values of the delivery ratio very close to 1.

When river processes become dominant it would probably be better substituting the static concept of "delivery ratio" by a dynamic concept of "response delay", in which the time scale also plays a role (Di Silvio and Marion, 1997). Indeed, if the watershed is large, it is not correct assuming that the very same particles detached from the watershed slopes during a certain storm can reach the closure section of the basin during the corresponding flood. The sediments moving as bedload or as suspended transport along the river (including the very fine ones, usually called “washload”), have continuously phases or deposition and re-entrainment with the river bed, banks and floodplains. Repeated deposition and re-entrainment may produce relevant granulometric, altimetric and planimetric changes at different time scale and, in any case, will strongly delay the response of river morphology (and river transport) with respect to the sediment input from the watershed slopes.

A direct evaluation of sediment yield is possible by utilizing regular (daily) measurements of turbidity and water dischange carried on at some stations along the river. This procedure assumes that there is a direct relationship between “turbidity” usually measured in one single point of the cross section and "transport concentrations" (ratio between total sediment transport and water discharge). This hypothesis is probably acceptable, especially on the long term, but it deserves further theoretical and experimental consideration (Walling and Webb, 1988).

The most precious and reliable information about sediment yield in terms of both quantity and grainsize composition, however, is given by the progressive sedimentation of existing reservoirs. The surveying technology based on the joint use of remote sensing and Global

Positioning System (GPS) has already been applied in similar circumstances. In assessing the sediment volume trapped in a reservoir, the time-dependent compaction of the deposited material should be taken into consideration (see for instance Morris and Fan, 1998). The data collected in existing reservoirs, as well as at measuring stations, may be used for calibrating reliable semi-empirical relationships (even if limited to a specific river configuration) which provide long-term sediment transport as a function of hydrological, geometrical and sedimentological characteristics of the river reach (see Sect. 7).

9. THE GEST (Global Evaluation of Erosion and Sediment Transport Processes) PROJECT

One of the main purposes of the GEST project is the evaluation of the global sediment yield, to be evaluated in a significant number of cross-sections of the main rivers of the world. Besides the present conditions, the assessment should also be repeated considering past and future scenarios.

As discussed in the previous Sect. 8, the evaluation of sediment yield can be carried out, in principle, either directly (by measuring or predicting the sediment transport rate in a given section, based on the local characteristics of the river), or indirectly, by measuring or predicting the soil production in the watershed slopes and routing this sediment down to the closure section. The application of either method depends on circumstance, as it appears from the following limit cases. For a very small and steep watersheds (e.g. a gully) and extremely fine materials (clay), the response of the system is very fast. In this case the sediment transport in the closure section (wash load) practically coincides with the soil production. By contrast, for large watersheds and relatively coarse material, the sediment yield (even transported in suspension) is quite independent from the soil production during the same event but solely depends on the local characteristics of the river.

For intermediate conditions the more or less delayed response of the river system should be taken into account. While a single (constant) “delivery ratio” appears to be inadequate for a satisfactory reproduction of this process, a relatively simple modelling of the river network may suffice.

The GEST project represents a unique occasion to assess and compare different methods. By utilizing experimental data on both the river and the watershed, collected in a variety of morphological, hydrological and sedimentological configuration, an acceptable (general) methodology might be developed, as one of the results of GEST, for predicting sediment yield at global scale where direct information is not available.

10. SUMMARY AND CONCLUSIONS

In the frame of the activities carried out by I.S.I. Task Force Group a brief review has been made of the state-of-the-art of knowledge about the dynamics of sediment erosion and sedimentation, especially regarding the following aspects:

- Surface erosion, mass movement and linear transport: three basic forms of sedimentary processes, usually approached in a different way, depending on the climatic, social and disciplinary "milieu". Taking advantage from variety and diversity of approaches.
- Sedimentary systems and sub-systems. How to cope with different time- and space-scales and how to select the necessary morphological models with different degree of details (1-D, 2-D, 3-D models).
- Increasing importance of large scale, long term processes in engineering practice.
- Sediment yield and sediment production. How can the "transfer function" be (reasonably) modeled?
- The GEST project: sediment yield assessment at global scale. An occasion for testing assumptions and methodologies.

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11. REFERENCES

ARMANINI, A., DI SILVIO, G. (1988), A one-dimensional model for the transport of a sediment mixture in non-equilibrium condition, J. Hydr. Res., vol. 26, n.3, 275-292.

ARMANINI, A. (1995), Non-uniform sediment transport: dynamics of the active layer, J. Hydr. Res., vol 33, 5.

BAUM, R.L., SAVAGE, W.Z. and GODT, J.W. (2002), TRGRS – A Fortran program for transient rainfall infiltration and Gps-based regional slope stability analysis, U.S. Geological Survey Open-File Report 02-424.

BEASLEY, D.B., HUGGINS, L.F., MONKE, E.J. (1980), ANSWERS: A model for watershed planning, Trans. Am. Soc. Agric. Engrs. 23, 938-944.

BOLLA PITTALUGA, M., SEMINARA, G. (2003), Depth integrated modelling of suspended sediment transport, Water Resour. Res. 39/5, 1137.

DE VRIES, M. (1993), Use of models for river problems, Studies and Reports in Hydrology Series, no. 51, UNESCO.

DI SILVIO, G., MARIN, A. (1996), Analytical approach to river morphodynamics : effects of space-and time-irregularities and grain-size non-uniformity. Commission of the European Communities, Directorate General XII for Science, Research and Development, Research and Technical Development Program in the Field of Environment, FRIMAR Project, Technical Report n. 2, 48.

DI SILVIO, G. (1996), Sediment yield estimates and prediction methods proceedings, Int. Conf. on Reservoir Sedimentation, vol. 2., Sect. v, 643-660, Ft. Collins, Colo., Sept. 9-13.

DI SILVIO, G., MARION, A. (1997), About delivery ratio: how does it change in time and space?, XXVII Congress IAHR., S.Francisco (USA), 10-15 August 1997. Vol. D/b, 90-95.

DI SILVIO, G. (1991), Sediment exchange between stream and bottom: a four layer model, Int. Workshop on Grain Sorting in Rivers. Ascona (Switzerland), 21-25 October 1991, pp. 163-192.

GALAPPATTI, R., VREUGDENHIL, C.B. (1985), A depth-integrated model for suspended sediment transport. J. Hydr. Res., vol. 23, 4.

MONTGOMERY, D.R., DIETRICH, W.E. (1994), A physically based model for the topographic control on shallow landsliding, Water Resources Reseach, 30, 1153-1171.

MORGAN, R.P.C., QUIINTON, J.N., SMITH, R.E., GOVERS, G., POESEN, J.W.A., ANERSWALD, K. CHISCI, G., TORRI, D., STYCZEN, M.E. (1998), The European soil erosion model (EUROSEM): A process-based approach for predicting sediment transport from fields and small catchments. Earth Surface Processes and Landforms, 23, 527-544.

MORRIS, G.L., FAN, J. (1998), *Reservoir Sedimentation Handbook*, McGraw-Hill, pp. 746.

NEARING, M.A., FOSTER, G.R., LANE, L.J. and FINKER, S.C. (1989), A process-based soil erosion model for USDA, WEPP (Water Erosion Prediction Project technology), Trans. Am. Soc. Agric. Engnrs., 32, 1587-1593.

PARKER, G., PAOLA, C., LECLAIR, S. (2000), Probabilistic Exner sediment balance equation for mixtures with no active layer, J. Hydr. Engrg., vol. 126, 818.

WALLING, D.E., WEBB, B.W. (1988), The reliability of rating curve estimates of sus pended sediment yield; some further comments. In Sediment Budgets (Eds. Bordas, M:P:, Walling, D.E.) IAHS Publications no. 174, IAHS Press, Walingford, U.K., 337-350.

WALLING, D.E., WEBB, B.W. (1996), Eds., Erosion and sediment yield : global and regional perspectives, IAHS Publication no. 236.

WICKS, J.M. and BATHURST, J.C. (1996), SHE-SED: A physically based, distributed erosion and sediment yield component for the SHE hydrological modelling system, J. Hydrol., vol. 175, n. 1-4, 213-238.

WINTERWERP, J.C., (2001), Stratification effects of cohesive and noncohesive sediment, J.Geophys. Res., vol. 106, no. C10, 22,559-22,574.

YALIN, M.S. (1977), *Mechanics of Sediment Transport*, Pergamon Press, Oxford.