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Third International Workshop on

Gravel-Bed Rivers

"DINAMICS OF GRAVEL-BED RIVERS"

Firenze - Poggio a Caiano - 24-28 Sepetember 1990

AD-A225 520

Florence July 31, 1990

Your reference: AMXSN-UK-RE/R&D 6510-EN-02 (70-1s)

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Thank you for your support.

Sincerely yours

3rd International Workshop
on
Gravel - Bed Rivers
Firenze, 24-28 September 1990

Dr. Paolo BILLI

"DINAMICS OF GRAVEL-BED RIVERS"

3rd International Workshop on Gravel-Bed rivers

Firenze - Poggio a Caiano - 24-28 Sepetember 1990

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SECRETARIAT

Dipartimento di Ingegneria Civile
Via S. Marta, 3 - 50139 Firenze, Italy
Phone (055) 4796225
Fax (055) 404375

VENUE

The workshop will be held in the Hotel Hermitage, Via Ginepraia 112, Poggio a Caiano, phones: (055) 877040-877244-8777085-8777045.

How to reach Poggio a Caiano:

From 13:00 to 22:00 a Workshop Reception Point will be specially set up in the Florence main station (Firenze Santa Maria Novella). On your arrival, please, look for it. During the same hours a shuttle service between Santa Maria Novella Station and the workshop hotel will be held by minibus; please, meet at the reception point. The organizers invite the attending people to take this chance as it is not easy to reach the conference site by public bus services.

Alternatively the workshop hotel can be reach privately by taxi at an approximate rate of Lit. 50,000.

Florence has only a small airport, while there are good train links from Rome (2 hours of travel), Milan (3 hours), and Pisa (1 hour) to Florence Santa Maria Novella.

However, for any problem please contact the workshop secretariat at the conference hotel (see above) which will be working from 9:00 am of Sunday 23rd.

USEFULL INFORMATIONS

- Obviously bank offices are closed on sunday 23rd. Currency can be changed at the hotel reception desk.
- The weather in september is commonly mild, but recent years experiences suggest to consider a wide range of conditions from rather hot and bright to relatively cool and rainy.
- No programme for accompanying persons is scheduled
- A registration desk will operate at Hermitage Hotel (Poggio a Caiano) on Sunday 23 from 9:00 am to 9:00 pm and on Monday 24 from 8:00 am to 9:00 am.
- A minibus shuttle service will operate from the conference hotel arriving to Firenze Santa Maria Novella Station on the evening of Friday 28 at 8:00 pm and on the morning of Saturday 29 at 9:00 am.
- The field trip on Tuesday 25 will start from the conference hotel at 3:00 pm. The return to the hotel is forecast at about 11:00 pm. Accompanying persons are mostly welcomed (a sweater could be usefull!).
- The field trip on Saturday 29 will start from the conference hotel at 7:00 am and will end at the conference hotel at about 9:00. On the way back, a stop at Firenze Santa Maria Novella station can be scheduled on request (again don't forget your sweater!).

RESPONSABILITY

The organizing committe takes no responsability for any damage to attending people, baggage and properties.

TECHNICAL PROGRAMME

Scientific sessions will usually start at 9.00 a.m and will end at 5:30 p.m.

In the conference room a 5X5 slide projector and a overhead projector will be available.

Those having produced scientific videos on gravel-bed rivers and related aspects are kindly invited to present them on the evening of Wednesday 26 using the standard european VHS and the Video 8 systems.

Monday 24 September

Morning

I INTRODUCTION AND KEYNOTE ADDRESSES

1. G. Wolman: Gravel-bed rivers. Some possible keynotes
2. M. Newson: Gravel-bed rivers. floods and thresholds

II SEDIMENT YIELD DYNAMICS

3. J. Bathurst, S. White: Modelling sediment yield
4. S.Moretti. F. Brunori, L. Chiarantini: Modelling sediment supply
5. G. Leeks: Fluvial sediment transport and plantation forestry: monitoring and application - Case studies from upland Mid-Wales, UK

Afternoon

III DEBRIS FLOW DYNAMICS

6. F. Iseya. H. Ikeda, H. Maita: Fluvial deposits in a torrential gravel-bed stream by extreme sediment supply: sedimentary structure and depositional mechanism
7. P. Canuti. C. Garzonio. G. Rodolfi: The historical and catastrophic debris flow phenomena of Mt. Falterona (Tuscany. Italy)
8. D. Rickenmann: Analysis of massive sediment transport processes in torrents

IV ALLUVIAL FAN AND STEEP STREAM SEDIMENT PROCESSES

9. A. Harvey: Sediment transport processes on alluvial fans: influence on fan morphology
10. A. Schick, J. Lekach: Bed material conveyance in a flash-flood system: magnitude, frequency, mechanics
11. P. Carling, A. Kelsey, M.S. Glaister: Effect of bed roughness, particle shape and orientation on initial motion criteria

Tuesday 25 September

Morning

V CHANNEL MORPHOLOGY DYNAMICS

12. C.R. Thorne: Flow processes in gravel-bed bends
13. M. Colombini, M. Tubino, P. Whiting: Topographic expression of bars in meandering channels
14. L.B. Leopold: The sediement size that determines channel morphology

Vi BRAIDED CHANNEL DYNAMICS

15. R.I. Ferguson, P.J. Ashworth: Spatial pattern of bedload transport and channel change in braided and near-braided rivers
16. J. Laronne, M.J. Duncan: Bedload transport path and gravel bar formation
17. P. Ashworth, M. Powell: The mode of medial bar formation or the sorting processes in braided rivers.

Afternoon

VII DOWNSTREAM FINING

18. A. Werritty: Downstream fining in a gravel-bed river in S. Poland: lithologic controls and the role of abrasion
19. A. Armanini: Variation of bed and transport mean diameters in erosion-deposition processes

- Field trip to Virginio stream

Wednesday 26 September

Morning

VIII GRAIN-SIZE DISTRIBUTION

20. H. Ibbeken: The mystery of bimodality: falsification of current concepts, proposal of a new one
21. I.Becchi, P. Billi, A. Moro: Grain-size distribution characteristics: an advanced problem for statistical tools
22. K. Bunte: Counting instead of weighing: grain-size composition of coarse material bedload in a mountain stream expressed in particle number transport rate

IX ENTRAINMENT AND MOVEMENT AT LOW TRANSPORT RATES

23. E. Andrews: Marginal bedload transport rates
24. P. Diplas, G. Parker: Deposition and removal of fines in gravel-bed streams
25. M. Kondolf: Salmonid spawning gravels: size distribution and modification by spawning fish

Afternoon

X ARMOURING

26. B.B. Willetts. S.J. Tait, J.K. Maizels: Laboratory observation of bed armouring and changes in bed load composition
27. E. Paris, A. Lamberti: Analysis of armouring processes through laboratory experiments
28. T.Lisle. M A. Madej: Spatial variation in armouring in a channel with high sediment supply

XI MOVEMENT OF MIXED-SIZED SEDIMENT

29. P.D. Kumar, Shyuer-Ming Shih: Equal grain mobility versus changing bedload grain size in gravel-bed streams
30. P.R. Wilcock: Bed-load transport of mixed-size sediment
31. R.A. Kuhnle: Fractional transport rates of bed load on Goodwin Creek

Evening

- Video session: informal projection of videos on gravel-bed rivers

Thursday 27 September

Morning

XII MONITORING OF PARTICLE MOVEMENT: NEW TECHNIQUES

32. P. Tacconi. M. Rinaldi, S. Moretti, M. Matteini: Monitoring of particle movement on Virginio gravel-bed stream.
33. M.A. Hassan. M. Church: The movement of individual grains on the stream bed
34. W. Emmett: Movement of coarse particles monitored by micro radio transmitters

XIII MODELLING SEDIMENT TRANSPORT

35. G. Di Silvio: Modelling sediment transport: dominant features to be simulated in different hydrological and morphological circumstances
36. R.R. Copeland, W.A. Thomas: Numerical modelling of gravel movement in concrete channels
37. A. Lamberti, L. Montefusco: Sediment transport in steep plane beds: a general model.

Afternoon

XIV BED DYNAMICS

38. I Reid, L.Frostick, A.C. Brayshaw: Microform roughness elements and the selective entrainment and entrapment of particles in gravel-bed rivers
39. P. Ergenzinger: River bed adjustments in a step-pool system (Lainbach, Upper Bavaria).

40. R. Hey, D.J. Needham: Dynamic modelling of bed waves

Evening

- Workshop banquet

Friday 28 September

Morning

XV ANALYSIS OF HISTORICAL CHANGES

41. C. Fritz, G. Giada, V. Villi (IT): Long term sediment budget in an alpine catchment
42. J. Hooke, C.E. Redmond: Causes and nature of river planform changes
43. M.G. Macklin, B.T. Rumsby, M.D. Newson: Historic floods and vertical accretion of fine-grained alluvium in the lower Tyne valley, North East England

XVI
HUMAN IMPACTS

44. M. Jaeggi: Effect of engineering solutions on sediment transport
45. D. Sear: The effect of river regulation for hydro-electric power on sediments and sediment transport within riffle-pool sequences
46. B.R. Hall, W.A. Thomas, M.L. Pearson: Computing bed-load discharge and channel adjustment in a cobble bed river for flood control channel design
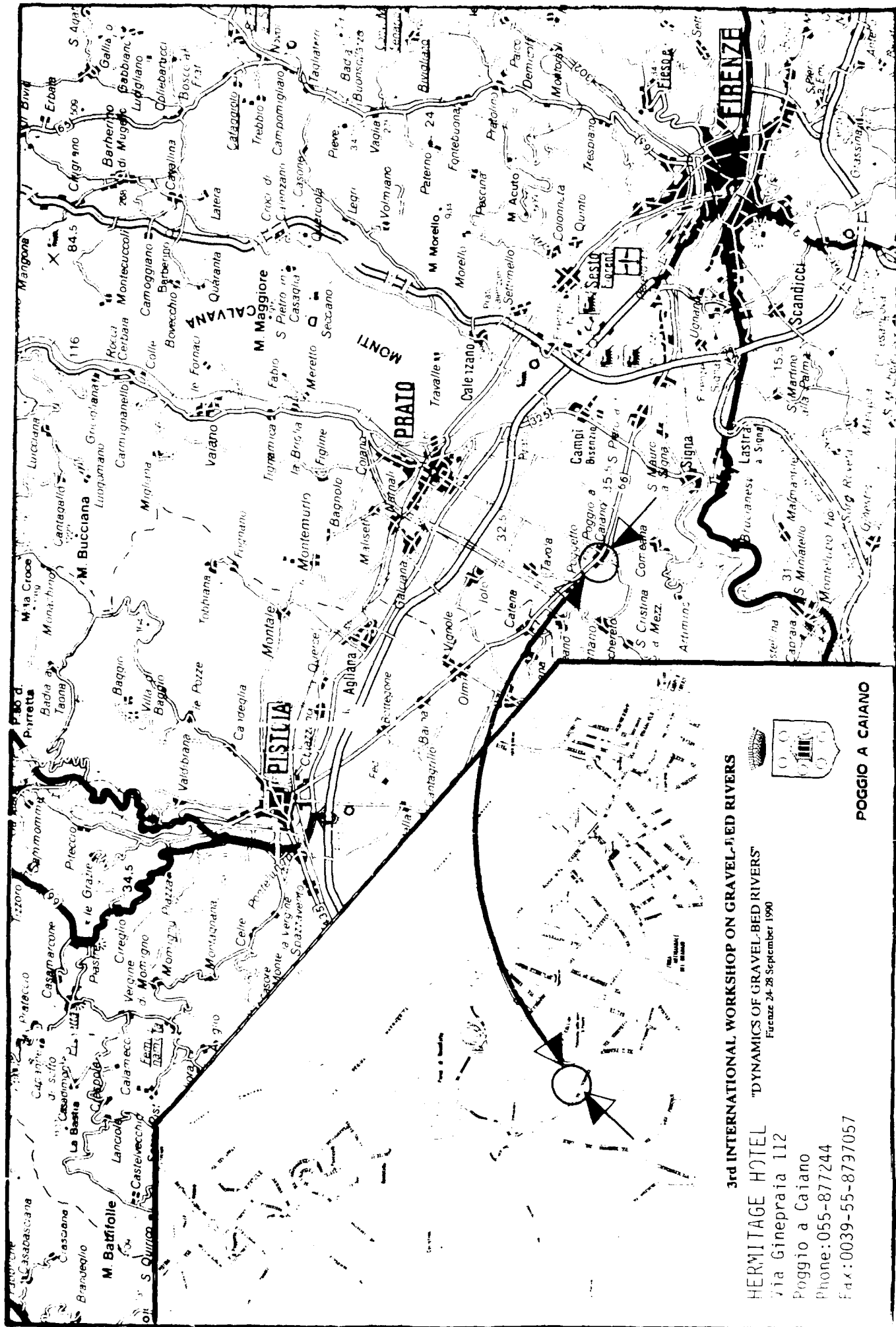
CONCLUDING REMARKS

47. L. Leopold: Overview of gravel-bed river dynamics.

Afternoon

- Departure for people not going on the field trip

Saturday 29 September
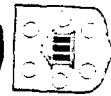
- Field trip: Gravel-Bed River Dynamics in Southern Tuscany

3rd INTERNATIONAL WORKSHOP ON GRAVEL-BED RIVERS
"DYNAMICS OF GRAVEL-BED RIVERS"
Firenze 24-28 September 1990

HERMITAGE HOTEL
Via Ginepraia 112
Poggio a Caiano
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POGGIO A CAIANO

Gravel-Bed Rivers
Some Possible Keynotes
M. Gordon Wolman
The Johns Hopkins University
June, 1990

Everyone knows what a gravel-bed river is, yet describing the particle size distribution on the bed of a river tells only a small part of the story. The patterns of gravel-bed rivers range from braided channels with many branches to straight flumes or canal-like natural rivers. The banks of gravel-bed rivers may be composed of gravel or a wide variety of materials and particle sizes ranging from clay and clay sizes to mosses and sedges. The gradients of gravel-bed rivers may be very steep or quite gentle as in many trout streams treasured by local fly fishermen.

Not only do gravel-bed rivers range in size, plan, profile, and cross section, but our interest in them spans a similar range. At one extreme human concern for gravel-bed rivers focuses upon attempts to maintain the "natural" river and riparian system in a world dominated by human activity. At the opposite extreme, the concept of control is evoked to fix the position and form of an unruly natural stream to prevent it from behaving in a "natural" fashion, to preserve the adjacent lands from erosion, or to assure a fixed channel to speed the passage of floods or to assure navigability.

Given the variety of gravel-bed rivers and the range of objectives inspiring interest in such rivers, few specific topics clearly warrant priority of inquiry over others. In the preface to the remarkable volume representing the deliberations of the Second International Workshop on gravel-bed rivers it is noted that at the First International Workshop in 1980, "the major obstacle to achieving goals (of better management guidelines) was identified as a lack of knowledge concerning sediment transport processes..." Both understanding of process and its application to management require continuing work refining the understanding of transport processes which control the behavior of gravel-bed rivers. The second conference made major contributions to this understanding. Despite such progress, uncertainties remain about sediment transport even when "gravel" alone is transported in the bed of the river channel. As the papers in the 1987 volume made clear, the full range of hydrologic and hydraulic complexities is also reflected in the behavior of gravel-bed rivers.

Papers in the second workshop contain careful evaluations of transport models along with cautions about the variety of materials in gravel-bed rivers, possible effects of ranges in particle size including suspended load, and the relative confidence in the application of transport equations to conditions where the channel geometry is assumed and water flow and sediment movement calculated, as opposed to those conditions estimates of channel features are to be derived from principles of hydrology and hydraulics in channels with mobile boundaries. Has a consensus emerged on the present

state-of-the-art of sediment transport estimation and observation?   Is it premature to provide confidence intervals for the estimation of sediment movement in gravel-bed rivers, qualified or structured according to a rough classification of the variety of gravel-bed rivers?

Equilibrium concepts applied channel form and plan have assumed steady state and uniform flow in characterizing the relationship between geometry and flow in specific reaches. From the outset, it has been known that these assumptions are simplifications of the natural world. Analogies between regime canal systems and natural rivers have recognized the disparity in hydrologic regimen between the two while marveling at the apparent similarity of the resulting river features. Design criteria for river control mimicking canal design have sought to minimize problems associated with the variability of flow and transport in natural rivers. "Dominant" and "channel-forming" discharge are surrogates for a more complex reality, and recognized as such. Is sufficient information now available to do without the single flow concept implied in dominant, channel-forming or bankfull discharge in the hydrology relating to river form or pattern?   The literature on flow and transport variability grows, but the association of flow variability and channel behavior is more weakly supported. The time history of channel behavior coupled with flow variation is poorly known and hence models of temporal change are hard to verify. Perhaps it is time for a structured inquiry designed at an international level to monitor carefully both flow and channel behavior in a well chosen sample of gravel-bed rivers.

As one moves from interest in the geomorphic characteristics and behavior of gravel-bed rivers to a focus upon society's demands and interest in such rivers, the scope of inquiry expands. A simple classification of the range of interest can be shown based on an artificial dichotomy between the natural and the controlled river (Table). The dichotomy is artificial because even the so-called natural channel is beginning to require management either because human beings really do not like the "natural", or because the natural channel system is being markedly altered by human activities on most areas of the drainage basin which control the downstream fluvial system. In addition, one cannot separate the channel itself from the riparian margin. The channel alone, however, has received the lion's share of attention in research and management. Moreover, in each domain, the aquatic and terrestrial, there are plants and animals whose habitat and behavior are essential features of the management of both natural and controlled river systems.

The spatial and temporal characteristics of fish habitats are receiving increasing attention. Riparian terrestrial vegetation, particularly trees, have been the subject of intense study; first from the standpoint of evapo-transpiration and water loss, and more recently because of the relationship between water loss and instream flow requirements. Aquatic plants, rooted and lotic are vital to the habitat and to the quality of the river waters themselves and, of course, vegetation has long been recognized as an important component of boundary material confining the channel cross-section. Relatively few observations, however, quantify these effects. While not new, emphasis upon the riparian scene as a whole, including the biota as well as upon the channel, appears to be critical for understanding and hence, management of gravel-bed rivers. The humorist Robert Benchley discussed an

international fisheries agreement from the standpoint of the fish. Perhaps a view of the gravel-bed river from the standpoint of a fish, a bird, or a tree plant might be useful in suggesting hypotheses to be tested of observations to be made of different parameters in different ways. What are the interactions of biota, channel, and riparian margin?

Managing the controlled or regulated natural channel whether for navigation, erosion control, gravel extraction or recreation faces the same issues as those encountered in a natural scene. Further, multiple use, while sometimes a euphemism glossing over conflicting objectives, dominates in many gravel-bed rivers. Such multiple objectives can be seen where groins are used in bank control and the habitat for birds within the slack water between the groins becomes a major environmental objective issue. A habit of visualizing the controlled or natural river from the standpoint of the fish, the birds, and the trees might mitigate the likely regret associated with failure to satisfy a variety of social values expressed by society as such values change over the years. The relationship between bank erosion control, floodway design, and habitat for biota might provide a framework for review and summary in a more sophisticated handbook for the management of some gravel-bed rivers than currently exists.

As noted earlier, the words gravel-bed rivers encompass an enormous range of channel and riparian characteristics. The continuum represented by both the downstream channel system within a drainage basin and by the variety of channel forms and patterns is well known. At the same time, the challenge of classification based on quantitative measures remains. We are hard pressed to quantify the description of bank material in the natural scene, let alone relate the full range of hydrologic behavior, geologic setting, and temporal variations to one another as determinants of channel form and pattern. Yet, in the absence of quantitative classification, it is difficult to transfer information derived from experience and observation in one place to that in another. In theory, it is correctly assumed that understanding and hence modelling based upon first principles is the basis for transferability. But, we also recognize the nearly infinite variability of the natural scene which requires reasoning by analogy as well as by the application of principles. Any classification of a continuum will require arbitrary "splitting or lumping", as the paleontologists recognize, but diagrammatic representation of channel forms would be much improved if the axes were quantitative rather than qualitative. Can this be achieved today for gravel-bed rivers?

SOCIETY'S INTEREST IN GRAVEL-BED RIVERS

Objective	Maintaining the Natural Scene	Controlling the River

Specific Uses	Preservation of Ecosystem
	Recreation
	    Fishing
	    Hunting
	    Hiking
	    Walking
	    Viewing
		Navigation
		Erosion Control
		Floodway
		Gravel Mining
		Recreation
		    Fishing
		    Hiking
		    Biking
		    Walking

Location	Channel Riparian Margin	Channel Riparian Margin
Biota	Aquatic Terrestrial	Aquatic Terrestrial

GEOMORPHIC THRESHOLDS
IN GRAVEL BED RIVERS-
REFINEMENTS FOR AN ERA OF ENVIRONMENTAL CHANGE

by
Malcolm Newson
Professor of Physical Geography
University of Newcastle upon Tyne
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Contents

1. Introduction

In 1973 Schumm introduced geomorphologists to the concept of thresholds in the fluvial sediment system. The concept has spread to all facets of the subject, often without question as to its true applicability; those who work with gravel-bed rivers, with flood effectiveness and with engineering applications of geomorphology are prone to unquestioning acceptance. The present author is therefore particularly vulnerable to "threshold spotting", and at the outset it is essential to list the qualitative appeals of the concept to the geomorphologist:

> a. Threshold concepts work best at the basin scale. Intrinsic "geomorphic" thresholds establish the value of geomorphological study as a valid, even crucial, contribution to the multidisciplinary field of fluvial systems,

> b. Thresholds are paradigmatic, representing a useful philosophical template for observations, located between castastrophic and uniformitarian explanations of the landscape. Thresholds normalize the role and relations of fluvial studies with other sciences which have employed the concept with great success.

The prospect of an impending era of marked changes in extrinsic climatic variables and of a determined anthropogenic response makes this is a good time to evaluate the validity and utility of threshold concepts in gravel-bed systems (these two values are not perfectly correlated!) . River systems appear to have caught the imagination of politicians and planners concerned with environmental management; possibly the river basin outline represents a helpful circumscription to "the problem". The balance between anthropogenic and climatic influences in river basins is highly debated (Newson & Lewin, in press); if the former already dominate, there is less apparent need to "prepare" for global warming and, if the latter dominate, we need to decide urgently whether and how to increase the anthropogenic imprint in order to conserve human resource systems. A new alternative has also emerged: that of a sustainable, mainly natural response in which we largely withdraw from some or all of the hazardous locations in river basins in favour of the conservation of nature (Newson, in press). To all of these scenarios there attends the need for predictions of the "where and when" of river responses.

The threshold concept has impinged on the public debate on climate change, notably through the spectacular example of the potentially threshold-like behaviour of the West

Antarctic Icesheet and the results of its melting (Grove, 1987). However, in the decade since the highly enthusiastic Binghampton Symposium devoted to the subject, (Coates & Vitek, 1980) thresholds in geomorphology have been increasingly questioned (e.g. Carlson, 1984; Ferguson, 1987), either because threshold behaviour at one scale becomes progressive at another or because the abruptness of change implied by the concept itself becomes questionable after the arrival of observations of intermediate conditions and forms.

The dust-jacket of the volume containing papers from the 9th Annual Geomorphology Symposium at Binghampton (Coates & Vitek, 1980) proclaims that "critical limits, boundary conditions, and yield points - indeed thresholds" are important in many fields of science and technology. Although a stricter definition was offered at least once inside the covers ("within the natural system boundary conditions exist that when exceeded can cause sudden and vast changes") such definitions have been little refined, and even less adhered to. Table 1 lists the alternative definitions of thresholds found in the Binghampton volume; it is not offered as supercilious criticism, but instead to demonstrate that the threshold concept is very compelling - indeed those criticising it often fall to using it within their critical writing (e.g. Ferguson, 1986). Coates and Vitek (1980) themselves tacitly admit to a lack of rigidity in definition when they say that "'the judgment criteria that are adopted may determine whether to label an event on landform change as a true threshold" (p 11). Nevertheless, they are in optimistic mood, finding that "thresholds are present in all geomorphic processes" and that, in an era of increasing application for geomorphology "ignoring thresholds invites tragedies in every aspect of human utilization of the surface" (p 21). It is the latter remark which now needs some substantiation in this review.

TABLE 1: Definitions of thresholds offered at the
          1980 Binghampton Symposium

"The point at which a stimulus begins to produce a
response"   McKerchar (1980) p 171

"a threshold is a turning point or boundary condition
that separates two distinct phases of interconnected
processes, a dynamic system that is powered by the
same energy source"   Fairbridge (1980) p 48

"That critical distance from equilibrium at which
unstability may occur therefore represents a thermo-
dynamic threshold beyond which a perturbation
introduced to the system does not decay to the
steady state ..."   Karcz (1980) p 220

"Although thresholds are generally defined as an
abrupt temporal change in morphology or dynamics
of landforms, similar abrupt spatial transitions
are common (stream banks, escarpment brinks, riffles
and pools, stream junctions etc.).
                                 Howard (1980) p 227

"A threshold may be regarded as a balance betweeen
opposing tendencies"   Bull (1980) p 260

"'Threshold' or 'threshold concept' are employed
as self-explanatory terms to denote existence of
zones or critical conditions at which change in
topographic form or process occurs"
                                    Ford (1980) p 345
                            quoting Frederking & Vilek.

"The Nile seems to be a classic example of a
hydrologic threshold"   Fairbridge (1980) p 45

2. Thresholds: original scales and definitions

Schumm (1973) began his influential paper on the threshold phenomena of river basins with the words, "The alluvial and morphologic details of drainage systems are much too complex to be explained by progressive erosion alone." (p 299). He wrote as a geologist and gemorphologist - about alluvium and landform. At our last gathering in Pingree Park (Thorne, Bathurst & Hey, 1987) the present author made a plea that each contribution be accompanied by a time- and space-scale cartoon locating its disciplinary origins in space and time - to avoid fruitless debate between the "zones" of Schumm and Lichty's (1963) table of dependent and independent river basin variables. Such attention to detail is also a requirement of any review of the origins and subsequent critiques of the threshold concept. To place this (geomorphological) contribution on the working diagram we need to refer to Schumm's original definition of geomorphic thresholds:

"A geomorphic threshold is one that is inherent in the manner of landform change; it is a threshold that is developed within the geomorphic system by changes in the system itself through time" (Schumm, 1973, p301)

In Schumm's treatment the external variables change progressively (much of the work having been done experimentally in the CSU basin-scale "facility"), but a complex response is evoked by the interaction of sediment supply and transport processes within the fluvial system. Refining the concept, Schumm (1977) is quick to point out that the alternating phases of erosion and deposition he defines as comprising complex response do not simply represent the system hunting for an equilibrium: "the complexity is not simply negative feedback that restores a former condition: rather it involves the crossing of a threshold to a new equilibrium state" (p 77). Schumm tends to restrict the utility of geomorphic thresholds and complex response to areas of high sediment production or of "rugged, youthful topography".

In view of these apparent restrictions the main task of the present review becomes one of investigating whether the original constraints placed on the applicability of threshold concepts have been validly breached or not and, if not, whether it is important that we now localise genuine threshold phenomena further in time and space because of their practical potential in interpreting and responding to environmental change.

The main extensions to the threshold concept appear to have been largely away from the fluvial system itself; thresholds are applied particularly to processes which include mechanical failures or overflow of storages (or a

mixture of both). Graf (1982) examining the potential of catastrophe theory in geomorphology concluded that, with the possible exception of arroyo formation, rapid system changes were much more likely in mass movement, aeolian processes, glacial surges and breaking waves; in other words, where the mechanical action of failure is critical. Thresholds have also been associated with the magnitude/frequency spectrum of extrinsic variables (and we may include anthropogenic activity as well as storms and floods). More in-keeping with Schumm's geomorphic thresholds has been the extension of threshold concepts to the interface between processes leading to sediment production/supply and those producing sediment transport Intrinsic thresholds, in particular, may be associated with many of the multitude of storages associated with cascading natural systems such as those of river basins: "Where catchment processes incorporate storages a new mode of operation commences, or a new process is included when a storage is filled" (McKerchar, 1980, p 172). Kirkby (1980) writes of "domains of dominance" for particular processes, with the changeover from one to another commonly progressive, but which "in several important cases involves an instability which sharpens the transition" (p 53). Even if the processes operate in a linear fashion, says Kirkby, the change of domain produces a non-linearity and this is especially true of storage/release processes.

To the present author's knowledge there has been but one attempt in geomorphology to refine a working definition of threshold behaviour in terms of generic attention to process behaviour (one of Graf's obstructions to the application of catastrophe theory in the subject was our ignorance of control processes). At our last Workshop Hey (1987) suggested that the normal interaction of load and transport in rivers could produce threshold effects whilst Pitlick in his discussion of Hey's paper concluded that a reach downstream from a catastrophic dam burst had not shown threshold responses! However, Chappell (1983) draws attention to an important divergence in the nature of thresholds; defining a threshold as the boundary between different states of a system, he goes on to define them as transitive or intransitive according to whether the new state represents a persistent or short-lived change of external boundary conditions. Chappell implies that the "perturbation of boundary conditions" which occurs over an intransitive threshold is one "within the normal magnitude - frequency spectrum". "Bistable behaviour" can occur in some systems containing intransitive thresholds, as defined by the process-resultant curve. The crossing of transitive thresholds is, however, more akin to Karcz's (1980) thermodeynamically-defined perturbation which does not decay to steady state.

Chappell stresses the need to identify thresholds at the process level because apparent boundaries in geomorphological systems can masquerade as thresholds. Concluding, he says that any study of geomorphological change must involve
- a) the boundary magnitude - frequency distribution
- b) the range of sensitivity of the threshold process
- c) the system lag and feedbacks

3. Thresholds: developing doubts and criticisms

Threshold interpretations may now be said to have come to occupy a number of niches in the fluvial system, as shown by Table 2:

Table 2 Summary table of threshold interpretations within the time and space scales of the fluvial system

PARTICLES	SEDIMENT YIELDS	PATTERNS	FORMS

Initiation/cessation of transport
        sedimentation features/zones
                channel pattern/metamorphosis
                        climatic geomorphology

One may consider three factors which have helped extend the popularity of the concept:

a. In hydraulics there are thresholds between sub- and supercritical flow regimes; thresholds also said to describe the entrainment, transport and deposition of sediment grains, indeed there are "threshold channels" designed to throughput sediment whilst retaining form. Hydrologists have produced more evidence for non-progressive behaviour of runoff processes and the "rare, great" flood event has been a major focus of geomorphological study.

b. The methods available to those studying the fluvial system may have influenced the adoption of threshold concepts; geomorphology is rooted in descriptive classifications and "classic" forms dominate over intermediate states. "Snapshot" historical evidence of fluvial change from maps, air photos etc. is an essential methodlogical component. In process studies, trapping of sediments has necessarily discretized what may be continuous properties.

c. There has been something of a cultural proclivity in late 20th century science to interpret rapid change - for example catastrophe theory (Woodcock and Davis, 1980), Chaos (Gleick, 1988) and mass extinction (Albritton, 1989).

By the mid 1980's, however, doubts were growing over the application of threshold concepts, especially to the transformation of channel planform.

Leopold and Wolman's (1957) location of planform domains for channel patterns on a slope/discharge plot attracted Schumm's attention as a candidate for the threshold concept; however, his 1973 paper does not expressly classify the transition from meandering to braided planforms as representing a process threshold. Clearly the importance of slope attracted him to the potential for a true geomorphic threshold but he did this only at a valley scale, not at a single site through time. The temporal reference in the paper is to valley metamorphosis such as that which occurred in the Cimmaron River (Schumm and Lichty, 1963); it therefore implies a change in sediment load calibre as the result of crossing the threshold.

Ferguson (1987) says of the Leopold and Wolman paper: "Despite its authors' emphasis that natural channel patterns form a continuum..... the paper is remembered mainly for its graphical discrimination between braided and meandering channels in terms of just two controlling variables, discharge and slope" (p 129). Ferguson also writes, "Surprisingly, few fluvial scientists have paused to ask whether Leopold & Wolman's threshold between meandering and braiding is quantitatively, conceptually or even morphologically correct. Ferguson is firmly of the view that "we should think in terms of transitions in channel pattern rather than sharp thresholds". The evolution of bar forms, leading eventually to channel change, is also said to involve transitions rather than thresholds, although the shift in chute location across point or lateral bars might well be considered by Chappell to be worthy of the "intransitive threshold" classification. Furthermore, any channel observed in the field, says Ferguson, is likely to be recovering from the last major flood during which an essentially equilibrium channel form resulted. Again, the Chappell terminology would label the flood channel and recovering low flow channel as existing either side of an intransitive threshold unless the flood had set up entirely new catchment or reach dynamics (e.g. by creating a new river cliff source of sediments). A similar point was made by Harvey, Hitchcock and Hughes (1982) - see Section 4 below.

Carson and Griffiths' (1987) impressively practical review (from the perspective of the management of New Zealand's extensive gravel-bed systems) incorporates a criticism of two threshold concepts - the initiation (and cessation) of gravel movement and meandering/braiding channel planforms. They vindicate Ferguson's view that a more complete field survey of channel types would yield considerable evidence of intermediate forms, for example the "wandering" channels classified by Neill (1973). Carson and Griffiths stess (p 119), "in attempts to differentiate meandering and braided channels in terms of some 'threshold', care is needed not to include such channels misclassified as one of the two end-members".

Turning to the mechanisms of the "wandering" channel, Carson and Griffiths report that the locus and pattern of secondary flows in the asymmetric bends of many gravel-bed meandering reaches produce both upvalley migration between bends and local shoaling of the thalweg in the bend-entry zone. As is demonstrated below for northern England, such "fill and cut" behaviour, producing a new thalweg is an important aspect of channel change both laterally and vertically. The lateral activity is far less likely to cross a transitive threshold than the vertical activity if it occurs.

Carson and Griffiths restate Carson's (1984) conclusions that channel classification into threshold types is illusory, that the conventional discharge/slope threshold is one of particle size and that there is weak support for an association between braiding and high slope-discharge values.

In terms of the initiation of movement the New Zealand review supports the view that "the critical mean velocity for the onset of bed material movement is significantly higher than that needed to maintain movement once particles have been entrained by the flow". (p 10). There are now significant proofs from the field that bed condition (overloose/underloose - Church, 1978), bedforms (Reid & Frostick, 1986) and particle exposure (Andrews, 1983) all produce this (intransitive) threshold-like effect. However, particularly in the wider channels reviewed by Carson and Griffiths, it is clear that whilst there is theoretical support for grain transport thresholds, the inter-relationships between mean and convective stresses and variability of particle size on the bed mean that bed movement may in reality be progressive.

There is clearly a dire need to measure the specific location of bed material entrainment and routes taken during bed material transport so as to confirm or refute the hypothesis that the apparent thresholds observed by trapping do not result from spatial variability of a bed material

source and of the trajectory of movement. Thus the bed
material discharge may be more or less progressive but the
yield from a reach may be represented by apparent
thresholds. Meigh (1987) pointed out that channel storage
and confinement of transport routes within the cross-section
of a "near threshold" gravel-bed channel produced a temporal
record of irregular throughput. Of equal relevance to the
argument is the role of the supply-limited "throughput"
material (Carson & Griffiths' term); because it is supply-
limited and because bank collapses are a key mode of supply,
its transport will appear quasi-threshold. If we have erred
in applying threshold concepts to channel patterns because
of classification problems we may have made errors in
assigning sediment transport phenomena on the basis of
observations which must be highly dependent upon sampler
design and sampling position within a reach.

4. Field evidence of basin-scale threshold behaviour during extreme floods: Northern England

At the outset of this review we considered the
importance of the threshold concept to an era in which
climatic change scenarios seem destined to force river basin
managers into a greater familiarity with equilibrium and
instability concepts. It seems reasonable, therefore, for
the case of Northern Britain (in which both average rainfall
and "storminess" are likely to increase) to examine the
recent flood record for gemorphological changes which might
fit the more rigorous of the threshold definitions presented
above. It is of particular interest in the two field areas
discussed below to consider the additional problems of
anthropogenic effects, especially via biogenic controls on
the important feedback loops which may control the
transitive or intransitive nature of a threshold event. The
relationship between lateral and vertical responses is also
of interest.

The "Hurricane Charley" floods in Northern England in
August 1986 were remarkable for their geomorphological
effectiveness throughout major drainage basins (Newson,
1989), with vertical channel adjustments dominating over
planform change (by incision in the headwaters and
floodplain accretion down-river). Geomorphological studies
of the affected headwater areas were made to investigate the
process thresholds which might have resulted in incision
during the event. Schumm, Harvey and Watson (1984) have
referred to the fact that "when geomorphic thresholds are
exceeded incision will occur" (p 11/12). .

The conclusions reached by Newson and Macklin (1990)
are that, although the Charley event may have been
relatively common in the hydrological record (estimated 50-

year return period) it produced very high rates of coarse
sediment transport, partly because of the availability of
mining wastes in some valleys. Although large fan-like
depositional zones were created where headwaters joined
larger valleys, incision through valley fill and bedrock
predominated. Bedrock incision appears to have been
lithologically-controlled and intrinsic thresholds were
crossed as alternating sandstone and shale strata were
breached in the narrower, confined reaches of the affected
valleys. A basic process leading to incision in boader
valley reaches appears to be the "fill and cut" sequence
which occurs at minor depositional storage sites, for
example in the short meandering reaches. At the upstream end
of such reaches flood deposition causes afflux which causes
incision across the valley floor; if the incision(s) headcut
successfully within the duration of the peak flow there is
positive feedback to further incision and this proceeds to
the upstream reach. Coxon, Coxon and Thorn (1989) describe
an idential sequence of channel blocking, chute cutting in
the Yellow River (Ireland) floods of 1986 (return period
several hundred years). After the event, the sediment
supply to the flood-torn channels in such valleys is
entirely changed; in the bedrock reaches it may be reduced
because the weathering profile has been stripped, but
elsewhere the incised stream flows between unconsolidatea
banks and there are fresh bluffs and slope failures to form
additional point sources of glacial material. A transitive
threshold has clearly been crossed in respect of the
"normal" magnitude/frequency relationship.

The rapidity of the incicion by these northern Pennine
streams is confirmed by small terraces which have been dated
by lichenometry performed on the flood boulders which mantle
them. Both this positioning of boulders and the concordance
of lichenometric dates with large floods from the
documentary record lead to the conclusion that phases of
incision represent transitive threshold changes in processes
operating in these systems. These are non-engineered
fluvial environments and the prime human influence appears
to be on the intrinsic threshold behaviour of the basins
where mining waste has been redistributed by fluvial action.
Mining has had another influence in that much of the waste
is polluted with metals and this prevents the biogenic
stabilization of near-channel sediment storages between
threshold-crossing events.

The Forest of Bowland study by Newson and Bathurst
(1990) was prompted by severe, costly management problems
resulting from both erosion and deposition within two
Victorian water supply schemes - the Brennand and Whitendale
catchments. The North West Water Authority was alerted to
the possible impact of land-use and land-management factors
on the catchments by the fact that sedimentation rates to
the nearby Abbeystead Reservoir had almost doubled since

approximately 1932; drainage, roading, vegetation control and afforestation seemed to be the most likely factors. They needed a remedial explanation for the damaging instability, particularly in the Brennand catchment, and Newson & Bathurst (1990) conclude that the explanation may be the crossing of a sediment transfer threshold by a spectacular flood in August 1967.

The flood-producing rainfall of August 8th totalled 109 mm. in the Brennand catchment and 89 mm. in the Whitendale, with a duration of approximately 90 minutes. The estimated discharge peak in the Brennand was 15-16 times the mean annual flood, equivalent in this region to a return period in excess of 10,000 years! Accounts written at the time speak of mudslides, landslides, boulder dams, channel diversions, severe erosion, "phenomenal" movement of bed material and, significantly, heavily-scoured channels of enlarged dimensions. Aerial photographs for 1945, 1948 and 1968 confirm almost complete valley-floor adjustment by channel switching or overbank deposition. New exposures of sediment include a large number of river cliffs created by mass movement and incision (but maintained by the latter process) and extensive moorland gullies stripped of their peat cover.

The present-day channel network of the Brennand catchment has, therefore, as a result of the one event, much larger sources of sediment supply than at any time since the growth of the present vegetation cover. A transitive process threshold was apparently crossed in that both supply and transfer of gravel material became adjusted to new conditions through a number of feedback processes.

Without more detailed study, however, particularly of catchment conditions prior to the 1967 flood, it would be difficult to judge the significance of the estimated 10,000-year+ return-period of the event. In the neighbouring Whitendale catchment the morphological legacy of the flood is minimal (flood peak approx. 7 times the mean annual flood). The evidence of incision in the Brennand headwaters and of the "wandering" channel which now occupies the transfer-zone is that lesser events would not cross the threshold. Indeed, in the neighbouring Langden Brook catchment Thompson (1987) concluded that channel planform fluctuations could be accommodated within a dynamic equilibrium framework despite the repeated crossing of (intransitive) channel thresholds. Valley-floor widths are important to this discussion - in the "Charley" flood areas it was the alternation of severe confinement and storage zones which set up the threshold conditions; in Bowland the wider valley floors are influenced only by a much higher magnitude event.

It is interesting to note in this respect that, in writing of a number of similar upland British catchments, Harvey, Hitchcock and Hughes (1982) see no evidence of recent approach to a major system threshold. They identify process thresholds and quote time intervals for events controlling overall channel morphology (2 - 4 per year) and adjustment of channel forms (14 - 30 per year). They observe that "there is no evidence that any past events larger than those occurring during the study period have had any lasting influence on the contemporary morphology of the active landforms" (p165).

However, in one of their study areas a 100yr + storm in June 1982 crossed major slope stability thresholds as well as those for a range of processes of fluvial transport (especially debris flows) to produce profound landscape and, on the resulting fan features a potential instability during subsequent fluvial action (Harvey, 1987; Wells and Harvey, 1987). There was on this occasion a "semi-permanent" change to braided channels sustained by a large number of new sediment sources. Channel slopes were also influenced by the calibre of the new inputs and presumably, therefore, an almost entirely new set of process-response relationships now exists.

At the slightly longer timescale there is considerable overlap between the catchments on which "rare, great" floods have been documented in Britain (Acreman, 1989) and the regions in which channel instability is pronounced in downstream, piedmont reaches (Hooke and Redmond, 1989)

One further conclusion by Newson and Bathurst is relevant; that is the contribution of land management practices to restricting biogenic recovery mechanisms in the Forest of Bowland catchments. The use of all-terrain vehicles, bracken control, heather burning and failure to control an infestation of burrowing animals (rabbits) has helped keep the process system in its current state of relative instability.

From these field studies we may conclude that, from the standpoint of long-lasting process switches induced by flood events there are a number of regularities:

a. If the extrinsic signal is a "rare, great" flood initiating slope changes (note the channel/slope flood demarcation of Newson, 1980), there are likely to be transitive threshold changes, during recovery

b. Sediment storage in headwater valley floors permits the development of intrinsic thresholds

c. When they are crossed the processes of incision represents the transitive threshold change, involving positive feedbacks to sediment supply and stream power during lesser events

d. Biogenic and land-use influences are important if they influence recovery by colonisation of deposits, maintain sediment supply sources or impinge directly on valley floor sediment stores.

5. Threshold behaviour and the spatial and temporal patterns of sediment throughput in river basin systems

Lewin (1989) has described the background to the domination of the "fluvial processes" school in the 1960's in terms of the availability of a spatially-restricted fluvial database, i.e. one derived largely from gauged alluvial channels in large basins. "Belief in systems that equilibriated in the short term allowed students to look at functional inter-relationships between discharge and other environmental variables such as slope, soil and drainage network parameters" (p 266). The origins and emphasis of data widely used in the subject are also considered by Lewin as of significance in forming what he describes as the "unorthodox challenges" to the process school. "There appears to be a growing impression that some environments are more prone to long-lasting landforming effects of exceptional events than others" (p 273): Lewin quotes semi-arid environments and small headwater catchments as susceptible; in both of these environments sediment supply processes are significantly altered by "rare great" floods and the growth curve of flood frequency is steep.

It is important to remember that Schumm's treatment of threshold phenomena in "The Fluvial System" (1977) was also spatially limited; he restricts thresholds to those parts of the system dominated by highly stratified climate, proclivity to large flood events, steep relief and neotectonics. These are the regions of high sediment production and storage, i.e. "The drainage basin", "valleys and valley fills" and "the piedmont" (largely alluvial fans).

Warner (1987) has confirmed a predictable spatial variability in channel response to the pronounced climatic variability of New South Wales. "Gorge", "armoured" and "backwater" zones react little and slowly to the alternation of flood- and drought-dominated regimes; "transition" and "mobile" zones react profoundly and rapidly (zones after Pickup, 1984). Because the mobile zone is composed of sand-bed rivers we restrict ourselves here largely to the transition zone, where-ever it may occur (as with many zonal schemes the individual units of this one may repeat throughout the fluvial system; this is a point further confirmed by Maizels (1985) for a major Scottish river.

It may be preferable to remain less specific about labelling zones of the river long-profile in relation to their susceptibility to threshold response and instead to refer to those parts of any drainage basin where considerable sediment storage occurs either on slopes or in channels, the result of either current, or as Church, Kellerhals and Day (1989) have shown, long-past, processes. Clearly, therefore, we can include both the apparently different valley fills of the narrow British upland headwaters and Warner's gorge zone - see Nanson's (1986) "floodplain stripping" mechanisms.

It may well be that the proper definition of geomorphic thresholds needs to reconcile the spatial ditribution, volume and erodibility of basin storages with the magnitude/frequency concept. Not only is the perturbing event of importance here, but also the recovery process. Pitlick (1988) has a "zone-free" classification of the recovery process from the impact of large floods; the reach factors which control recovery (if it is to occur) include:

    magnitude of the initial perturbation
    sediment storage potential of the reach
    continuity of sediment supply from upstream
    frequency of flows subsequent to the big event

The storages, too, need a formal classification. Kelsey, Lamberson and Madej (1986) describe the movement of a "slug" of sediments introduced by a flood to the 720 km Redwood Creek basin; it traversed the active storages of the basin in 18 years but the important contribution made by this study is that of a model which simulates the role of active, semi-active, inactive and stable sediment stores in the basin. Not all valley-floor storages will respond in the classsic threshold fashion. Macklin and Lewin (1989) describe the 115-year fluctuations of five "sedimentation zones" in a 22 km reach of the South Tyne. These zones, which average 1.69 km long appear to be largely a glacial inheritance; the more stable zones which link them are controlled by bedrock or by confining Quaternary fills and have much steeper channel gradient. Key controls are operated by the ability of the intervening "stable" (i.e. laterally stable - Lewin & Macklin's study is primarily from maps) reaches to temporarily store coarse sediments and by the ability of major floods to transfer such sediments (in-channel) to the downstrream sedimentation zone. Once again the relationship between storage location, volume and flood frequency is confirmed.

Obviously the emerging hypothesis that vulnerability to transitive threshold behaviour during the extreme events which may typify climate change can be localised and predicted in drainage basin systems can be partially tested by reference to past climate changes, such as those of the

Holocene. From this perspective Rose and Boardman (1983), comparing the sedimentological response of two contrasting rivers of the same order to the same climatic deterioration (the lowland, alluvial, Gipping and the upland, boulder-bed, Mosedale Beck), conclude that gradient-controlled valley floor energy conditions explain why upland basins exhibit continuing complex response following a perturbation. Nevertheless, in both river systems there are alternating phases of incision and aggradation at first because of the delay between climate change and the sediment supply response. Richards et al (1987) review the major phases of anthropogenic sediment-supply changes in the British environment; they too contrast the sediment size and supply cycles of the lowland and upland headwater sites which have been reported and dated. The upland catchments are notable for the additional influence of localized meteorological events and for their episodic use and abandonment by agriculture.

Green and McGregor (1987) describe the significance of the thresholds implicit in dynamic metastable equilibrium for the formation of Holocene valley floors. Thus "relatively shortlived conditions of environmental change" can produce a terrace whose detail is then added by longer spells of quiescence and "morphostasis" Thanks to the temporal and spatial variability of threshold conditions, they claim that geomorphological response, even to global change is unlikely to be synchronous.

6. Conclusions and implications for managed fluvial environments.

Acceptance or rejection of the threshold concept by those disciplines and environments represented at this Workshop may either confirm the paradigmatic nature of thresholds or support the contention that thresholds are poorly defined. However, we might also consider that their proper investigation refers us to an new emphasis on sediment storage in river basins.

Geomorphologists are beginning to lay down guidance on the techniques for properly locating and evaluating basin storage, for example Laronne and Duncan (1989) have measured the storage potential and duration for braided river sections; channel sediment delivery ratios (Phillips, 1989) are also a conceptual advance and can be modelled. One of the most influential recent workshops on the topic of fluvial sediment storage concluded (Kelsey, 1982) that there was considerable potential in the division of drainage basins on the basis of supply- and transport-limited sediment fluxes and the appropriate magnitude/frequency growth curve. At the same meeting Reid formalised an

approach to sediment routing through drainage basins on a "from", "to" and "process" matrix. Swanson, Janda and Dunne (1982) concluded that "geomorphic recovery from a sediment-routing standpoint could be viewed as the refilling of storage sites and their readiness to fail again".

Knox (1989) suggests that the tendency for net storage in modern drainage systems implies a degree of disequilibrium in the erosional and depositional system. Furthermore, "the effectiveness to erode, transport and store sediment in watersheds is subject to considerable spatial variation within drainage hierarchies and the zone of maxium effectiveness may shift with changing processes over time" (p158). It required a complete change of climatic conditions to remove the colluvial deposits of the Late Pleistocene in the Upper Mississippi; this was then followed by a disjointed sequence of anthropogenic sedimentation from agriculture and removal of these storages once conservation treatments were applied to the headwaters. Humid environments can exhibit significant volumes of sediment storage within small basins, for example 90% of the material excavated by modern gullying is stored within the 13.8km Fernances Creek basin in south-east Australia (Melville and Erskine, 1986).

In terms of management of fluvial systems it is significant to note that certain characteristics of human manipulation of both land and water resource systems set up transitive or intransitive geomorphic thresholds. For instance, in the release of large volumes of Quaternary sediments by ploughing steep land for afforestation in Britain, or by closing off an impoundment to remove sediment from the downstream reaches we produce a hiatus which may or not recover (see Petts, 1979, 1987). Urbanisation can produce a discontinuity between the release of large amounts of sediment (during building) and the release of increased flood flows (after completion) - see Roberts, 1989.

However, management should now be concentrating on the sensitive zones of each major basin and the current condition of the anthropogenic/ biogenic controls on sediment stores in relation to the effect of climate change in either filling or emptying those stores. Brown (1987) writes of the importance of storage zones in middle and lower basins and suggests that up to 3,000 years' volume of stored sediments may be removed from the Severn basin (UK) were it not for extensive flood- and erosion-protection works on the river. Burrin (1985), writing about rivers in Sussex (UK) concludes that their valley floors have been artificially separated from formative fluvial action for up to 500 years as a result of engineering. If the response to climate change is further structural protection we further delay the time during which these sediments will be remobilised. Because of the popularity of sustainable

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design and accounting timescales nowadays these are
considerations which are politically and, through
geomorphology, practically relevant.

Clearly the threshold concept has already had some practical
utility; even the now discredited threshold of channel
planform has encouraged a more sympathetic engineering
treatment of channels in the field (Newson, 1986; Leeks,
Lewin and Newson, 1988). However geomorphologists are now
contemplating the benefits of thirty years of process
measurements designed to elucidate the controlling variables
in river dynamics. Richards (1990) has concluded that we
should now judge results on the basis of their ability to
explain the development of landforms. "Real" geomorphology,
writes Richards, is emerging in which "criteria for
acceptance of a theory are not based on predictive success
but on explanatory power".
The critical phase now entered "must involve the emergence
of new structures and relationships employing the identified
mechanisms". Richards advocates the most likely line of
progress as "the development of iteratively and spatially-
distributed form-process-form feedback models".

A properly-defined, spatially-discrete threshold approach to
the fluvial landforms created by sediment storage in river
basins would conform to Richards' wishes. "Threshold
spotting" now has to enter a new era and the signals for its
methodlogies have come from those working with both
historical and contemporary storages in a variety of
climatic zones.

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ANALYSIS OF MASSIVE SEDIMENT TRANSPORT PROCESSES IN TORRENTS

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Abstract

As a result of a laboratory study, new formulae are proposed to calculate bed load transport rates and velocities of torrent flows with high suspended fine material concentrations. Using the bedload transport equation in a case study, sediment volumes of flood events are determined and compared with independent estimates. In a second part, several empirical equations are compared which were developed to estimate the total sediment volume that may be mobilized in a torrent catchment during an extreme rainstorm event by either a debris flow or by fluvial sediment transport.

1 INTRODUCTION

With increasing population densities and with increased tourist activities in mountain areas, the problem of sediment hazards from mountain torrents has to be given more attention. A flood event involving massive sediment transport can endanger human life, property and traffic routes particularly on the fan area. If a lot of material is deposited at the mouth of the torrent, the main stream in the valley may be dammed which will cause trouble for the next village downstream. In order to plan countermeasures against such disasters, it is therefore important to know more about massive sediment transport processes in torrents and to develop methods to estimate sediment volumes resulting from flood events.

Torrent catchments are characterized by steep gradients, and thus the sediment transport capacity in the channels can be very large. During periods with no major flood events, however, a lot of debris and sediment may be stored in and along the channel. In addition, there are often steep side slopes composed of loose material (e.g. moraines). If an intense rainstorm occurs over such a catchment, the resulting flood may mobilize major parts of these sediment resources, and under certain conditions the transport can also be in form of a debris flow. Accompanying landslides enhance both the sediment supply into the torrent and the probability that a dangerous debris flow will form.

2 FLUVIAL SEDIMENT TRANSPORT

Bed load transport experiments

During a flood event a lot of fine material may enter the channel from the side slopes. If this material is suspended in large quantities by the water flow, a slurry of fine material will result which can eventually form the fluid matrix of a debris flow. In order to examine the transition region between "normal" (fluvial) sediment transport and debris flows, an experimental study was performed (Rickenmann, 1990). A clay suspension of increasing concentrations was used to simulate the fine material slurry. In a steep flume with slopes between 5 and 20 %, the bed load transport capacity (q_B) and the flow behaviour of the slurry were determined, looking at the effect of the increasing fluid density and viscosity with increasing clay concentration of the flow.

The main results of this study can be summarised as follows: As long as viscous effects remained negligible (particle Reynolds numbers above approximately 10) the bed load transport rates increased with increasing fluid density. As compared to the corresponding clear water flow (having the same flow rate and the same slope), the increase in the experiments reached up to 300 %. Once the flow around the grains became laminar (particle Reynolds numbers below 10), however, a marked decrease of q_B with further increasing clay concentration could be observed. All the experiments were performed with steady uniform flow. The fact that the transport capacity of debris flows is still much greater than in these slurry flows, is probably associated with the unsteady flow behaviour and the different rheologic behaviour of such grain-water mixtures.

The majority of the experiments were performed under flow conditions with negligible viscous effects. The analysis showed that this experimental data can be described by a similar bed load transport formula as proposed by Smart and Jäggi (1983), only the density factor had to be adjusted. Since these authors had performed similar experiments in the same steep flume facilities, using clear water as transporting fluid, both data sets were analysed together (totalling 115 tests). The resulting calculation procedure is shown in Fig. 1.

At steep slopes (>~ 10%) the bed load transport capacity becomes so large that the concentration of moving grains increases the flow depth. Therefore a distinction is necessary between the mixture flow depth h_m and the fictitious fluid depth h_f (fictitious depth in the case of no transport). For the same reason two different calculation procedures were proposed, depending on whether the flow rate (q) or the flow depth (h_m) is given or assumed. In the following case studies we start from the flow rate, therefore only the first procedure (labelled "design case") is presented here. The second calculation procedure can be found in Rickenmann (1990).

As mentionned, equ. (2) is the result of a regression analysis of both the clay suspension experiments of the author and the tests of Smart and Jäggi (1983). Equ. (1) represents a slight modification of a formula proposed by Bathurst et al. (1987), to include also density effects. This equation is partly based on steep flume experiments (up to 20% slope) and it applies essentially to beds with more or less uniform grain size distribution. From the flow resistance analysis of the steep flume tests, a formula very similar to equ. (3) had been developed. Interestingly, the measured velocities fitted even better to equ. (3) which was adapted from Takahashi (1978); his formula is based on theoretical considerations, and a "calibration" with his experimental debris flows results in the constant 1.3. Equ. (4) is a purely empirical relation that was developed from the steep flume data of the author and of Smart and Jäggi (1983).

The channel bed of a torrent is often composed of an armour layer containing very coarse pebbles and boulders. The main difficulty in assessing the sediment transport is to determine the critical discharge above which substantial transport will occur. For rivers with flat slopes, full sediment transport (close to the calculated transport capacity) may be assumed after the armour layer is destroyed. In a torrent, however, the coarsest boulders (grains sizes above about 1 m) may not be moved even at higher flow rates. Since transport formulae

were developed with comparatively uniform grain sizes, effective sediment transport rates in torrents might only make up for a part of the calculated values.
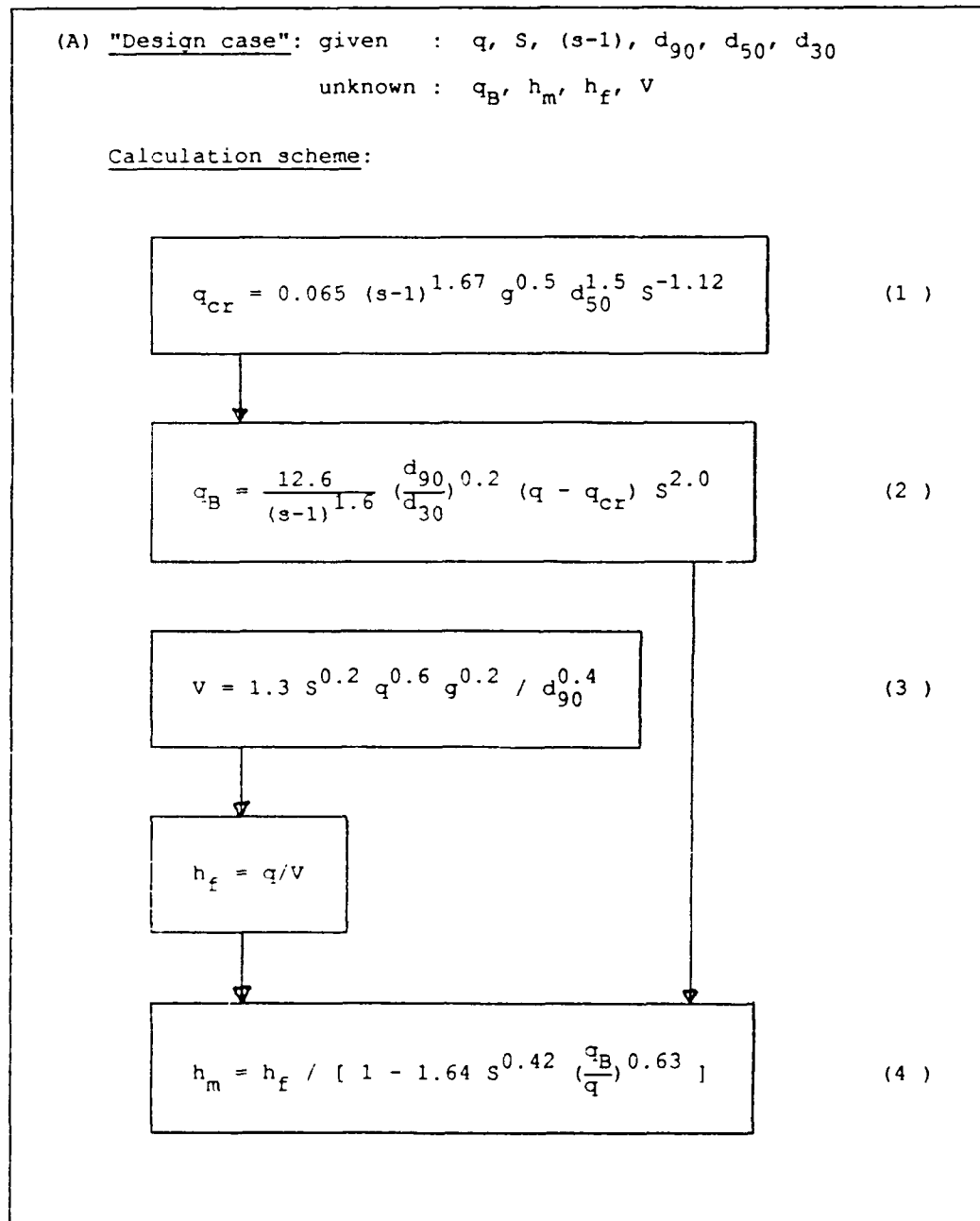
(A) <u>"Design case"</u>: given : q, S, $(s-1)$, d_{90}, d_{50}, d_{30}

unknown : q_B, h_m, h_f, V

<u>Calculation scheme</u>:

$$q_{cr} = 0.065 \ (s-1)^{1.67} \ g^{0.5} \ d_{50}^{1.5} \ S^{-1.12} \qquad (1)$$

$$q_B = \frac{12.6}{(s-1)^{1.6}} \ (\frac{d_{90}}{d_{30}})^{0.2} \ (q - q_{cr}) \ S^{2.0} \qquad (2)$$

$$V = 1.3 \ S^{0.2} \ q^{0.6} \ g^{0.2} \ / \ d_{90}^{0.4} \qquad (3)$$

$$h_f = q/V$$

$$h_m = h_f \ / \ [\ 1 - 1.64 \ S^{0.42} \ (\frac{q_B}{q})^{0.63} \ ] \qquad (4)$$

Fig. 1: Calculation procedure for the "design case". Recommended range of application: 5 % <= S <= 20 %, $q >= 5 \cdot q_{cr}$, h_m/d_{90} <= 20.

In Fig. 1, the following notation is used:

q : fluid flow rate per unit width [m³/(s·m)]
S : slope
s = ρ_s/ρ_f: ratio of sediment to fluid density
d_{90} : characteristic grain size than which 90 weight percent is finer
d_{50} : characteristic grain size than which 50 weight percent is finer
d_{30} : characteristic grain size than which 30 weight percent is finer
q_B: bed load transport rate per unit width [m³/(s·m)]
q_{cr}: critical flow rate at initiation of motion
h_m: mixture flow depth (including part occupied by moving grains)
h_f : (fictitious) fluid flow depth (= q/V)
v: mean fluid velocity

Günter (1971) performed experiments to study the effect of a wide grain size distribution on the critical dimensionless shear stress to break up the armour layer, $\theta_{cr,D}$, which can be expressed as follows:

$$\theta_{cr,D} = \theta_{cr} \left(\frac{d_{90}}{d_m}\right)^{0.67} \tag{5}$$

where θ_{cr} is the critical dimensionless shear stress at initiation of motion for a uniform sediment (θ_{cr} = 0.05), and d_m is the mean grain size. If equ. (5) is combined with the formula of Manning-Strickler, a critical discharge for the destruction of the armour layer, $q_{cr,D}$, can be determined. Comparing this relation with equ. (1, the following first approximation results (putting d_{50} = d_m):

$$q_{cr,D} = \frac{d_{90}}{d_m} q_{cr} \tag{6}$$

For the bed material of the examples in the following case studies the ratio (d_{90}/d_m) varies between 2 and 2.6. In this case it may be concluded that the critical discharge to destroy the armour layer is about 2 to 3 times larger than the one for initiation of motion with a relatively uniform sediment (with the same mean grain size d_m).

In the bed load transport equ. (2), the effect of a wide grain size distribution is accounted for by the factor (d_{90}/d_{30}), and the density effect by the factor (s-1). These influences may be replaced by a single factor F. Putting F = 1 for clear water with (s-1) = 1.65 and for a uniform sediment with $(d_{90}/d_{30})^{0.2}$= 1.05 (Smart and Jäggi, 1983), equ. (2) can be simplified to:

$$q_B = 5.9 F (q - q_{cr}) S^{2.0} \tag{7}$$

The respective influence of these two effects is as follows (if both effects are to be taken into account, the two factors have to be multiplied) :

$$d_{90}/d_{30} = 5 \quad ----> \quad \text{increase of } q_B \text{ by factor } F = 1.3$$
$$10 \qquad\qquad\qquad\qquad\qquad\qquad\qquad 1.5$$

$$\rho_f = 1.15 \text{ t/m}^3 \quad ----> \quad \text{increase of } q_B \text{ by factor } F = 1.5$$
$$1.28 \qquad\qquad\qquad\qquad\qquad\qquad\qquad 2.0$$

Case Study

During the summer of 1987 two heavy rainstorm events occurred in July and August, resulting in large floods and widespread debris flow activity. These natural events killed one car driver and caused a lot of damage to traffic routes, buildings and cultivated land. In order to study the causes and the processes associated with these events, a national research program was initiated by the Swiss Federal Board of Water Resources Management. In one subproject, the debris flow and flood events in torrent catchments were studied (Haeberli et al., 1990).

Four flood events in torrent catchments were studied in more detail. The sediment volumes transported to the mouth of the catchment were between 30'000 m³ and 250'000 m³. In all these torrent catchments there was both a flood event and debris flow activity during long-lasting precipitation of about 1 1/2 to 3 days. Based on field observation and eyewitness accounts, an attempt was made to distinguish between the sediment volumes resulting from fluvial transport and those volumes associated with the debris flows.

The following Table 1 gives the main characteristics of the catchment and the transported sediment volumes:

name	date	catchment area	mean creek gradient	fluvial sed. volume [m³]	debris flow sed. volume [m³]
Varuna	18.7.87	6.6 km²	37 %	max. 20'000[1]	200'000[2]
Plaunca	18.7.87	3.8 km²	34 %	?	240'000[2]
Zavragia	18.7.87	13.0 km²	26 %	?	30'000[3]
Münster	24.8.87	15.3 km²	16 %	max. 20'000[4]	30'000[4]

Table 1: Observed sediment volumes resulting from fluvial transport or debris flows, for four flood events in torrent catchments.

In Table 1, the sediment volume estimates are based on:

1) sediment retention capacity of channels in the fan area, with no deposition outside channels according to eyewitnesses

2) analysis of aerial photographs before and after the event

3) damming of the main stream within about 15 to 45 minutes

4) debris flow occurred clearly before flood event; photographic documentation of deposition in the village area before and after the flood

In all four creeks, information on the characteristic grain sizes of the bed material was obtained from several frequency-by-number transect samples of the armour layer. Subsequent transformation into a sieve-by-weight distribution of the whole grain mixture was performed according to a procedure given by Fehr (1987). The following values represent the bed material characteristics near the head of the fan; also listed are the bed slope of the fan head and the mean width B of the channel there:

name	d_{90} [m]	d_m [m]	d_{30} [m]	d_{90}/d_{30}	slope of fan head	B [m]
Varuna	0.46	0.18	0.05	9.2	0.24	5
Plaunca	0.29	0.14	0.025	11.6	0.11	4
Zavragia	0.52	0.21	0.04	13.0	0.083	10
Münster	0.42	0.16	0.04	10.5	0.13	7

Table 2: Grain size and channel characteristics.

Since no discharge measurements (during the flood events) exist for any of the four creeks, flood hydrographs were constructed from rainfall intensity data for three hour intervalls (based or interpolation from nearby raingauge stations). Runoff coefficients were varied between 0.25 and 0.7, depending on catchment characteristics and preceeding rainfall conditions. The combination of the flood hydrograph with the bed load transport equation (7) yielded fluvial sediment volumes.

Preliminary calculations showed that these volumes depend to a great extent on both the value of the critical discharge for initiation of motion and the time during which substantial sediment transport is

assumed to have occurred. A breaking up of the armour layer can be expected with increasing discharge, but what is the effect of the largest boulders not moved by the flow ? And the armour layer will probably be rebuilt during the receding part of the hydrograph. Calculations of the flow depth at peak discharge according to the procedure in Fig. 1 resulted in values less than 1 m at the time of the flood peak. In view of the uncertaincies related to the beginning and ceasing of transport, it was decided to neglect any correction for sidewall friction effects, and a rectangular flow cross section was assumed. Thus, discharge values over the whole width were obtained as: $Q = B \cdot q$, $Q_{cr} = B \cdot q_{cr}$, $Q_B = B \cdot q_B$.

In a first case (A), the critical discharge q_{cr} (and Q_{cr}) was determined with d_m. A comparison of the resulting sediment volumes with the observed ones (Varuna and Münster creek) showed a clear overestimation (although with d_m a larger grain size was already used than with d_{50} according to equ. 1). A considerable higher critical discharge for the breaking up of the armour layer is obtained by equ. (6), the increase is a factor d_{90}/d_m. For a second set of calculations (case B), it was decided to choose an even larger upper extreme value: Using d_{90} instead of d_m in equ. (1), there is an increase in q_{cr} by a factor $(d_{90}/d_m)^{1.5}$.

The combined effects of an increased fluid density and of a wide grain size distribution are taken into account by the factor F. For the Varuna, Zavragia and Münster creek this factor was varied between 1.0 and 1.5. Extremely high fine material concentrations were indicated for the flood event in the Plaunca creek, muddy and humid deposits could be observed even some days after the event. In this case the factor F was varied between 1.5 and 2.0. Several combinations between the cases A or B and different values of F were introduced in the bed load transport calculations. Case 1 refers to the greater, case 2 to the smaller F value. Here, only the two extreme combinations are considered. The results are given in Table 3 and in the Figures 2(a)-(d) showing the constructed hydrographs and the calculated sediment graphs.

A further assumption concerns the onset of substantial sediment transport during the event. For three creeks (Varuna, Zavragia, and Münster) this time is mainly based on eyewitness accounts. The flood event in the Plaunca creek was not observed by anybody. There it was assumed that substantial transport began either at the flood peak (linear increase from 14^{00}h to 15^{30}h) or after the presumed time of debris flow occurrence (21^{30}h). For the Varuna creek, no substantial bed load transport was reported for the time before 17^{00}h. In the Zavragia
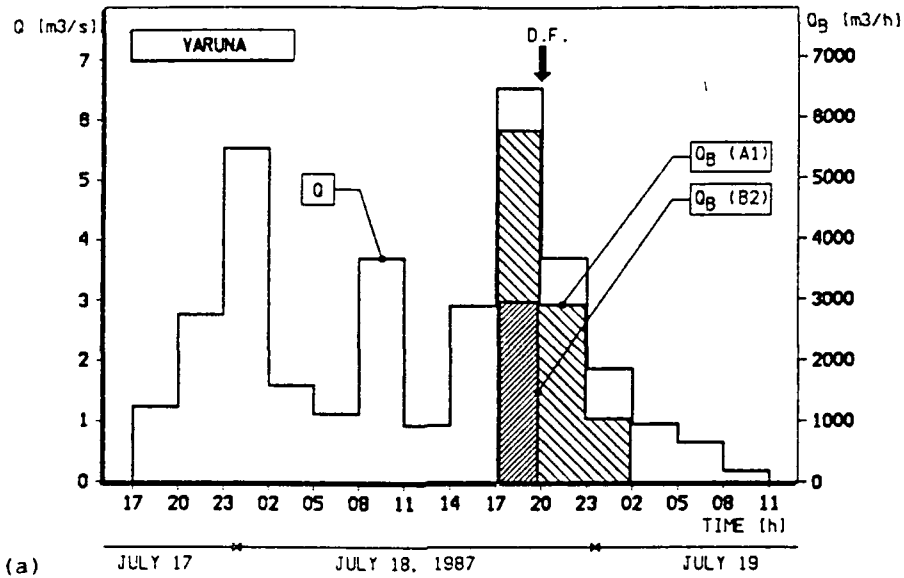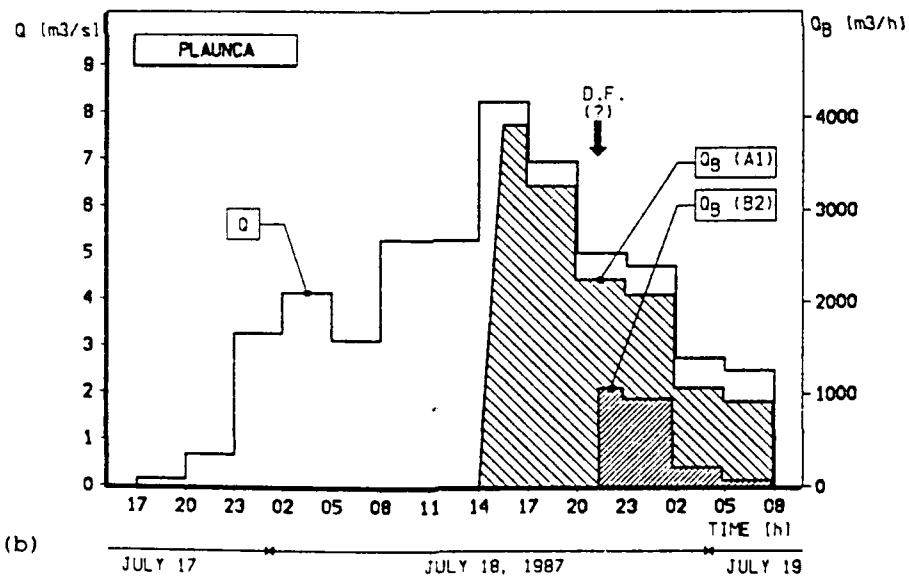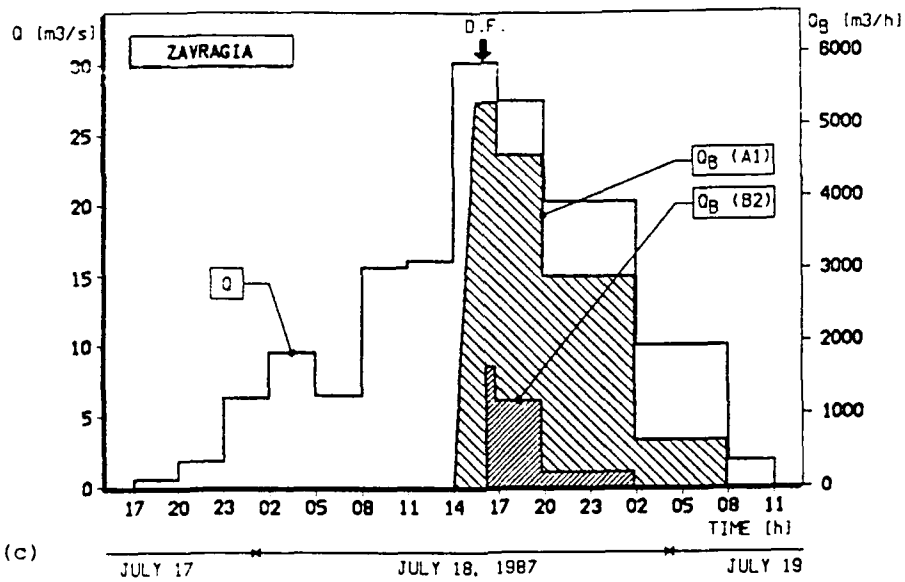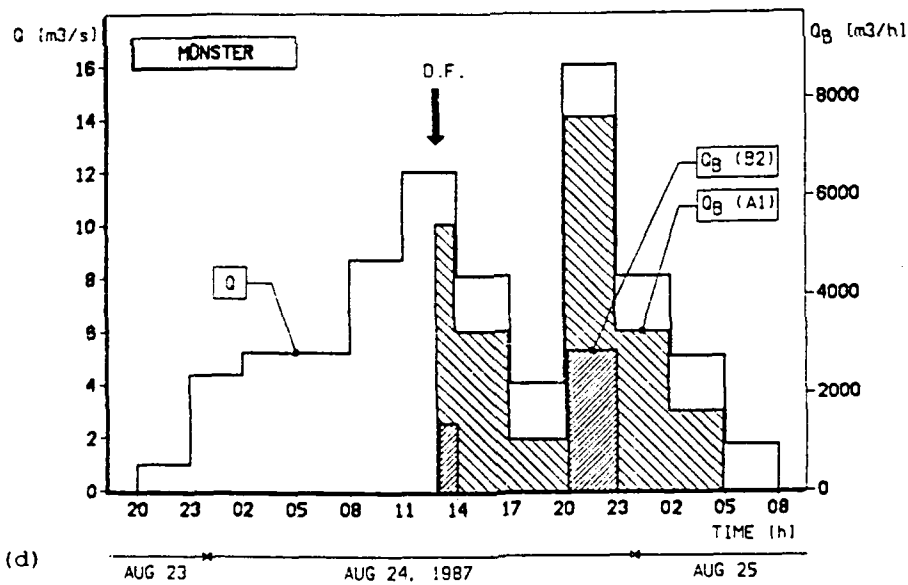
Fig. 2(a)

Fig. 2(b)

Fig. 2(c)

Fig. 2(d)

Fig. 2: Flood hydrographs constructed from rainfall data and sediment
graphs calculated by equ. (7), for four torrent catchments: (a)
Varuna, (b) Plaunca, (c) Zavragia, and (d) Münster. The arrow
with D.F. marks the time of the (first) debris flow occurrence.

case, bed load transport is said to have started in the afternoon. With Münster, the debris flow (13^{00}h) occurred clearly before the flood peak. It may be assumed that the armour layer was destroyed afterwards, and that some bed load transport occurred in the afternoon, although the eyewitnesses reported substantial transport during the evening (of August 24) and overnight.

name	case	d [m]	Q_{cr} [m³/s]	F	Fluvial sed. volume [m³]
Varuna	A1	0.18	0.9	1.5	45'000
	B2	0.46	3.6	1.0	11'000
Plaunca	A1	0.14	0.8	2.0	37'000
	B2	0.29	2.3	1.5	5'000
Zavragia	A1	0.21	7.4	1.5	48'000
	B2	0.40*	19.3	1.0	6'000
Münster	A1	0.16	2.1	1.5	57'000
	B2	0.40	8.2	1.0	10'000

* With d_{90} = 0.52 m the critical discharge becomes larger than the peak flow rate, and no bed load transport would result. Therefore a smaller characteristic grain size is assumed in case B.

Table 3: Range of calculated fluvial sediment volumes.

The figures for the Varuna and Münster creeks can be compared with the observed sediment volumes given in Table 1. It is seen that the predicted volumes based on d_{90} (to determine Q_{cr}, case B) are closer to the observations (s. Table 1) than those based on d_m (case A). This is in general agreement with equ. (6) which indicates that a larger critical discharge is required to break up the armour layer than the one for initiation of motion of a relatively uniform sediment (equ. 1).

A torrent bed is characterised by a very wide grain size distribution. A complete destruction of the armour layer is probably very rare and might not even be reached after the passage of a debris flow. During a flood event, sediment enters the channel also from undercutting and failure of side slopes; however, it cannot be expected that this supply

is sufficient to satisfy the (calculated) transport capacity of the flow. It may therefore be concluded that direct application of bed load transport formulae, developed under idealized laboratory conditions, tend to overestimate effective transport rates in a torrent.

3 DEBRIS FLOW OR TOTAL SEDIMENT VOLUMES

During flood events, enormous masses of sediment can be delivered to the mouth of a torrent catchment. In many cases, only minor parts of all the material is transported in fluvial form (as bed and suspended load) and large portions are moved in the form of a debris flow which may consist of one or several waves or pulses. Then it is no longer sufficient to estimate the sediment volumes transported from the catchment by using a bedload transport formula and a hydrograph (as done in the previous section). Instantaneous peak discharges of debris flow pulses can be one or two orders of magnitude higher than the peak water discharge during the same flood event (Haeberli et al., 1990). In such cases, the approach discussed above could clearly underestimate resulting sediment volumes.

Therefore numerous debris flow events of the summer 1987 in Switzerland were analysed on a statistical basis. As a result, a new empirical formula was developed to estimate the debris potential associated with such events. This relation is compared with similar equations which were developed in Austria and in Japan.

Analysis of Swiss debris flow events

The two major rainstorm events (July 18/19 and August 24/25) of the summer 1987 caused flooding (s. also case study of previous section) and debris flows in several regions in the Swiss alps. A large concentration of debris flows could be observed in the scree and moraine slopes of the periglacial belt, because warm temperatures resulted in rain- instead of snowfall at high altitudes. Most of the debris flows analysed in this study deposited between 1000 m^3 and several 10'000 m^3 of material on the fan. However, there was also a number of events in which up to several 100'000 m^3 of sediment were eroded in and along torrent channels and transported down to the main valley (s. also Table 1).

A detailed analysis was performed with 82 of these debris flow events (Haeberli et al., 1990). It was examined whether any general relationships exist between simple parameters such as total debris volume, average debris yield rate per unit channel length, catchment area, mean channel slope and fan slope. The most clear dependency was found on a diagram of channel debris yield rate versus fan slope. From this diagram, an upper limit defining the maximum debris yield rate, EL [m³/m], could be determined as a function of the fan slope in percent, J_K% :

$$EL = 110 - 2.5\ J_K\% \qquad\qquad (8\ )$$

For the examined cases, the fan slope varied between 8 and 45%, and the largest value of the average debris yield rate was about 100 m³/m. (It may be noted that negative values of EL result, if J_K is greater than 45%). Since the debris yield rate EL is defined as the ratio of the total debris volume, M [m³], to the channel length, L [m], equ. (8) can be transormed into:

$$M = (110 - 2.5\ J_K\%)\cdot L \qquad\qquad (9\ )$$

These equations express an upper limit of erosion to be expected in the case of a debris flow event. To illustrate the degree of overestimation, the ratios of calculated to observed debris volumes (M_{cal}/M_{obs}, termed "estimation error") are plotted in Fig. 3 as a function of the observed volumes. Also included are some Austrian events (taken from Hampel, 1977, 1980) to enhance the number of debris flows with large volumes. It is seen from Fig. 3 that very large estimation errors can result for events with smaller observed volumes, while for large events (with M_{obs} >= 100'000 m³) the error is less than a factor of 2.

Comparison with similar formulae

In Austria and in Japan, similar formulae were proposed to estimate the maximum debris potential that can be moved during extreme flood and debris flow events. Also with these equations, the maximum debris volume is estimated as a function of simple parameters characterising certain elements of the torrent catchment.
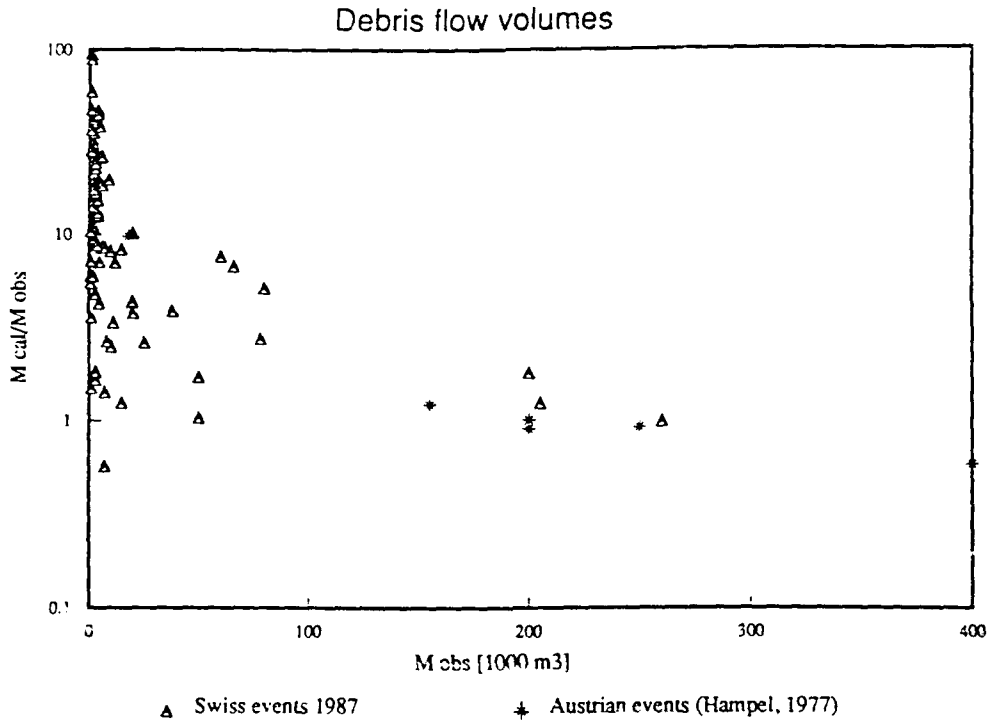
Debris flow volumes

Fig. 3: Performance of Equ. (9) in comparison to 82 Swiss debris flow
and some Austrian events, showing the ratio of calculated to
observed debris volumes (M_{cal}/M_{obs}) as a function of the (ob-
served) event magnitude (M_{obs}).

Hampel (1977) developed an equation to estimate the debris volume M
in [m³] as a function of the parameters catchment area, EG in [km²], and
fan slope J_K% :

$$M = 150\ EG\ (J_K\% - 3)^{2.3} \tag{10}$$

The exponent of the slope term is partly based on laboratory
experiments, and the constant 150 was "calibrated" with 15 debris flow
events with volumes between 4000 and 500'000 m³; the corresponding fan
slopes vary between 6 and 14%.

Another formula was proposed by Kronfellner-Krauss (1982), expressing
the debris volume, M in [m³], in terms of the catchment area EG [km²]
and the mean creek slope, J_m% in [%]:

$$M = K\ EG\ J_m\% \tag{11}$$

where K is a torrential erosion coefficient which varies between 500 (large catchments, with sediments already eroded) and 1500 (steep torrents with large sediment sources). In this study, a value K = 1500 was used. For the calibration of the coefficient K, Kronfellner-Krauss (1982, 1985 Tsukuba) used a part of over 100 events with volumes between 5000 and 20'000 m^3 and several events with more than 20'000 m^3. The corresponding catchment areas vary between about 1 and 80 km^2.

A third empirical equation is presented in Japanese planning and design guidelines against debris flow events (Public Works Research Institute, 1988). Using some coefficients given in this publication, the formula can be simplified to:

$$M = 1100 \ \Psi_o \ R_{24} \ EG \tag{12}$$

where R_{24} in [mm] is the accumulated rainfall 24 hours prior to the event, and Ψ_o is a runoff coefficient (given in a diagram), which decreases with increasing catchment area (Ψ_o ≈ 0.55 for EG = 0.1 km^2, and Ψ_o ≈ 0.1 for EG = 10 km^2); again, EG is in [km^2] and M in [m^3]. There is no mentioning of the range of the parameters used in the development of this equation.

For comparison with equ. (9) and with the corresponding data set, the three formulae (10) to (12) were applied to 82 Swiss debris flow events of the summer of 1987. The results are shown again in terms of the "estimation error" (M_{cal}/M_{obs}) as a function of the observed debris volume, s. Fig. 4(a)-(c). The largest ratio M_{cal}/M_{obs} shown in Fig. 4 is 100; even larger "estimation errors" result for small events (around 1000 m^3) according to equ. (10) of Hampel (up to a factor of 530).

It is seen that the three formulae (10) to (12) show a similar pattern in the diagrams as equ. (9) in Fig. 3. Again, there is a clear overestimation of the debris volumes for small events, while the error becomes smaller for larger events. It is noted that the three later formulae also underpredict observed volumes by up to a factor of about 5 (s. Fig. 4), whereas equ. (9) was developed on the basis of the presented data set to give an upper limit. The range of the estimation error is similar for all equations, except for the formula of Hampel (equ. 10) which yields a somewhat larger overestimation for some cases.
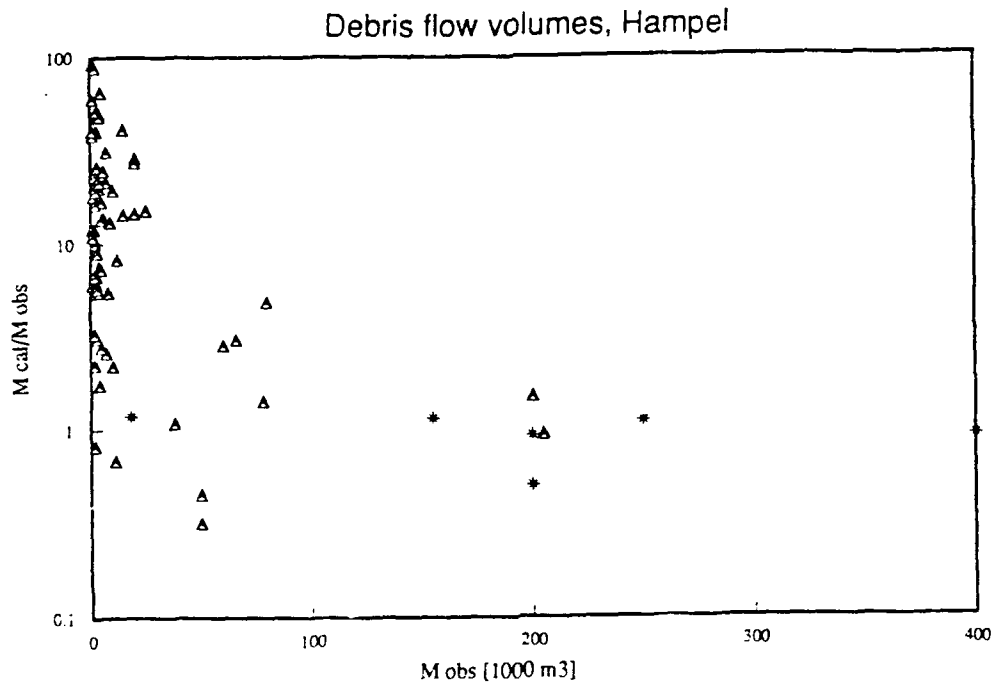
Debris flow volumes, Hampel

Fig. 4(a)   ▲ Swiss events 1987   ✦ Austrian events (Hampel, 1977)

Debris flow volumes, Kronfellner-Krauss
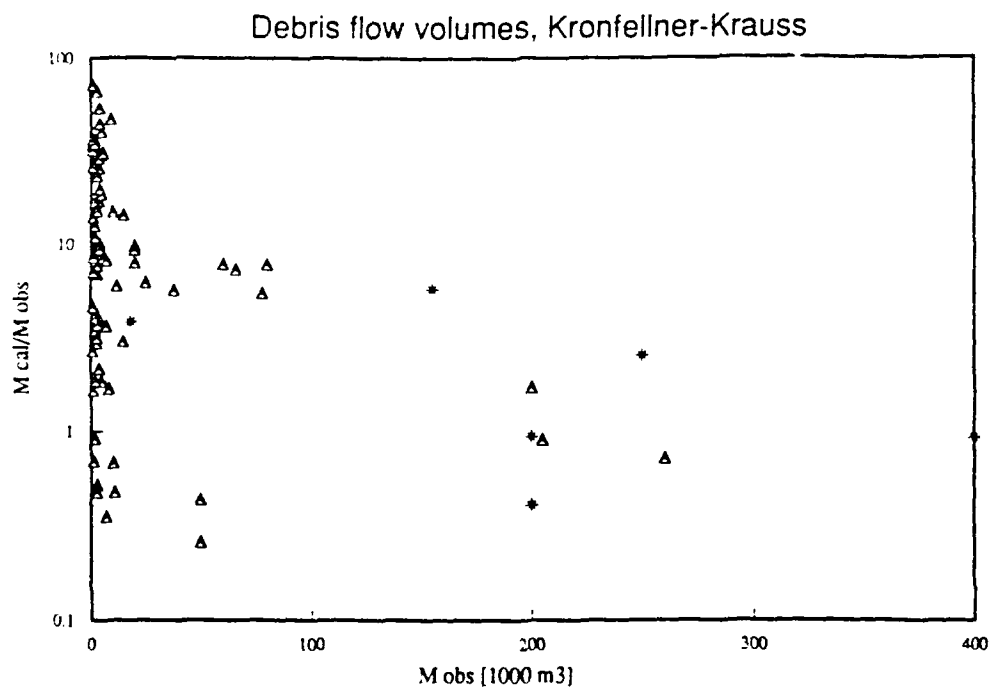
Fig. 4(b)   ▲ Swiss events 1987   ✦ Austrian events (Hampel, 1977)
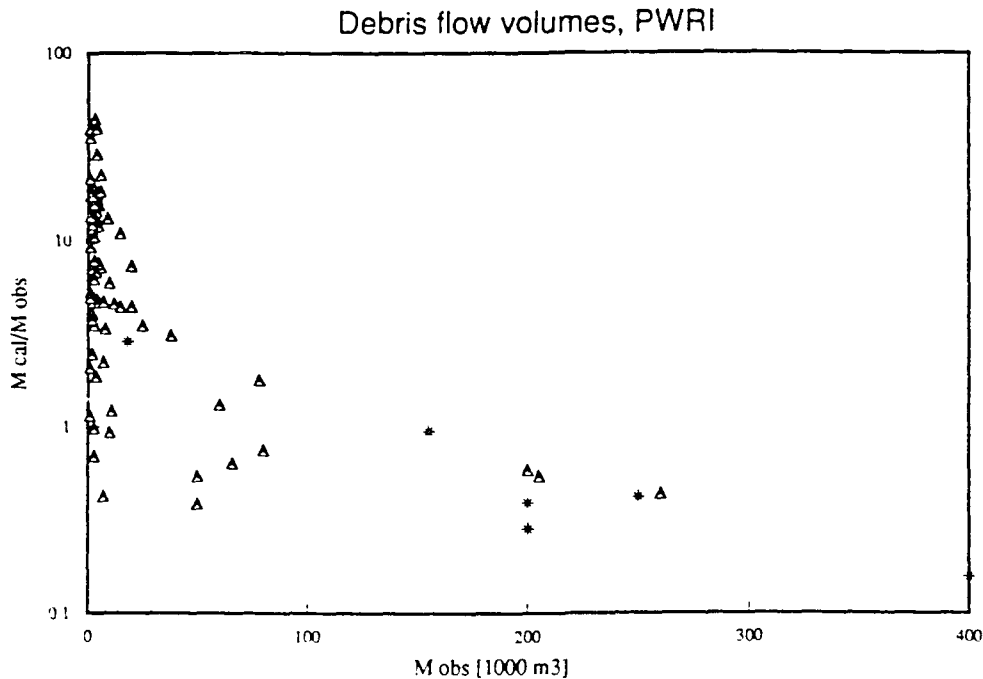
Debris flow volumes, PWRI

Fig. 4(c)

Fig. 4: Application of several empirical formulae to the same debris flow events as in Fig. 3, showing the ratio of calculated to observed debris volumes (M_{cal}/M_{obs}) as a function of the (observed) event magnitude (M_{obs}). M_{cal} was determined according to (a) Hampel's equ. (10), (b) Kronfellner-Krauss' equ. (11), and (c) the Japanese equ. (12).

For the considered debris flow events, the parameter R_{24} varied in a relatively narrow range (about 50 to 250 mm), so that it may be neglected in a first comparison of the four empirical formulae (in view of the much larger overall estimation error). Then it is not so much surprising that there is a similar estimation error for all four equations. For there are mainly two effects that are considered: First, a parameter characterising the catchment area (EG, L) represents the overall debris and water potential. Second, there is a "correction" factor (J_K, J_m, Ψ_o) which decreases with increasing catchment area. With equ. (9), the effect of an inreasing EG does not seem sufficiently accounted for by the parameter L (instead of EG); this is probably the reason why J_K has a negative sign, and thus the correction works in the opposite direction (as compared for example to Hampel's equ. 10).

It appears that the maximum debris potential can be very roughly estimated considering simple parameters that characterise the potential debris and water supply. This simplified view seems to hold approximately for larger catchment areas (>= 1 km²), involving still an estimation error up to a factor of the order of 10. With smaller catchments, other factor related to the initiation of debris flows become relatively much more important, and thus the error increases drastically. An upper limit of application may be given by the debris flow process as such, since with increasing catchment area and decreasing slopes, channel yield rates decrease markedly and furthermore a debris flow might not reach the mouth of the catchment. In alpine regions, a catchment area of about 25 km² may be suggested as this upper limit.

4 CONCLUSIONS

Based on laboratory experiments, a new procedure is proposed to determine bedload transport rates and flow resistance in steep channels. For four torrents, fluvial sediment volumes transported to the fan were calculated by combining the proposed bedload transport formula with an (estimated) flood hydrograph. A comparison with observed volumes shows that a relatively high critical discharge at initiation of motion should be selected, using rather d_{90} than d_m as characteristic grain size of the bed material. It seems likely that in a torrent the armour layer is very rarely completely detroyed. Thus full transport capacity (as determined from idealized experimental conditions) may be reached only during short periods of a flood event, since no continuous sediment supply can be expected from side slope failures.

To estimate the maximum debris potential of a torrent catchment, several empirical formulae are available. They are based on parameters characterising the water and sediment to be possibly discharged during a flood event. Such a simple approach gives a first, very rough estimate of the maximum debris volume that can be removed from the catchment during an extreme rainstorm event. An application to 82 debris flow events of the summer 1987 in Switzerland showed that the error of estimation decreases with increasing catchment area; observed debris volumes were overestimated by up to a factor 10 in catchments larger than about 1 km².

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Sediment Transport Processes on Alluvial Fans:
Influences on Fan Morphology and Development

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INTRODUCTION

Alluvial fans are important depositional landforms composed
predominantly of coarse sediments, laid down where stream systems
issue from steep mountain catchments into zones of reduced stream
power. They are transitional between dominantly erosional upland
or mountain catchments, within which the transporting capacity
exceeds the rate of sediment generation, and valley, piedmont,
or lowland areas where transporting capacity is much closer to,
and least during the deposition of the fan is or has been
exceeded by the rate of sediment supply. This situation has been
described by Bull (1979) as one where effective stream power
falls below critical stream power i.e. that required to transport
the available sediment (Figure 1). This typically occurs in
mountain-front locations but fans may also occur where steep
tributary streams join major valleys (Harvey 1989a). They are
common in arid and semi-arid regions (Bull 1977, Harvey 1989a),
but also occur in humid environments especially in mountain areas
where sediment generation is high in relation to transport
capacity (Rachocki and Church 1990).

Because the streams feeding alluvial fans are at or near the
threshold of critical stream power (Bull 1979) they are
especially important to the geomorphology of mountain or upland
regions for two reasons. First, they are sensitive to changes
in run-off and sediment generation from the mountain catchments
and preserve in their constituent sediments the most complete
record of the erosional history of the mountain catchment.
Second, from a functional point of view, they may act as major
sediment stores within the fluvial system, trapping most or all
of the coarse sediment generated within the mountain catchments
and therefore buffering the system downstream from the effects
of changes in sediment production. The effectiveness of this
buffering role will depend upon the processes on the fan itself.
An aggrading fan may trap all the coarse sediment generated
within the mountain catchment and therefore effectively break the
continuity of the system, but if the fan surface becomes
dissected then channel continuity and sediment transport
continuity from mountain areas to mainstream systems may be
restored.

The factors controlling alluvial fan behaviour are firstly
those controlling water and sediment supply to the alluvial fan
and secondly the transport efficiency through the fan. The first
group depend on catchment characteristics, on for example the
tectonic context the geological and geomorphological history,
relief, erosion rates, climate and hydrology. The second group
depend in part on the transport mechanism through the fan
environment but also on the fan morphology, itself the product
of previous alluvial fan deposition. This paper examines the

influence of catchment characteristics on the sediment transport
characteristics and in turn their influence on fan morphology
paying special attention to progressive changes in fan morphology
resulting from environmental change. It draws on examples from
a wide range of climatic environments from the arid Mojave
Desert, California, to semi-arid Mediterranean region fans in
Spain and Greece, to humid region fans in northwest England, over
a range of timescales from the mid Quaternary to the present day.

SEDIMENT TRANSPORT PROCESSES ON ALLUVIAL FANS

Classically sediment transport processes on alluvial fans
have been grouped into three (Blissenbach 1954); debris flow,
stream flow and sheet flow processes. More recent work by Wells
and Harvey (1987), on the basis of the characteristics of
sediments produced by a major storm on small alluvial fans in the
Howgill Fells, northwest England, has identified four types of
depositional environment and therefore four types of sediment
transport mechanism. These are: (i) true debris flow deposits,
characterised by a matrix supported fabric showing push
structures, including vertical clasts, aligned parallel with lobe
fronts, and surface pressure ridges: (ii) transitional deposits,
characterised by clast-supported, structureless gravels with
matrix only in the lower part of the deposit. These deposits
have been interpreted as transported and deposited by
hyperconcentrated flows (Pierson and Scott 1985) whereby
transport takes place as a very wet mass flow but on deposition
the wet fluid matrix drains, or at least partially drains.
Sediment concentration in the fluid gives the material sufficient
strength for mass transport but on deposition is sufficiently
dilute to drain, causing the clasts to settle and giving the
deposit its almost structureless clast-supported fabric. (iii)
fluvial bar deposits, showing a well developed fluviatile
imbrication and a clearly developed nose to tail sorting. These
deposits show clear evidence of fluvial transport, often with a
strongly developed 'bar and swale' relief. They include features
similar to those described as sieve deposits (Hooke 1967). (iv)
fluvial sheet deposits: these are well sorted thin sheets of
fluviatile gravels locally showing a weak bar and swale
topography on the surface.

Matrix particle size varies between these four types of
deposit. The matrix of the true debris flows has an abundance
of silt and clay, whereas the matrix in the fluvial deposits is
almost wholly of sand. Where present, that in the transitional
deposits has particle size characteristics mid-way between those
of the true debris flows and of the fluvial deposits (Wells and
Harvey 1987).

In that study a general model was proposed relating the
transport and depositional processes to the relative fluidity of
the water sediment mix fed in from the catchment, and identifying
the controlling catchment characteristics. (Fig.2) Although
based on the response of the Howgill fans to one major storm, the
same general principles should apply to the dominant mode of
deposition over a period of time on any group of fans. The

Howgill fans were then grouped into two groups; fans dominated by debris flow/transitional deposits and fluviatile fans, and the two groups were successfully discriminated on the basis of a variety of catchment characteristics relating, on one axis, to drainage area, and on the other to various measures of catchment steepness (Wells and Harvey 1987). The results accord in a general sense with the trends identified by Kostachuk et al. (1986) working in the Canadian Rockies, and with those identified by Harvey (1984a) working in Spain. In all cases debris flow fans with Trollhein-type sedimentary sequences (Miall, 1977) are associated with small steep catchments, whereas fluvial fans with Scott-type sedimentary sequences (Miall 1977) are associated with larger and less steep catchments.

In this paper the catchment characteristics discriminating between debris flow and fluvial fans are examined for fans in a variety of climatic environments and over a range of timescales.

STUDY AREAS
Zzyzx fans, California.
The Zzyzx fans (Fig.3) issue from the eastern flank of the Soda Mountains in the Mojave Desert, California, an arid region with c.100mm mean annual precipitation. Geology of the source area comprises Mesozoic granites and metamorphic rocks. The fans occur along a faulted, but long since stabilised, mountain front and backfill into the mountain catchments. They toe out at the margins of Pleistocene Lake Mojave, now a dry playa, Soda Lake. The fan surfaces have been mapped and segments assigned to late Pleistocene and late Holocene ages on the basis of desert varnish and soil development (Wells et al., in preparation), and their relations to dated late Pleistocene and Holocene lake shorelines (Wells et al. 1987). The fans (Fig.4) range from debris flow fans issuing from small steep basin to fluvial fans issuing from larger and less deep basins, but there is an overall trend towards increasing fluvial activity and deceasing debris flow activity from Pleistocene to Holocene times. On the larger fans this has been associated with the development of fanhead trenches and the progradation of the distal fan area (Wells et al., in preparation; Harvey and Wells, in preparation).
Southeast Spain
Numerous fan groups occur in south east Spain (Fig. 2) issuing from mountains of the Betic Cordillera into adjoining lowland areas or at the margins of adjoining Neogene sedimentary basins. The region has a semi-arid climate with mean annual precipitation ranging from c.350mm in the northern part of the area in Alicante Province to less than 200mm in the south in Almeria. The fan surfaces range in age from mid-Pleistocene to Holocene and most are dissected by deep fanhead trenches, some also by mid-fan trenching, and some are dissected throughout (Harvey, 1987a, 1989, 1990).
For the purposes of this paper the fan groups are divided into three: (i) those in southern Alicante and northern Murcia provinces, which issue from catchments on Mesozoic limestones and on tough low grade metamorphic rocks. They include both debris flow (Fig. 4) and fluvial fans. These areas have had a more complex aggradation and dissection history during the Pleistocene

than areas further south (Harvey, 1984b), (ii) fans in southern
Murcia and Alicante provinces issuing from schist catchments,
mostly simple fans (Fig. 4) which are dominated by fluvial
deposits (Harvey 1984a), (iii) other fans in southern Murcia
and Almeria provinces issuing from non-schist catchments which
include debris flow fans but have a less complex Pleistocene
history than those in the area to the north (Harvey 1984a).

Methana, Greece

Around the margins of the Methana peninsula (Fig. 3) fans
issue from catchments on early to mid-Quaternary Dacite/Andesite
volcanic rocks. They include both debris flow and fluvially
dominated fans of late Quaternary age. The larger fans have
been dissected by deep fanhead trenches (Fig. 7) and in some
cases Holocene fan segments prograde in the distal areas. The
climate is a dry Mediterranean climate, with mean annual total
of strongly seasonal rainfall of c.400mm.

Howgill and Bowland Fells, Northwest England

Holocene alluvial fans have been studied in two upland areas
of northwest England, the Howgill and Bowland Fells, humid
regions with mean annual precipitation ranging from c1500mm in
the Howgills to c1900mm in Bowland (Harvey 1986). In both areas
the fans date from the younger Holocene (Harvey et al 1981;
Harvey and Renwick 1987; Harvey 1985), apparently resulting from
accelerated hillslope erosion induced by vegetation changes in
Bowland at c2000Brand in both areas at c1000BP. Source area
geology in the Howgills is of late Pleistocene periglacial
hillslope sediments overlying tough Silurian mudrocks. In Bowland
the source area geology is of late Pleistocene periglacial
hillslope sediments overlying very coarse massive sandstones of
Carboniferous age. Both areas include both fluvial fans and
debris cones Some composite fans occur in the Howgills (Fig. 4).
In both areas sedimentation on the fan surfaces has effectively
ceased, apart from sedimentation induced by rare catastrophic
flood events, with a return period perhaps of c.100 years (Harvey
1986, Wells and Harvey. 1987).

In each study area fan surfaces have been mapped and
assigned to dominantly fluviatile or dominantly debris flow
origins, for the older fan surfaces on the basis of the
sedimentology of the deposits exposed in fanhead trenches, and
for modern fan surfaces on the basis of surface sedimentary
features. For each fan surface segment identified, the fan slope
has been measured by profiling along the axial slope of the fan
segment. For each fan or fan segment the drainage area has been
delimited in the field onto large scale topographic maps, then
from these maps catchment morphometric data, drainage area,
basin length, basin relief, the latter two giving mean basin
slope, have been derived from these topographic maps. This
provided the primary data set for the consideration of the
influence of catchment characteristics on sediment transport and
deposition processes, and the implications for fan morphology
and development.

CATCHMENT CONTROLS OVER DOMINANT PROCESSES

The previous work, referred to above, has identified that
for given climatic and geological context the water:sediment mix
fed to the fan environment from the mountain catchment may act

as a major control over sediment transport and deposition
processes. If we plot drainage area as an index of water
discharge from the mountain catchment, against some measure of
slope steepness as an index related to potential sediment supply,
it may be possible, on any one group of fans, to identify
threshold combinations of these two variables to differentiate
between debris flow and fluvial processes. Two immediate
problems arise in this simple methodology. First, it only
relates to conditions at the fan apex and does not take into
account down-fan variations in the water:sediment mix which lead
to the proximal to distal facies variations on many alluvial fans
(Harvey 1989a). This may not be too great a problem in simple
mountain-front fans, where there is little extra water or
sediment supply away from the fan apex and down-fan facies
changes may be predictable. However on back filled fans,
drainage area increases down-fan and this must be taken into
account. Furthermore, there may be addition of proximal type
sediments in mid-fan from adjacent side slopes, which may
otherwise complicate simple thresholds. This problem is
considered below especially in the context of the Zzyzx fans.

Zzyzx fans

The Zzyzx fans were mapped in the field grouping the fan
surface segments into late Pleistocene, Holocene or active modern
age categories (see above). For each segment the dominant
depositional process, debris flow or fluviatile, was inferred
either from exposures in fanhead trenches or at the margins of
incised channels, or by the sedimentary characteristics of the
exposed surfaces. For each fan segment, drainage area was then
plotted against drainage basin slope, the latter being the
steepness-related variable, which when plotted against drainage
basin area gave the clearest separation of plotting position.

Three separate plots were derived for the Zzyzx fans. The
first (Figure 5a), deals only with fan apex environments and is
therefore directly comparable with true mountain front, rather
than backfilled fans. Also shown on this plot are data derived
from other fans in the Mojave and Sonoran deserts (Harvey 1987).
The fan surfaces shown all relate to late Pleistocene deposition.
The Zzyzx fans (shown by the solid symbols on Figure 5a) are
nearly all dominated by debris flow deposition at their apices,
and contrast with the other, dominantly fluvial fans. It is
clear that the Zzyzx debris flow fans are associated with small
steep catchments, contrasting with the much larger and less steep
catchments for the fluvial fans, but because of an absence of
data relating to drainage areas of between 0.2 and 1 km^2, the
pecked line shown on Figure 5a, suggesting the approximate
threshold between debris flow and fluvially dominated systems,
can only be very approximate.

To examine this threshold more precisely and specifically
in relation to the Zzyzx fans, drainage area characteristics
feeding the mid-point of all mapped fan segments have been
determined and plotted on Figure 5b. Here three types of fan
surface are shown, Pleistocene, debris flow and fluviatile fan
surfaces and Holocene fluviatile fan surfaces. The pecked line
on Figure 5b suggests a threshold between debris flow and
fluviatile depositional environments, one about which there are
relatively few anomalies. The plotting positions of the Holocene

fluviatile surfaces, dominantly to the lower right of the graph reflect the tendency for Holocene fanhead trenching and distal aggradation by fluvial processes (Wells et al. in preparation).

It has also been possible to consider the controls over modern, active processes on the Zzyzx fans (Figure 5c). Five categories of modern processes can be recognised. In some fan apex areas and locally further downfan there are recent debris flow deposits. Two fans show little evidence of any geomorphic activity, and these are classified as 'passive', in contrast to deep narrow channels actively incising into older fan surfaces. Two styles of fluvial transport or depositional activity are apparent, either within wide shallow channels, identified here as 'fluvial transport', or as sheets of fluviatile gravels deposited in the distal zones of the modern fans, described here as 'fluviatile fan deposition'. Two thresholds can be suggested here (shown by pecked lines on Figure 5c) one between debris flow deposition and channel incision, both associated with relatively steep catchments, but with incision occurring only on fans draining the larger catchments, presumably as a result of higher stream power. The other threshold, approximately defines the upper limit for fluvial transport and deposition processes. Its position when compared with that on Fig. 5b suggests that at the present day fluvial processes are dominant in some areas which were producing debris flows during the late Pleistocene. This reflects a progressive trend from the Pleistocene through the Holocene of increasing fluvial dominance on these fans apparently reflecting increased aridity, and presumably associated with a reduced sediment supply from the mountain slopes but an increased stream power during flood events typical of arid regions (Baker 1977).

Mediterranean fans, Greece, Spain.

Similar methodology has been used for the Mediterranean region fans in an attempt to identify the catchment controls over debris flow and fluvial processes. For the Methana fans (Fig. 6a) data relating only to Pleistocene fan surfaces are included. Most of the fans are deeply trenched and Holocene fan segments are restricted to small distal areas. The dominant sedimentary process in each case has been identified from exposures in fan trenches. The fans are not simple mountain-front fans, but a back-filled fan complex at Kipseli, and coalescent, multi-sourced fan aprons in the other areas. (Fig. 3). Despite the problems of drainage area delimitation a reasonable threshold can be identified on Figure 6a with only two apparently anomalous fan segments.

Previous work on the Spanish fans (Harvey 1984a, 1987b, 1990) has shown that different groups of fans have had rather different Pleistocene aggradation and dissection histories and show rather different morphometric characteristics. For this reason the Spanish fans are treated here as three separate groups. The fans issuing from the limestone Pre-Betic mountains in Alicante and those issuing from the massive low grade metamorphic rocks of the Sierra de Carrascoy and the coastal Sierras in northern Murcia all show strongly calcreted Pleistocene fan surfaces, deep fanhead trenches and a relatively complex history of Pleistocene aggradation and dissection. Some are simple mountain front fans, others are back-filled fans.

For individual fan apices (Figure 6b) a reasonably clear threshold discriminates between debris flow and fluvial fans with few major exceptions. One of the exceptions relates to a small fan in the Torre group (location, Fig. 3) which produces a sediment sequence dominated by stratified fluviatile gravels rather than the debris flows that would be expected on the basis of the steep mountain catchment, however, this source area is in a relatively high grade fissile schist which may not be conducive to the production of debris flows.

On the basis of previous work, fans in southern Murcia and Almeria are sub-divided into those issuing from high grade fissile schist catchments and those from non-schist areas. For the non-schist areas (Fig. 6c), with the exception of fans from the Cabo de Gata volcanics, a reasonably clear threshold can be identified, despite the wide range of bedrock geologies. Most of these fans are simple mountain front fans with well developed calcrete crusted fan surfaces, shallow fanhead trenches and Holocene fluviatile distal fan deposits. Only the Pleistocene upper fan segments are included here.

In the schist areas (Fig. 6d) only two small fans are dominated by debris flow deposits, but three others show occasional debris flow deposits at the base of the sedimentary sequences. If these are grouped with the two debris flow fans a threshold can be identified separating this group from the majority fluviatile fans. Some of these fans are simple mountain front fans, but many, especially issuing from the Sierra de los Filabres are back-filled fan complexes. The fan sequences are relatively simple, with aggradation often until late Pleistocene, often with little calcrete development, limited fan trenching and Holocene distal aggradation. Again only proximal fan surface segments are included here.

Northwest England, Howgills and Bowland Fells

The sedimentary style for Howgill and Bowland fans has been determined either from exposed sections or from fan surface morphology. For the Howgill fans there is a contrast between debris flow fans or debris cones, composed of ill-sorted matrix-rich debris flow deposits whose surfaces preserve lobate debris flow morphology, and fluviatile fans composed of stratified, better sorted matrix-poor cobbles whose surfaces are either smooth or lightly channelled. Some of the Howgill fans are composite forms with proximal debris flow deposits from which distal fluvial fan segments have been flushed. If these are considered primarily as debris flow fans a reasonably clear threshold with few anomalies can be identified (Fig. 7a). Two anomalous fluvial fans issue from steep catchments, both in the same valley, Upper Bowderdale; both are small fans and could be younger than the main group of fans.

The debris cones in Bowland show ill-sorted largely unstratified boulder or cobble deposits with a sandy matrix whereas those classified as fluvial fans show better sorting and clear stratification. Because of the coarse sandy nature of the periglacial hillslope parent materials the debris cones were probably deposited by transitional rather than true debris flow deposition. Despite the limited number of fans in Bowland an approximate threshold can be suggested (Fig. 7b).

6

Catchment controls: summary

When the thresholds suggested on Figs. 5 to 7 are plotted together (Fig. 8), it is clear that within the various regions different catchment characteristics influence the relative importance of fluvial and debris flow processes. The most "fluvial" fans are those in the arid American southwest. In that area modern fan processes are more "fluvial" than they were during the Pleisto,cene. The Mediterranean region fans are more debris flow rich, though it must be remembered that these fans relate to Quaternary climates rather than present day Mediterranean climates. Climates during Pleistocene "glacial" phases appear to have been cold, dry, with relatively high rates of weathering in the mountain catchments but higher storm-induced seasonal run-off than today (Sabelberg 1977). It may be due to climatic or geological control that fans in the more arid south of Spain are more "fluvial" than those further north. Not surprisingly for a humid region, the Bowland fans are the least "fluvial", but the strongly "fluvial" plotting position of the threshold for the Howgill fans appears to be little anomalous. Perhaps the reason may relate to the low infiltration capacities and steep slopes in the Howgills coupled with the fact that the fans are the product of a major phase of gully erosion, whereas in Bowland slope failure appears to be more important in sediment supply, and with the sandier parent material run-off rates may have been lower.

A major implication of these results is that, not only do drainage basin characteristics, drainage area and slope characteristics, and presumably at a local scale geology, influence the threshold between debris flow and fluvial processes, but climate may also be a major control. This may be important not only in the context of contrasts between regions but also in the context of geomorphic responses to climatic changes. The major reason behind this appears to involve the relationship between stream power and critical stream power in the mountain front zone, or in more general terms the relationship between sediment and water generated in the mountain catchment during flood events.

MORPHOLOGICAL IMPLICATIONS

Previous work on alluvial fan morphology (Bull 1962b, Hooke 1968, Harvey 1989a) has indicated that sediment transport processes on alluvial fans influence fan morphology, especially through their influence on depositional slope. Deposition by debris flows takes place at greater slopes than deposition by fluvial processes, and within fluvial processes deposition by unconfined sheet flow takes place on greater slopes than deposition by channel processes. Therefore, as catchment characteristics influence sediment transport processes so do they influence fan morphology. In addition to this influence, controlled by the water : sediment mix, fed to the fan environment, is an influence related to particle size, itself partly dependent on catchment topography and partly on catchment geology (Elissenbach 1952, Bluck 1964, Lustig 1965, Denny 1965, Hooke and Rohrer 1979). Coarser sediments will be deposited on steeper slopes than finer sediments. Furthermore, these influences are likely to change downfan from proximal to distal environments with selective deposition of coarse material at the

fan apex and possibly dilution downfan, leading to a general concavity in the longitudinal profile of the fan. This tendency is accentuated on backfilled as opposed to mountain-front fans, where drainage area increase downfan. For particular groups of fans the simplest of these relationships can often be expressed by a regression equation in the form:

$$G = aA^b$$

where G is fan slope and A is drainage area, but the strength of the correlation can often be improved by taking into account some measure of catchment steepness (Harvey 1987a). In this paper these relationships are considered for the several study areas, taking into account the differences between debris flow and fluviatile fans. For simple mountain front fans the fan slope is measured for the proximal third of the fan but for backfilled and composite fans slopes are measured and appropriate drainage basin characteristics derived for individual fan segments in proximal, mid-fan and distal fan locations, thereby taking into account the increase in drainage areas downfan.

For the American fans (Fig. 9a), there is a clear distinction between debris flow and fluviatile fans, with the debris flow fans having much higher gradients per drainage area. Given the moderate correlations no real difference is apparent in the slopes of the regression lines. When the Zzyzx fan surfaces are treated separately (Fig. 9b) there is again a clear separation between the debris flow segments and the fluvial segments. The correlation for the Pleistocene fluvial surfaces on their own is not significant and only becomes so when all fluviatile fan surfaces, Pleistocene and Holocene are considered together.

For the Mediterranean region fans (Fig. 10), in all cases there are again clear separations between debris flow and fluvial fans. On Methana (Fig 10a), the slope of the regression line for the debris flow fans is less than that for the fluvial fans. For the Spanish northern group (Fig. 10b) the correlation for the fluvial fans is not significant. This group includes a wide range of fan types from basins with a wide range of geology and topography (Harvey 1987a) and includes the tectonically disturbed Carrascoy fans (Harvey 1988). However, the debris flow fans clearly have higher gradients per drainage area than do the fluvial fans, and when an all-fan regression is calculated for this group, although still not significant the 5% level, residuals for the debris flow fans are in all cases but one positive and for all but four of the fluvial fans are negative. The mean residuals for the debris flow and fluvial fans are significantly different at the 1% level on the basis of Student's t test. For the two southern Spanish groups there are again clear distinctions between debris flow and fluvial fans, and in both cases the slopes of the regression lines for the debris flow fans are less than those for the fluvial fans.

There is a similar picture for the two groups of English fans (Fig. 11), with higher gradients per drainage area for the debris flow fans, and lower slopes to their regression lines. Although there are only four fluvial fans in the Bowland group it is clear that they follow a similar behaviour to those in the Howgill group.

Considering these regression analysis together (Fig. 1?, Table 1), several patterns emerge. In some, but not all cases, the correlations are stronger for the fluvial fans. As would be expected, in all cases the fan gradient per drainage area for the debris flow fans is greater than that for the fluvial fans. In most cases the slopes of the fluvial regression lines are greater than those for the debris flow regression lines, suggesting that the fluvial fans are more sensitive to selective sediment transport either within the mountain source areas or within the fan environment itself. The overall differences in plotting position (approximately equivalent to 'a', Table 1) do not seem to reflect climatic differences, and are probably related to geology and relief differences within the source area.

In order to examine the influence of relief characteristics within each fan group, multiple correlation and regression analyses were carried out (Table 2), using drainage basin relief and drainage basin slope as measures of the basin steepness. In almost all cases correlation coefficients significantly improve and standard errors of the estimate significantly decrease when one or other of the two steepness characteristics are taken into account together with drainage area. In most cases basin relief gives better results than basin slope. For most of the major relationships multiple correlation coefficients are at least 0.7; even those relationships that had non-significant correlation coefficients for drainage area alone improve to significance and correlation coefficients of at least 0.5. With two exponents involved it is a little difficult to make direct comparisons between the multiple regression equations, but for the all-fans relationship has there are still major differences between the residuals for the debris flow and fluvial fans, reflecting the different nature of the catchment controls over each process.

DISCUSSION

It is clear that drainage basin characteristics influence the sediment transport processes on alluvial fans which in turn influence fan morphology, especially fan slope. The most important drainage basin characteristics can be divided into three groups, those related to topography, geology, and climate/vegetation. The topographic controls, dealt with in this paper, are the size of the drainage basin influencing the volume of water supplied to the alluvial fan, and the steepness of the drainage basin relating to the potential for sediment supply. Geological controls may account for differences between fan groups within any climatic environment, and include the availability of fine sediment, hence the propensity for debris flow processes, and the particle size characteristics of the coarse sediments which influence depositional slope. The climate/vegetation controls influence the generation of water and sediment on the hillslopes of the drainage basin, and the transport efficiency through the drainage net to the alluvial fan environment. Particularly important are the effectiveness of storm rainfall, and the rate and style of sediment generation, whether by slope failure or by overland flow processes. Of these groups the third is the most likely to change with time,

therefore the most likely to cause changes in alluvial fan morphology.

If climate or vegetational cover were to change in such a way that altered the effective water: sediment ratio during storm events, the sediment transport processes on the alluvial fan would also change (essentially involving a lateral shift in the position of a distribution curve on Fig.2). As the new process would be operating on fan slope or fan channel slope created by former processes a change in the aggradation/dissection regime of the fan could occur. This would be especially important if a change took place from debris flow to fluvial processes in that fan dissection could be initiated. There are three zones within alluvial fans that may be prone to dissection. The first zone is at the apex of an undissected fan where dissection could lead to the development of a fanhead trench (Schumm et al. 1987). The second zone is in mid-fan on a fan whose proximate areas are dissected by a fanhead trench, and where the fan channel emerges onto the fan surface at an intersection point (Hooke 1967). On most fans deposition occurs here (Wasson 1974), but on some dry-region fans, especially where the fan surface is cemented by calcarete, mid-fan incision may occur (Harvey 1987a). The third zone is at the fan toe, but this is related to different controls, either to base level changes causing incision or to the lateral migration of a mainstream to which the fan is tributary.

The relationship between process change and morphological change has two important implications for fan dynamics, first in the context of the history of fan development, and second in the context of the vulnerability of alluvial fans to contemporary environmental change. Providing it can be distinguished from the general long term trend of fanhead incision and distal progradation (Eckis 1928), sequential changes in fan dynamics produce a suite of fan profile types (Fig.13, and Harvey 1987a), ranging from type O for an aggrading fan, types A and B for fanhead trenched but distally aggrading fans, type C for fan with mid-fan scour but no substantial incision, types D and E for fans subject to mid-fan erosion, type F for a wholly dissected fan and type G for a fan subject to fan toe incision. Harvey (1987a) has attributed mid-fan erosion to conditions where unit stream power increases rather than decreases in mid-fan, on fans where the channels are relatively narrow and there is a major slope difference between fan surface slope and channel slope (Fig.12). Profile types for the study fans (Table 3) can be compared with the changes in the dominant processes during fan development (Table 4). For the American and Mediterranean groups these trends relate to timescales from the Pleistocene into the Holocene, and for the English fans for environmental changes during the Holocene. The Zzyzx fans show a trend away from debris flow processes and a wide range of profile types. The dominantly fluvial other American fans characteristically have A/B profiles. The Methana fans show some shift from debris flow towards fluvial processes, but the most important profile type is type F where dissection is base-level induced (type G). Similarly the Spanish northern group show some trend from debris flow towards fluvial processes and include many examples of mid-fan scour or dissection (Types C, D and E), but the southern

Spanish groups are overwhelmingly fluvial and dominated by A/B type profiles. The Howgill fans include many simple aggrading fans (Type C) and many of types A/B often associated with a switch from debris flow to fluvial processes, involving the flushing of the deposits towards distal fan environments. This is very similar to some fans in western Scotland, an area more recently glaciated than the Howgills, where late Pleistocene to early Holocene paraglacial debris cones have been flushed out later in the Holocene to produce distal alluvial fans (Brazier et al 1998). Many of the Bowland fans have been dissected at the toe by lateral erosion by the mainstream. Putting these two data sets together it is possible to see some association between process trend and profile type (Table 5). Active and stabilised debris cones are largely of type O. Fans which have seen a switch from debris flow to fluvial processes show profiles dominated by fanhead trenching (types A/B) or mid-fan scour or dissection (type C,D,E). Dominantly fluvial fans show a range of profile types similar to those which have switched from debris flow to fluvial processes. No fans show a switch from dominantly fluvial to debris flow processes and on most of those fans which have switched from deposition, either by debris flow or fluvial processes, to incision, the incision has been primarily related to incision at the toe (type G).

Interestingly, though the Pleistocene to Holocene process and morphological trends are similar in both the American and Mediterranean study areas, the underlying climatic causes do differ. In the American case the Holocene aridification has been associated with the cessation of debris flow activity on the hillslopes, the reduction of sediment supply to the fan environment, the increasing effectiveness of flood run off and the trend towards fan dissection. In the Mediterranean case, cold arid 'glacial' climates produced abundant sediment from the hillslopes, but an apparent decrease in aridity during the Holocene (Harvey 1984a, Sabelberg 1977) has been associated with some switch from debris flow to fluvial activity and an overall dissection trend. This is presumably in response to a reduction in weathering rate within the mountain catchments. The British fans were initiated apparently by human induced vegetation changes, with some subsequent trend from debris flow to fluvial processes or to stabilisation, presumably as revegetation took place within the source areas.

Because alluvial fans are dynamic features susceptible to environmental change, it is relevant to consider their potential response to currently postulated climatic changes resulting from global warming (Harvey 1989b). Climatically induced vegetation change could influence hillslope erosion rates and therefore sediment supply to the fan environment. In dry regions the climatic trend could be to increasing aridity and a reduction in vegetation cover and an increase in sediment supply, but in humid areas accelerated vegetation growth could lead to stabilisation and a reduction in sediment supply (Fig 14). In most regions an increased storm frequency or storm severity could increase runoff rates and hence stream power. These two changes could cause a shift in the position of the curves on Figure 2, representing a change in transport process and/or erosion/deposition relationships. The zones particularly

susceptible to change are the fan apex, especially on undissected fans, but fanhead trenches themselves could be subject to further dissection, and the intersection point/mid-fan zone. In addition a change in mainstream behaviour, where present, at the fan toe could alter the local base level.

In conclusion, alluvial fans are critical zones within the fluvial system, situated between mountain source areas and main drainages. Processes on alluvial fans influence the continuity of sediment movement through the system. Furthermore, fan morphology is both a product of sediment transport processes and influences the effectiveness of sediment transport through, or deposition on the fan. That these processes may change through time makes it especially important that both the morphology and the processes are studied within their temporal contexts.

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Table 1

Correlation and regression analysis for the relationships:-
$G = aA^b$ (where G is fan gradient, A is drainage area km^2, R is
correlation coefficient, SE is standard error of the estimate, log
units)

Fan group		n	R	SE	a	b
American	Debris flow	13	-0.788	0.061	0.161	-0.188
	Fluvial	13	-0.779	0.034	0.033	-0.141
Dayton	Debris flow	33	-0.681	0.066	0.148	-0.101
	Plastic flow	15	no correlation			**
	Wet flow	11	-0.339	0.164	0.151	-0.134*
	All flows	33	-0.473	0.157	0.151	-0.134
Methana	Debris flow	11	-0.765	0.047	0.171	-0.177
	Plastic flow	3	-0.911	0.099	0.131	-0.309*
	All flow	11	-0.681	0.111	0.100	-0.471
Spain N	Debris flow	11	-0.545	0.111	0.108	-0.108*
	Fluvial	15	no correlation			**
	All fans	17	-0.500	0.099	0.104	-0.000*
Spain S convection	Debris flow	11	-0.403	0.111	0.074	-0.066*
	Fluvial	11	-0.551	0.077	0.051	-0.171
	All fans	11	-0.500	0.161	0.051	-0.109
Spain S coast	Debris flow	5	-0.866	0.159	0.101	-0.101*
	Plastic flow	16	-0.800	0.099	0.051	-0.100
	Wet flow	7	-0.651	0.119	0.109	-0.101*
	All fans	17	-0.777	0.111	0.059	-0.100
	All fans	17	-0.105	0.101	0.151	-0.105
Howgills	Debris flow	11	-0.767	0.066	0.061	-0.101
	Fluvial	13	-0.714	0.061	0.059	-0.171
Scotland	Debris flow	5	-0.617	0.159	0.215	-0.191*
	Fluvial	4	-0.099	0.199	0.109	-0.994**
	All fans	11	-0.101	0.111	0.109	-0.419

* significant levels $P<0.1$

** insignificant <0.1

Table 1 Logarithmic multiple correlation and regression analyses of Drainage area. km² A, Basin relief r B and Fan slope S on Fan gradient G. R is correlation coefficient, SE is standard error of the estimate, log units.

Fan Group		A → G		A,B → G		A, S → G		"Best" Equation
		R	SE	R	SE	R	SE	
American	Debris flow							
	Fluvial							
Oregon	Debris flow							
	Pleist. flow							
	Rel. flow							
	All flow							
Montana	Debris flow							
	Pleist. flow							
	All flow							
Spain N	Debris flow							
	Fluvial							
	All fans							
Spain S non schist	Debris flow							
	Fluvial							
	All fans							
Spain S Schist	Debris flow		5 fans only					
	Pleist. flow							
	Rel. flow							
	All flow							
	All fans							
Anguilla	Debris flow							
	Fluvial							
Scotland	Debris flow							
	Fluvial		4 fans only					
	All fans							

Figures in brackets are for non-significant > 5% correlations

Table 3: Occurrence of various styles of fan/channel profile relationships for the study areas.

Profile Style (see Fig. 13)

Fan Group	O	A/B	C	D	E	F	G*
Zzyzx	7	5	-	3	4	-	6
Other American	-	6	1	-	-	-	-
Methana	-	3	-	1	-	4	4
Spain, North	1	8	2	10	2	1	1
Spain, South (non-schist)	4	11	3	1	-	2	4
Spain, South (schist)	-	21	1	1	-	6	2
Howgill	14	13	-	-	-	4	5
Bowland	6	-	-	-	-	6	8

* numbers showing toe incision, also included in other groups

Table 4 Progressive changes in dominant fan processes, for
American and Mediterranean fans changes relate to
Pleistocene to Holocene timescales, for English fans
changes relate to the Holocene.   D Debris flow, P
Passive i.e. non erosive/non depositional, F Fluvial
Deposition, I Incision.

Dominant Process Change

Fan group	DD	DP	DF	DI	FD	FP	FF	FI
Szyrx	4	4	6	3	-	-	2	1
Other American	-	-	1	-	-	-	6	-
Methana	-	-	3	1	-	-	4	-
Spain North	-	-	8	2	-	-	15	1
Spain South (non-schist)	-	2	3	3	-	1	11	2
Spain South (schist)	-	-	-	2	-	-	26	1
Howgill	9	5	7	1	-	-	6	3
Bowland	-	3	-	5	-	-	3	1

Table 5   Relation of dominant process change (see Table 4) to Profile
           Change (see table 3, Fig.13), Number of occurrences.

	Dominant Process Change**							
Profile style (see table 3, Fig 13)	DD	DP	DF	DI	FD	FP	FF	FI
O	7	14	1	2	0	1	6	1
A/B	4	0	16	1	0	0	47	·
C	0	0	5	1	0	0	1	0
D/E	2	0	4	6	0	0	11	1
F	0	0	2	7	0	0	9	6
G*	2	2	1	13	0	0	8	8

* Indicates number of fans subject to (local) base-level induced toe

incision, also included in other groups.

**D,P,F,I, as defined on Table 4.

a distally aggrading fans

b distally trenched fans

(see text)

Fig 1

Catchment Geo-morphology	Size of catchment —————— increasing ————————▶			
	Steepness of catchment ——— decreasing ————————▶			
	Erosion area ————————— decreasing ————————▶			
Fan process	Non-Newtonian flow		Newtonian flow	
	Deposition			
	Debris flow	Transitional flow	Stream flow or sheet flow	Erosion

RELATIVE FLUIDITY (WATER : SEDIMENT RATIO)

Fig 2.

Map 1 (Soda Mountains, California)

0 km 1

Major ridges
Alluvial fans

Soda Mountains

Soda Lake (playa)

ZZYZX

Map 2 (Spain)

Pre Betic Ranges

SPAIN

Study area

0 km 200

ALICANTE

MURCIA

Sierra de Carrascoy

Torre fans

Sierra de Almenara

Sierra de los Filabres

Sierra de Alhamilla/ Cabrera

ALMERIA

Cabo de Gata fans

Study sites
1 Northern fan groups
2 Southern fan groups (excluding schist areas)
3 Southern fan groups (schist areas)
Debris fans
Fluvial fans

0 km 50

Map (California)

CALIFORNIA

Study area

0 km 300

Map (Britain)

HOWGILL FELLS
BOWLAND FELLS

0 km 200

Map (Howgill Fells)

TEBAY

HOWGILL FELLS

Langdale

Bowderdale

Weasdale

Carlingill

River Lune

Study sites
Fluvial fans
Debris cones
Composite fans

0 km 2

Map (Greece)

KYPSELI

Western fan complexes

METHANA

0 km 2

GREECE

0 km 100

Study area

Main ridges
Study sites, fan complexes

Map (Bowland Fells)

BOWLAND FELLS

Study sites
Debris cones
Fluvial fans

0 km 2

Langdale

Brook

Dunsop Brook

River Hodder

a
b
c
d
e
f

(a) AMERICAN FANS (General)

Legend:
△ ▲ Debris flow fans
△ ▼ Composite fans
○ ● Fluviatile fans

(b) ZZYZX (Quaternary fans)

Legend:
▲ Pleistocene debris flow fan surfaces
● Pleistocene fluviatile fan surfaces
○ Holocene fluviatile fan surfaces

(c) ZZYZX (Modern processes)

Legend:
△ Debris flows
ρ Passive
ι Incision
. Fluvial transport
F Fluviatile fan deposition

Drainage area km

(a) GREECE (Methana)

(b) SPAIN (Alicante, Murcia North)

(c) SPAIN, SOUTH (Non Schist areas)

V Cabo de Gata fans

Basin slope

(d) SPAIN, SOUTH (Schist areas)

▲ Debris flow fan surfaces

• Fluviatile fan surfaces

⋮ Fan sediments which include
    debris flows   (d only)

Drainage area km²

(a) HOWGILL FELLS

▲ Debris flow fans

△ Composite fans

● Fluviatile fans

Basin slope

0.6
0.5
0.4
0.3
0.2
0.1
0

0.01   0.02   0.05   0.1   0.2   0.5   1   2   5   10

(b) BOWLAND

Basin slope

0.6
0.5
0.4
0.3
0.2
0.1
0

0.01   0.02   0.05   0.1   0.2   0.5   1   2   5   10

Drainage area km^2

Fig. 8

(a) AMERICAN FANS

Legend:
▲ Debris flow fans
⚠ ⚠ Composite fans
○ • Fluviatile fans

D

F

(b) ZZYZX

Legend:
▲ Pleistocene debris flow fan surfaces
• Pleistocene fluviatile fan surfaces
○ Holocene fluviatile fan surfaces

D

(P)

F

H

Fan slope

Drainage area km^2

(a) GREECE (Methana)

(b) SPAIN (Alicante, Murcia North)

(c) SPAIN, SOUTH (Non Schist areas)

Fan slope

(d) SPAIN, SOUTH (Schist areas)

▲ Pleistocene debris flow fan surfaces

• Pleistocene fluviatile fan surfaces

○ Holocene fluviatile fan surfaces

△ Fan sediments which include debris flows (d only)

Drainage area km^2

HOWGILL FELLS

Legend:
- ▲ Debris flow fan surfaces
- • Fluviatile fan surfaces

BOWLAND FELLS

Fan slope

Drainage area km^2

Fan slope vs. Drainage area km² (log-log plot)

Labels on curves: H-D, Z-D, B-D, SS-D, (Z-P), H-F, M-D, A-D, SO-D, Z-F, SN-D, M-F, M-P, Z-H, (SN-F), A-F, SN-A, SS-H, SS-P, SS-F, SO-F, (B-F)

O   NO DISSECTION

A   AGGRADING FAN 'AMERICAN' TYPE

B   LIMITED DISTAL AGGRADATION

C   INTERSECTION POINT SCOUR

D   INTERSECTION POINT HEADCUT DEVELOPMENT

E   DISTAL TRENCHING

F   TOTAL DISSECTION

G   BASE-LEVEL INDUCED DISTAL DISSECTION

Fan surface

Channel profile

Intersection point

Width excessive
Slope difference small

Slope difference great
Width small

Dissecting fans (Types D E)

Aggrading fans (Types A B C)

Crusted

Non crusted

Residuals from regression $S = 0.429 \ G^{0.89}$ (log units)

Residuals from regression $W = 7.12 \ A^{0.25}$ (log units)

		Effectiveness of storm runoff	
		DECREASE ←——————→ INCREASE	
Vegetation cover DECREASE ↕ INCREASE	Hillslope erosion rate INCREASE ↕ DECREASE	Decrease sediment supply Decrease stream power STABILISATION	Decrease sediment supply Increase stream power D' FLOW ⟶ FLUV OR INCISION
		Increase sediment supply Decrease stream power FLUV ⟶ D' FLOW OR AGGRADATION	Increase sediment supply Increase stream power ACCELERATE ACTIVITY (DIRECTION DEPENDS ON RELATIVE INCREASES)

Effect Of Bed Roughness, Particle Shape and Orientation On Initial Motion Criteria

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Introduction.

Considerable attention has been given to defining the initial motion criteria for spherical particles perched on planar beds composed of similar particles. In contrast, there is little information on entrainment thresholds and the mechanics of entrainment for non-spherical particles resting in natural positions on beds composed of particles dissimilar in size and shape to the entrained grain. This is surprising as data from gravel-bed rivers has shown that at a given shear velocity, a mix of grain-sizes and shapes can be entrained.

From this evidence it might be surmised that there is no size selection at all from certain bed fabrics or that selective entrainment is dependant on the characteristics of the bed-material fabric. The presence of an armour-layer is an evident structural element of the bed-fabric in many rivers and may act to impart a degree of equal mobility to the bed-material population but other factors have not been fully evaluated. For example, other than density differences, the nature of the over-all grain-size distribution will influence particle packing and bed roughness. In addition, particle shape influences packing and the probability of entrainment as does the exact position and orientation of the individual particles. Each of these various factors can be parameterized and the variability in entrainment thresholds evaluated. In addition, to some extent, these factors can be incorporated into models of the initial motion process which can be used to explain the equal mobility of divers particles at threshold conditions.

A proper consideration of equal mobility should apply only to initial motion conditions and not to material in transport. Consequently in this paper the variability in entrainment is explored and an initial motion criterion developed which reflects these observations.

In the first part of this paper the preliminary results of a laboratory flume experiment are reported. These data primarily concern the initial motion of individual particles of similar weight in natural mixtures, wherein particle shape and orientation on the bed are variables as is the roughness of the bed. The critical shear velocity associated with initial motion of the test particles was ascertained for each case and the range of threshold values determined for combinations of shape, orientation and bed roughness.

In the second part of the paper, the geometric relationship of ellipsoidal particles (ranging from spheroids to discs) perched on a bed of grains of dissimilar size or shape, is considered in the light of the observations obtained from the flume experiments. Contact angles and critical pivoting angles are derived and an example presented graphically. These data provide input to an initial motion criterion which caters for a variety of ellipsoidal shapes and orientation.

Methods

A series of test runs were conducted in a recirculating glass-walled flume which had an over-all length of 13m, a width of 0.79m and a maximum depth of 0.5m. Near uniform flow, with a discharge up to $0.3m^3s^{-1}$, could be maintained 6m from the inlet baffles over a 5m section. Flow from the channel was controlled by a vertical louvre gate and bed slope was altered by electric motor from zero gradient to $3.75^{O}(0.066)$ with a precision in the vertical elevation of 0.5mm. Artificial stable bed-roughness was created by lining the flume bed with concrete slabs into which ellipsoidal river gravels had been tightly packed with their long axes transverse to the flow such that approximately 50% of each cobble was submerged in the concrete. Three grades of gravel (one phi interval apart) were used (16-32mm; 40-64mm and 125-250mm) to create three roughness types. The largest roughness was similar in size to the test particles whilst the smaller roughness elements represented smoother beds over which coarse gravel is observed to roll in natural streams.

Four sets of test runs were designed to represent idealized bed particle shapes and particle orientations, which nevertheless resemble geometries encountered in natural gravel-bed rivers. In the first series, four artificial well-rounded particles of similar weight (1kg) and density (2.65) were compared. These were a rod, an ellipse, a disc and a spheroid; dimensions of which are given in Table 1. Each particle was placed in exactly the same

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position on the col between four adjacent bed particles such that initial motion would be by displacement over the col between the two downstream particles. Three particle orientations were considered; long axis parallel to flow, transverse and oblique, so that the projected area and stability of the perched particles varied with orientation and bed roughness. For a given discharge, the bed slope was increased steadily until the test particle was entrained into the flow. The velocity profile over the pivot point was then measured using four simultaneously recording Ott C2' flow meters (positioned between 22mm and 100mm above the base of the col) and the local shear stress computed using the log-profile method for rough turbulent flow. The test was repeated five times for each particle, orientation and roughness type.

Notes were made with respect to the mode of entrainment (eg. whether the particle vibrated, swiveled, imbricated, slid or rolled etc). The disc and ellipsoid displayed a tendency to imbricate on the roughest bed and so the shear velocity required to imbricate the particle and that required to entrain the particle into the flow were recorded as appropriate.

The second series of runs compared the entrainment characteristics of a sphere and a cuboid of similar weight (2.4kg) and density (2.65), whilst the third series examined the differences in the threshold shear stress for two natural cobbles (of similar shape but contrasting weight - Table 1), when the orientation and bed roughness were varied. There was little variability amongst the shear velocity values for the five replicates in the first test series, and so for the latter two series, three replicate runs

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were conducted for each set of test conditions.

The final series of experimental runs dealt solely with influences of flow velocity and bed roughness on particle velocity once entrainment had occurred. The mean velocity conditions were established and the velocity of the particle was measured over a distance of 5m.

Experimental Results

Mode Of Initial Motion

Particles frequently vibrated in situ before entrainment occurred. This was less vigorous in the case of the disc, but occurred with all roughness types. In the case of roughness types 1 & 2, all particle shapes (excluding the rod) placed with their A-axes parallel to the flow, tended to rotate out of place without assuming a transverse attitude first. This observation was true even for the disc which never slid out of place but instead flipped over; turning around a pivot point located near the edge of the disc. However, on the smoothest bed (Roughness 1) the ellipse and disc placed parallel or oblique to the flow, swiveled to a transverse attitude before rolling out of place. Only the rod, placed parallel or oblique to the flow always swiveled to a transverse attitude before becoming entrained. The sphere was entrained by rolling in all cases, whilst the cuboid always slide out of it's resting position but then rolled about it's A-axis. Rolling about the A-axis was the usual mode of travel once entrained for all shapes. The exception to this was the disc, at the highest velocities, when motion could be by rolling

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along either of the three axes or by sliding. On the roughest bed (Roughness 3) the ellipse and disc always slid to an imbricated position before being entrained by rolling.

Initial motion on the smoothest bed (Roughness 1)

The shear velocity data for the various particle shapes and attitudes are summarized in Fig. 1. On the smoothest bed the most stable attitude is for an ellipse positioned with the A-axis parallel to the flow, whilst the least stable is the rod, placed transverse to the flow. It is evident from the range of data values, that particle orientation is as important as particle shape and relative mass in dictating the critical threshold. Orientation alters the cross-sectional area of the particle exposed to the main stream-line and to a lesser extent effects particle stability vis-a-vis the contact points of the particle with the under-lying bed particles. For example, not only does turning a rod from parallel to oblique to the flow increase the cross-sectional area exposed to the flow, but the rod must be tilted and rested across both a bed particle and a col position rather than assume a horizontal attitude over two col locations.The threshold shear velocity for the parallel rod is more than double that for the transverse rod; the ratio being 2.16 which is greater than any ratios comparing shapes rather than orientation. A similar difference applies to the ellipse (1.86) but the differences for the spheroid and disc are not so acute (1.33 and 1.26 respectively) but still remain as important as differences in shape. The sphere of course can have only one orientation, and despite the difference in mass (2.4kg),compared with the other artificial particles (1kg), a small pivoting angle

ensures entrainment occurs at a low threshold value (9.8 cm s^{-1}) equal to that required to entrain the transverse ellipsoid. The cuboid, although the same weight as the sphere, requires a higher shear velocity for entrainment by sliding, than is needed to entrain the sphere, but even here the entrainment shear velocity is not dissimilar to that associated with the rod, ellipse and spheroid in a variety of orientations.

Initial motion on the intermediate roughness (Roughness 2).

Roughening the bed with coarser material, which results in moving the pivot points further apart, provided an opportunity for the test cobbles to assume a more stable attitude, sitting deeper into the col between the bed particles. Consequently, the pivoting angle in all cases is increased and the necessary applied critical shear velocity to dislodge the cobbles also increases (Fig. 2). As might be expected the relative difference between the critical shear velocity for the parallel and transverse orientations changes little. However, the change in the relationship between shear velocity, roughness and test cobble size is not independent of the cobble shape. Whereas on the smoothest bed the parallel ellipse was the most stable, on the rougher bed the parallel rod is the most stable. More importantly, as the amplitude of the bed roughness increases, the tilt of the oblique particles also increases so that they are less stable than on Roughness 1 and easily roll to a transverse orientation. Consequently there is very little difference between the threshold values for oblique or transverse orientations for any individual shape. Whereas the sphere was very unstable on Roughness 1, increasing the roughness

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amplitude provides an increasingly stable base. This increase in stability is not reflected in the case of the cuboid which cannot sit in the col between the bed particles selected but largely bridges across the gap and still sits on the tops of the bed particles. as it did on Roughness 1. Consequently, the difference between critical values for sphere and cube are now diminished.

Initial motion on the roughest bed (Roughness 3)

Selection of a roughness similar in dimensions to the test particles, meant that up to 17% of the cobble in some cases was not protruding into the flow. More importantly, for particle stability, the greater roughness amplitude provided an opportunity for cobbles to rotate within the col position and to imbricate. This latter series of tests shows quite clearly the importance of flattening of the C-axis to the initiation of imbrication and enhanced bed stability. Only the rod was entrained from any orientation without imbrication occurring first. The spheroid always rotated into an imbricated position $(22^{O}-32^{O})$ before entrainment. The high shear velocities needed to entrain particles on the rough bed, resulted in high initial particle momentum so that oblique (unstable) cobbles did not roll into new transverse (potentially more stable) locations but were entrained at lower threshold values than the parallel or transverse particles (Fig.3). This result is at variance with the observations from the two less-rough beds (Figs. 1 & 2). The ellipse could only be entrained when placed in the oblique orientation, whilst in other attitudes both it and the disc rotated into imbricated $(18^{O}-24^{O})$, stable positions. The shear velocity required to induce

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imbrication was greatest for parallel orientations and for the disc. Typically, the velocity required varied between 19 and 32 cm s^{-1}, which is in excess of that required to entrain the same particles from the smoother beds. The maximum shear velocity that could be generated was 44 cm s^{-1}, but this was insufficient to break-up the imbricated structures.

Particle Velocities

On the smoothest bed the rod exhibited the fastest particle velocity, followed by the spheroid, ellipse and disc (Fig.4). On the rougher beds, the motion of the spheroid was retarded by the roughness of the bed more so than any of the other shapes (Fig. 5 & 6).Oblate and flattened particles rolled more easily and directly over the roughness elements whilst the spheroid meandered around obstacles. Consequently on the rougher two beds the relative particle speeds were ranked; rod, ellipse, disc and spheroid.

The difference in particle velocities is maintained over the range of current velocities examined. Particle velocity increases as a linear function of current velocity over the range examined; the rate of particle velocity increase being somewhat less than that of the stream current (Table 2).

Entrainment Model

All of the particles, excluding the cuboid, moved by rolling and so this analysis is restricted to a consideration of entrainment

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by pivoting.

In a fuller exposition (Carling & Kelsey,in prep.) geometric relationships are advanced which calculate the grain contact angle and the pivoting angle for any isolated ellipse (2-D case) or ellipsoid (3-D case) positioned over the col between two (or four) bed-particles. The perched grain may be of dissimilar size and shape to the bed-particles and may be in a transverse or parallel orientation. Entrainment may be either by grain-top rotation (2-D case) or by pivoting over a col (3-D case). The geometry is a reasonably exact description of natural bed geometries where clustering of perched particles is not significant but for brevity the full details are not reported here.

Contact angles (θ) and pivoting angles (ϕ) calculated for the 2-D geometry of the test particles and the three bed roughnesses are summarized in Table 3. In addition the roughness parameter (z_p) and the relative degree of exposure (ψ) of the test particle are also tabulated. Because calculation of the pivoting angle for a test particle overlying a bed of particles of dissimilar size may be tedious without a computer routine, it is convenient to summarize typical data graphically. An example is given in Fig.7.

Other parameters can also be presented graphically but for brevity these relationships are not presented here. Appropriate values (as in Table 3) can be selected for insertion in an initial motion function.

Consideration of the usual force balance (Fig 8) acting on the area of an ellipse exposed to the flow gives an expression for the drag force,

$$F_d = \frac{C_d}{2} \varrho \, A \, U^2 \qquad (1)$$

where for the coefficient of drag data used $A = \pi(a_2 b_2 c_2)^{2/3}$

The force due to the mass of the particle is

$$F_g = V \, (\varrho_s - \varrho) \, g \qquad (2)$$

where for an ellipsoid $V = \frac{4}{3} \pi \, a_2 b_2 c_2$

and as

$$F_d = F_g \frac{\sin(\varphi - \beta)}{\cos(\varphi)} \qquad (3)$$

then including the relative exposure term (ψ) and the drag coefficient (C_d) the critical entrainment velocity is given by

$$U_c^2 = \frac{2 \, V \, (\varrho_s - \varrho) \, g}{\psi C_d \varrho \pi (a_2 b_2 c_2)^{2/3}} \frac{\sin(\varphi - \beta)}{\cos(\varphi)} \qquad (4)$$

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As the velocity at height z above the bed is given by

$$U = 5.75 \, u_* \log(30z / k_s)$$ (5)

then the critical shear stress for any shaped ellipse can be given by,

$$\tau_c = \frac{2 \, V \, (\varrho_s - \varrho) \, g}{\psi C_d \, \pi (a_2 b_2 c_2)^{2/3}} \, \frac{\sin(\phi - \beta)}{\cos(\phi)} \, \frac{1}{[5.75 \log(30z_p / k_s)]^2}$$ (6)

The appropriateness of equation 6 can be checked by comparing the calculated threshold shear velocity ($u_* = \sqrt{\tau_c/\varrho}$) and the observed threshold shear velocity. For the transverse rod on roughness 3 the observed threshold shear velocity was 17.29cm s^{-1} (Fig.1). Given $\phi=28.28^{O}$, $\beta=0.57^{O}$, $\psi=0.84$, $(\varrho_s-\varrho)/\varrho=1.65$, g=981cm s^{-2} and z_p=4.53cm, k_s=11.3cm it remains to select a value for C_d. As a critical evaluation is required it would be inappropriate to calculate C_d from the hydraulic test data, but it is expected to be large and so can be set to 1.0 in the first instance. Introduction of the parameters into equation 6 gives a critical entrainment velocity of 17.08cm s^{-1}; a result in excellent agreement with observation. For roughness 2 the calculated result was 12.97cm s^{-1}(observed=10.50cm s^{-1}) and for roughness 1, 9.61cm s^{-1}and 6.92cm s^{-1}respectively.

Discussion

The flume tests demonstrate that, within a limited size range, variation in cobble shape and orientation for a given bed

roughness are as important variables in the initial motion criteria as is the mass of the particle defined by weight or diameter. This variability means that a range of particles of varying size, shape or orientation may be entrained from a bed of mixed grains under identical flow conditions. These *susceptible* particles may be regarded as 'equally mobile' but other particles may not be entrained for the given hydraulic conditions because of their particular size shape or attitude on the bed.

Attention needs to be given to the influence of the complete size distribution and packing of the bed materials in individual rivers as field data (eg. Carling, 1989) show conclusively that where a wide gradation of particle sizes exists, then selective entrainment (defined by size) is an important process such that the probability of encountering larger particles in motion increases as the shear velocity increases. Any robust deterministic initial motion expression must include realistic definitions of the geometry and packing of the particles on the bed such that the mechanics of initial motion can be correctly specified. Further work is in progress and will explore the mechanics of imbrication and initial motion by sliding.

Council to PAC facilitated travel. For all this support we are grateful.

Table 1.   Dimensions of test particles

Shape	A-axis	B-axis	C-axis	Weight
	(mm)	(mm)	(mm)	(kg)
Rod	225	60	60	1
Ellipse	159	100	47	1
Disc	155	137	40	1
Spheroid	122	95	63	1
Sphere	116	116	116	2.4
Cuboid	129	96	75	2.4
UP60	81	60	49	0.35
UP130	159	130	115	3.2

Table 2. *Constants and coefficients for particle velocity (U_p) as function of near-bed velocity (U_b). $U_p = a + bU_b$*

	Roughness 1		Roughness 2		Roughness 3	
	(a)	(b)	(a)	(b)	(a)	(b)
Rod	-43.20	0.66	-41.57	0.73	-43.36	0.70
Ellipse	-69.22	0.76	-42.25	0.69	-55.84	0.75
Disc	-99.68	0.87	-37.10	0.68	-54.79	0.73
Spheroid	-54.16	0.71	-56.93	0.75	-69.56	0.80

Table 3. Calculated entrainment parameters for test particles.

	Roughness	Rod	Ellipse	Disc	Spheroid
	1	20.93	10.20	5.57	13.24
θ (deg.)	2	23.24	10.99	5.81	14.64
	3	28.28	16.05	9.59	21.11
	1	20.93	39.17	48.83	28.14
ϕ (deg.)	2	23.24	41.32	50.06	30.71
	3	28.28	52.82	63.23	41.28
	1	39	34	32	41
z_p (mm)	2	44	36	34	42
	3	45	35	34	39
	1	0.96	0.975	0.985	0.97
ψ (-)	2	0.95	0.96	0.97	0.95
	3	0.84	0.84	0.87	0.83

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Fig. 1
Rough 1 Shape/Orientation v.U*

U* CM S-1

	ROD	ELIP	DISC	S/OID	SPHERE	CUBOID	NP60	NP130
PARALLEL	14.948	18.278	11.715	16.006	9.812	17.736	16.691	16.887
TRANSVERSE	6.92	9.818	9.297	12.068		14.306	8.565	16.238
OBLIQUE	9.488	12.842	10.563	13.031		18.358	13.325	17.08

SHAPE

■ PARALLEL   ▨ TRANSVERSE   ☐ OBLIQUE
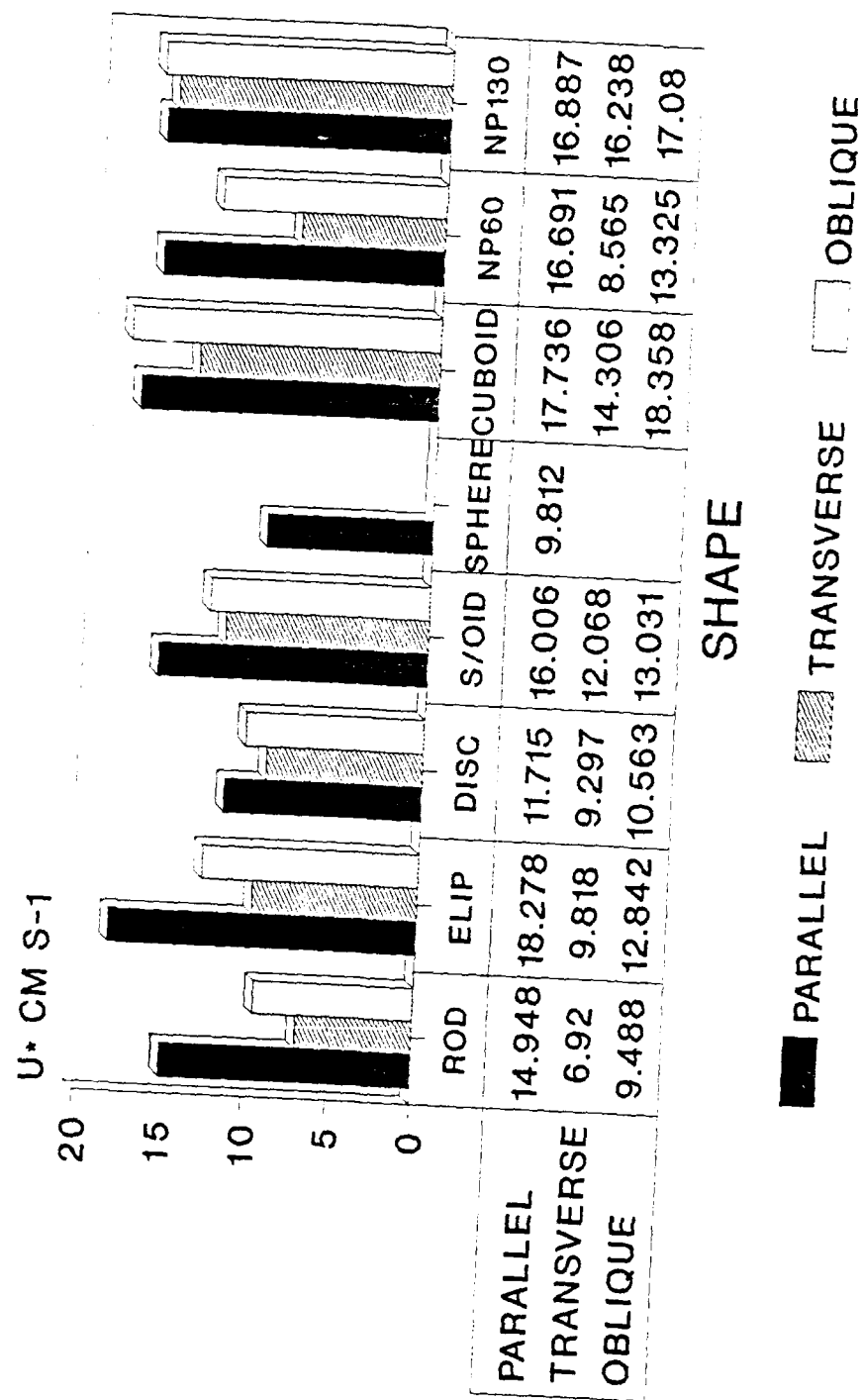
Mean U* (n=5)

Fig. 2

Rough 2. Orientation/Shape v. U*

U* (CM S-1)

	ROD	ELIP	DISC	S/IOD	SPHERE	CUBOID	NP60	NP130
PARALLEL	30.282	22.268	18.09	21.809	20.438	21.627	22.459	22.547
TRANSVERSE	10.505	15.238	12.769	16.783		22.615	13.261	20.193
OBLIQUE	11.519	15.593	12.979	16.338		21.745	15.7	20.08

SHAPE

■ PARALLEL   ▨ TRANSVERSE   ▢ OBLIQUE
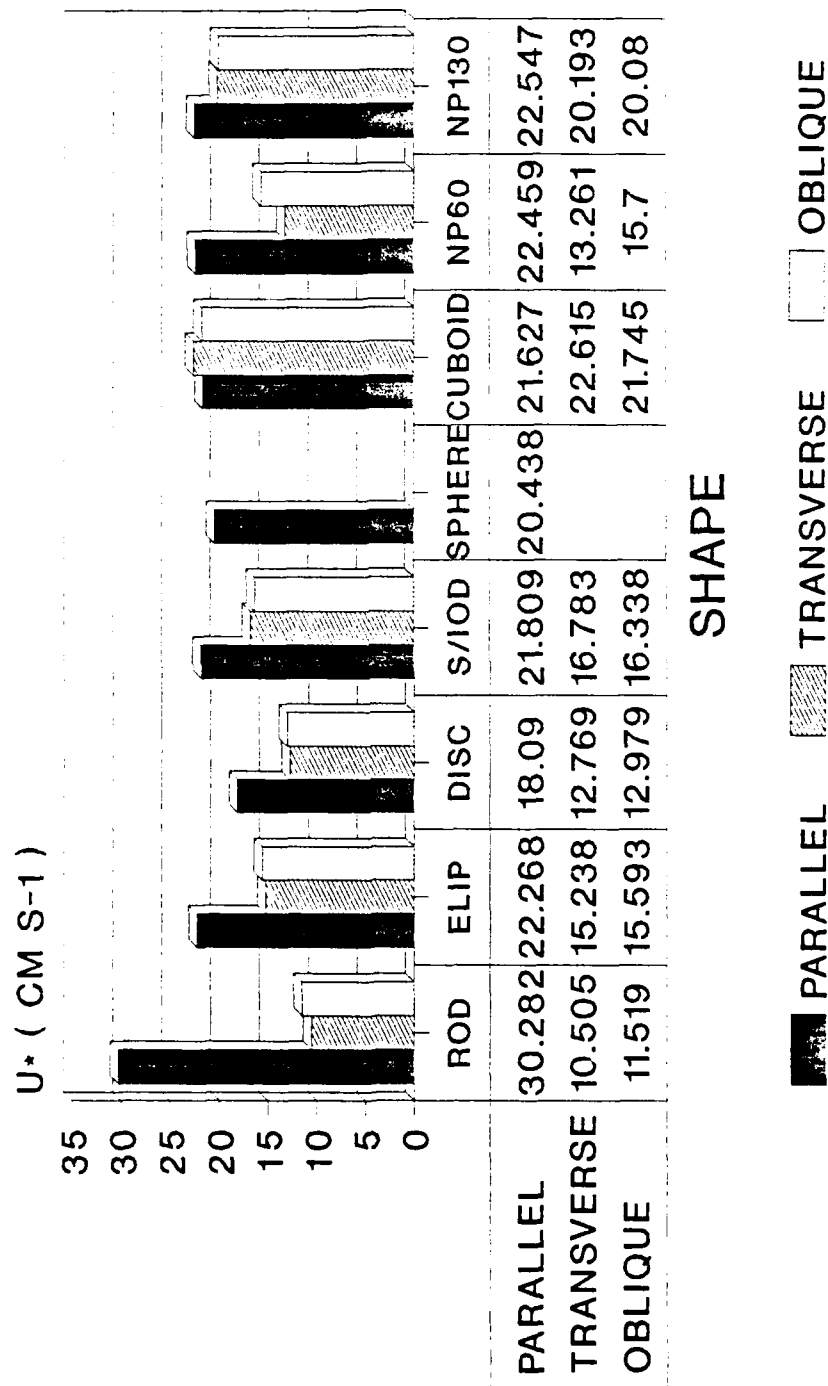
MEAN U* VALUES USED (n=5)

Fig. 3

Rough 3 Shape/Orientation v.U*

U* CM S-1

	ROD	ELIP	DISC	S/OID	SPHERE	CUBOID	NP60	NP130
PARALLEL	36.758			40.561	27.075	34.935	31.571	41.119
TRANSVERSE	17.292	22.675		40.352		35.794	22.169	36.329
OBLIQUE	14.827			27.744		29.723	22.4	39.32

SHAPE

■ PARALLEL   ▨ TRANSVERSE   □ OBLIQUE
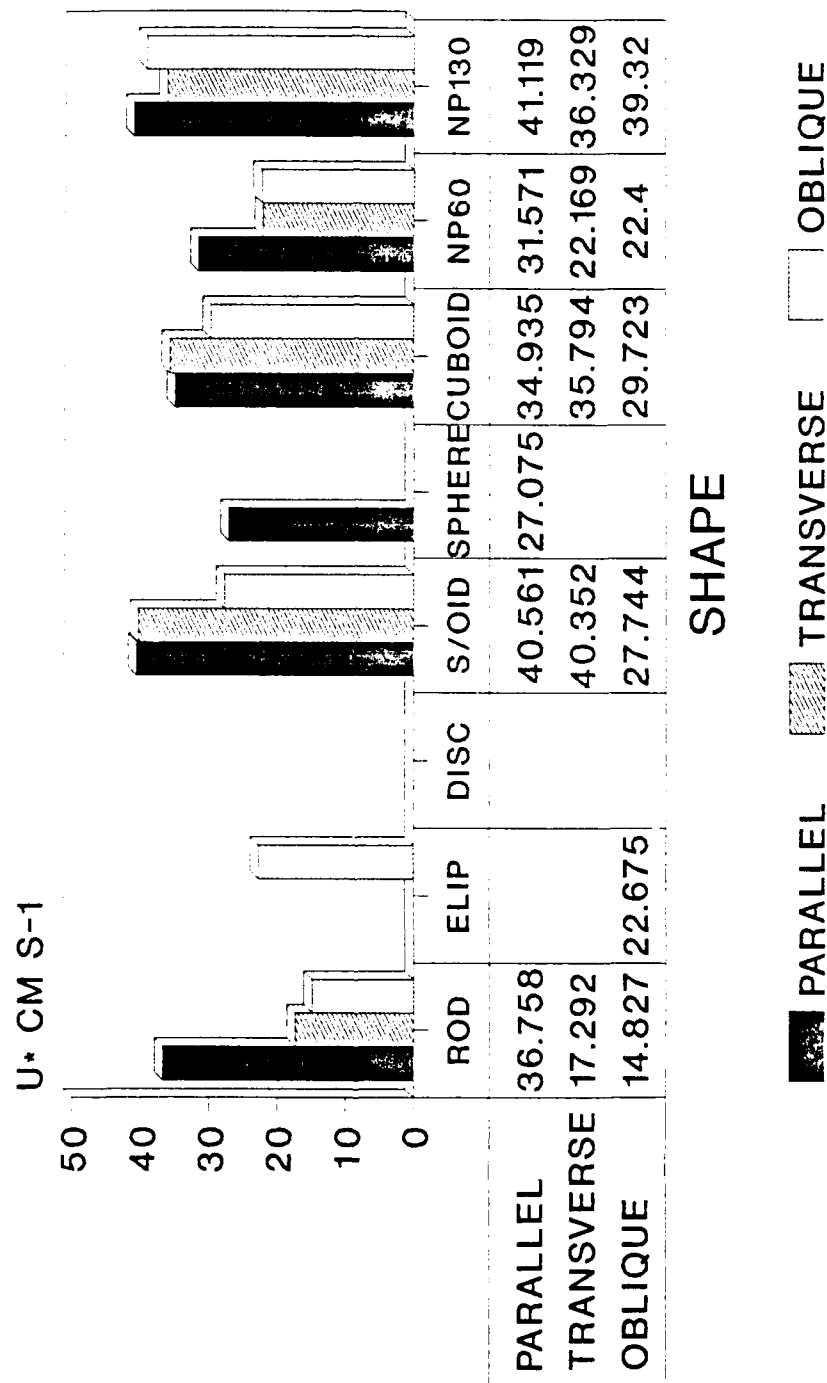
Mean U* (n=5)

Fig. 4. Particle Velocity
Roughness 1

Particle Velocity (cm s-1)

160
140
120
100
80
60
40
20
0

100 110 120 130 140 150 160 170 180 190 200 210 220 230 240 250

Current Velocity (cm s-1)

- - - Rod   —+— Ellipse   —+— Disc   —⊡— Spheroid

Fig. 5. Particle Velocity
Roughness 2

Particle Velocity (cm s−1)

Current Velocity (cm s−1)

Rod   Ellipse   Disc   Spheroid

Fig. 6. Particle Velocity
Roughness 3

Particle Velocity (cm s−1)

Current Velocity (cm s−1)

—— Rod   ·–·– Ellipse   ·–·– Disc   ·–·– Spheroid
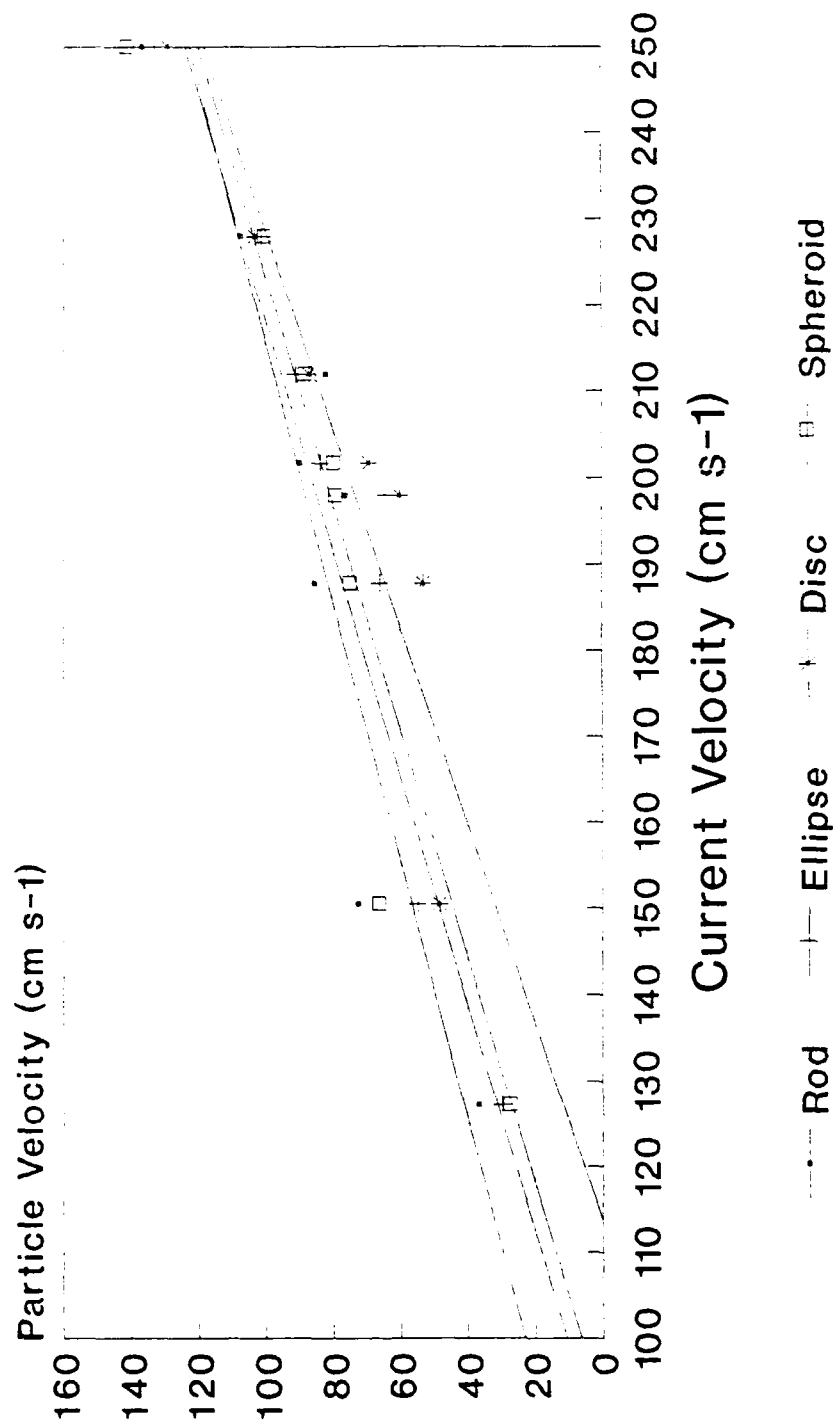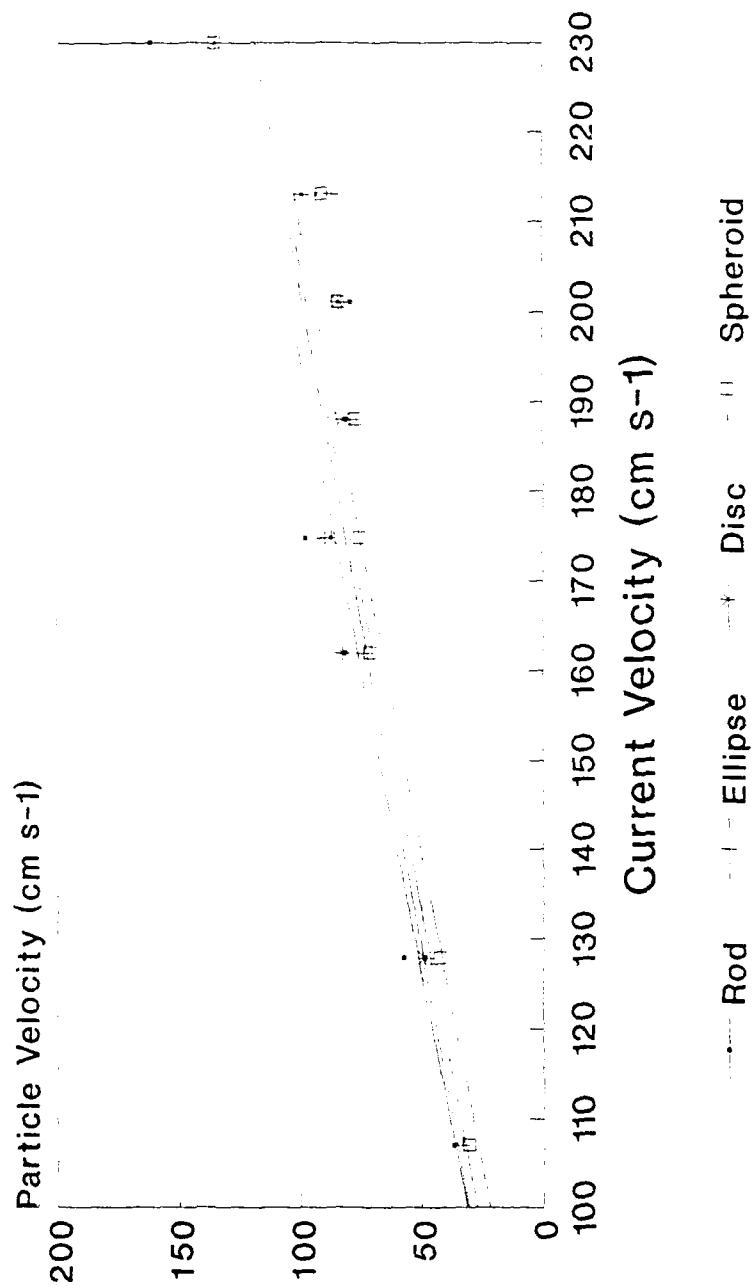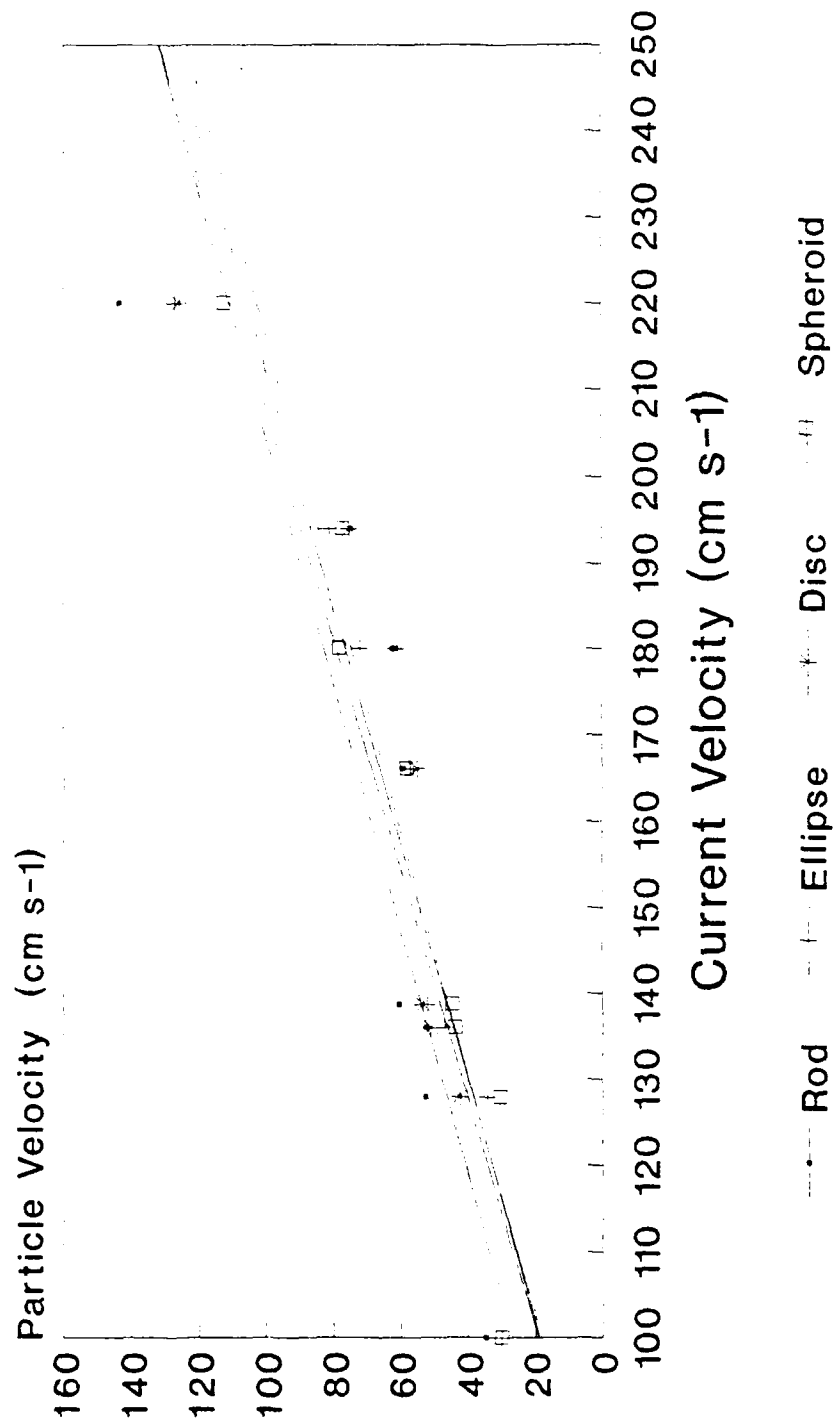
Fig. 7.
Pivot Angle vs. Bed Form
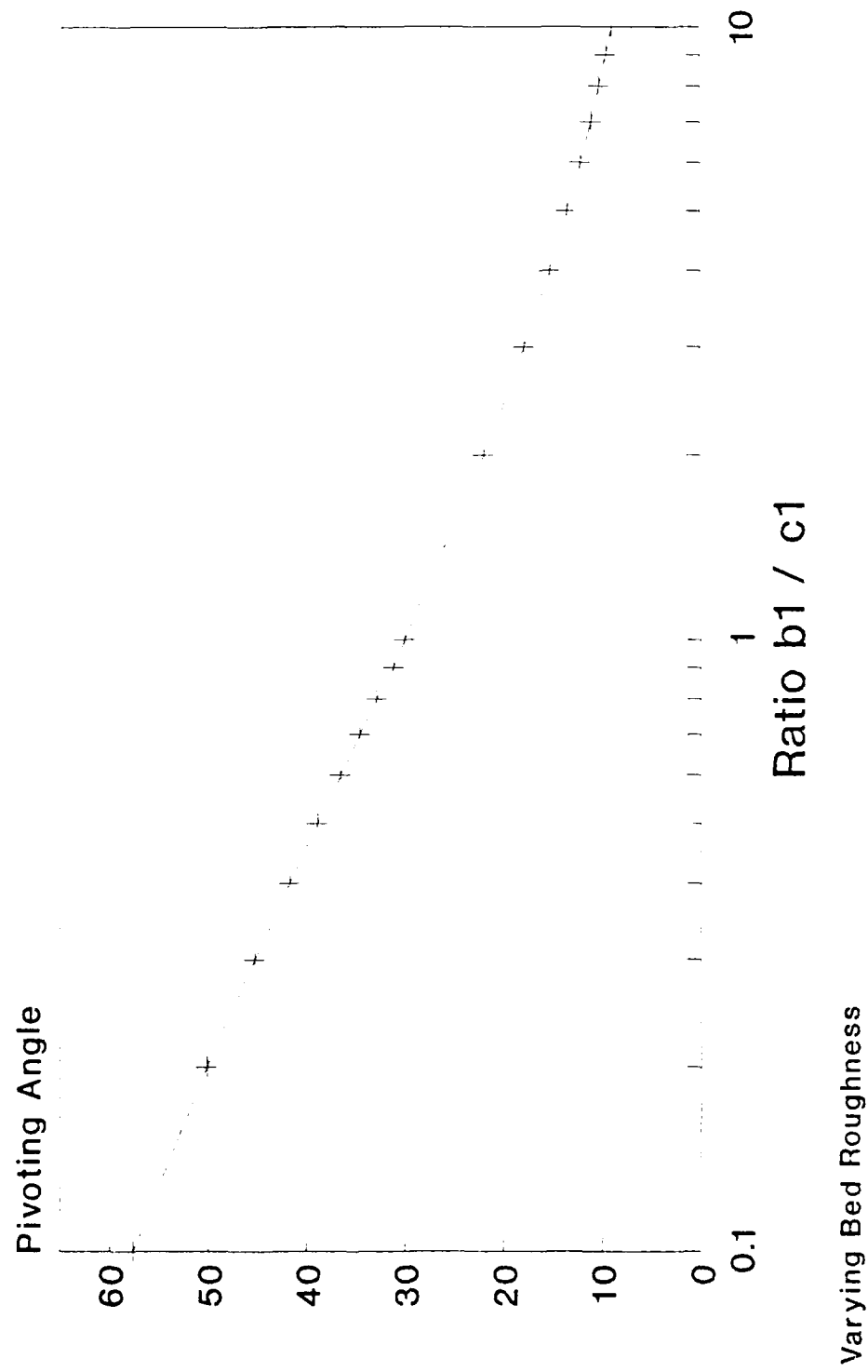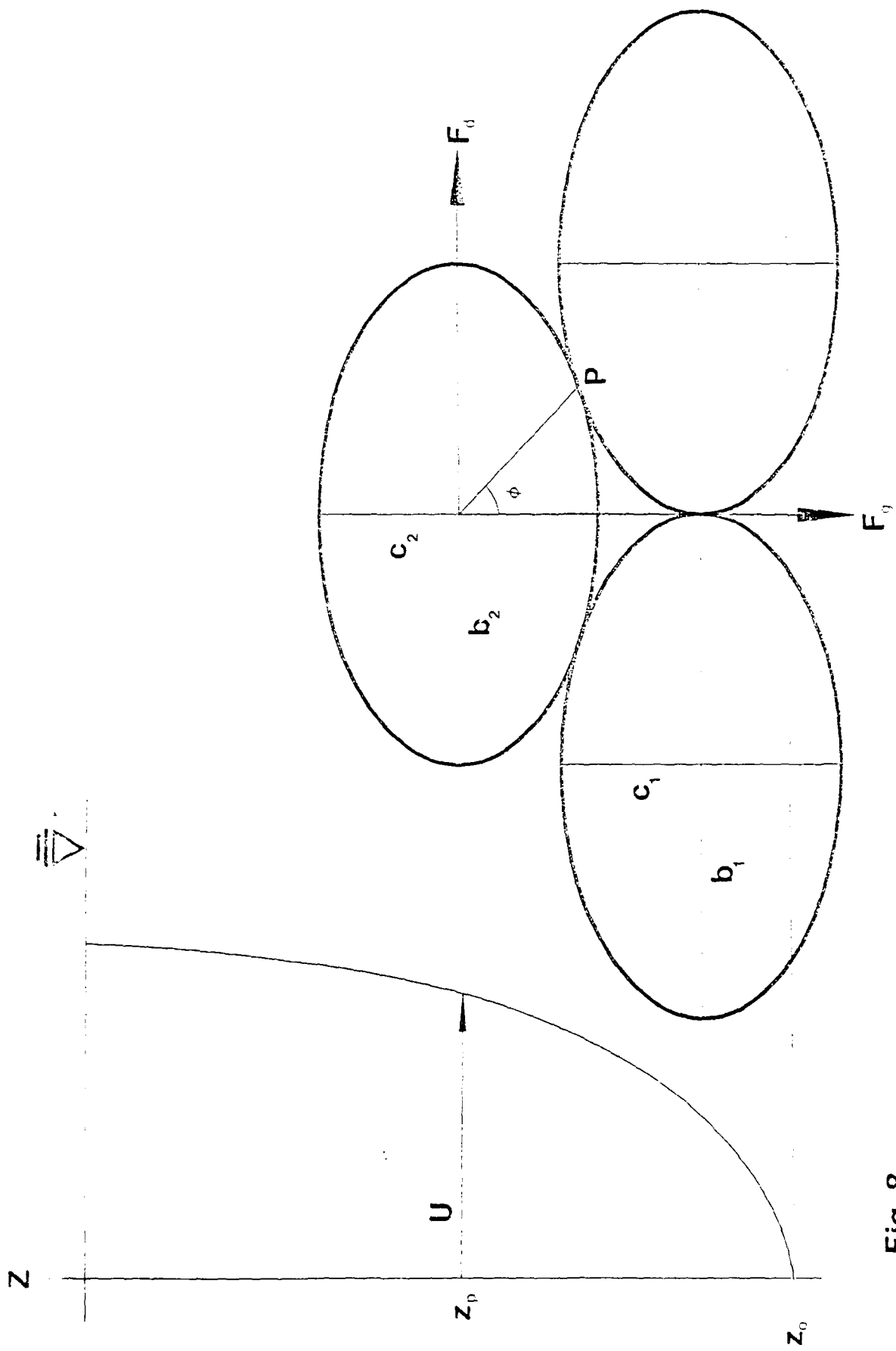
Pivoting Angle

Ratio b1 / c1

Varying Bed Roughness

Fig. 8.

TOPOGRAPHIC EXPRESSION OF BARS IN MEANDERING CHANNELS

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ABSTRACT - Bar morphology in meandering channels has been investigated through an extensive set of controlled experiments in fixed side-walls channels with erodible bottom. Different meander configurations have been tested under the same set of flow conditions. with values of channel wavelength and width to depth ratio falling within the range usually found in nature. Results of experiments may shed some light on the fundamental question concerning the triggering mechanism of bend instability which leads to meandering. It appears that within the range of wavelengths examined the experimental trend is far from being satisfactorily predicted by linear theories which have been so far employed to investigate bed topography in meandering channels.

1 - INTRODUCTION

Bars are prominent features of alluvial channels. They develop as a results of an instability process that arises from the coupling between flow and sediment transport in mobile channels, on a spatial scale typically of the order of channel width. and grow at a rate that is associated with the bed erosion-deposition process. The presence of bars exerts a strong influence on the flow field and sediment transport process; the resulting patterns greatly enhance the erosion of channel banks and consequently induce bend growth and channel shift. Thus, the presence of bars mainly determines the morphology of alluvial channels.

Besides the practical relevance of the subject (the development of bars affects several aspects of fluvial engineering), modelling flow and bed topography associated with bar development has attracted the interest of many researchers since it stands as the essential step towards the understanding of river meandering.

As a consequence of many theoretical works on meandering, supplemented by various field studies (Bridge and Jarvis, 1982; Dietrich and Smith, 1983, 1984, Thorne et al., 1985) and flume investigations (Friedkin, 1945; Hooke, 1975; Gottlieb, 1976; Fujita and Muramoto, 1982), a fairly advanced knowledge of the process has been reached. Based on these analyses, it is now possible to draw a consistent picture of various phenomena involved in the process of meander formation, even though recent applications of nonlinear mechanics to river systems seem to suggest that the state of the subject can by no means considered as fairly settled.

In single-thread alluvial channels two classes of bars are typically identified that arise from both natural and forced responses. Migrating free bars develop spontaneously in straight channels and persist in gently curved channels unless channel curvature exceeds a threshold value above which bar migration is suppressed. Forced (point) bars are nearly steady features that develop in response to the curvature of the channel and give rise in meandering channels to an alternate sequence of shoals on the inside of the bends and pools along the outer banks.

Research developed in the seventies and in the early eighties on flow and bed topography in meandering channels focussed mainly on linear theories of point bars. However further theoretical investigations have shown a clear connection between free and forced bars. Blondeaux and Seminara (1985), within the context of a unified approach, found that the amplitude of point bars in sinuous channels is likely to be controlled by the interaction between curvature and free bars. In particular they showed that the linear response of flow and bed topography to the forcing effect of curvature exhibits a resonant behaviour and peaks when meander wavelength and width to depth ratio of the channel attain values at which curvature forces a natural response consisting of steady free bars. When coupled with a bend stability analysis of the kind proposed by Ikeda et al. (1981), where meandering is explained in terms of a planimetric instability of the channel activated by secondary flow driven by bed topography in sinuous channels, the above theory shows that resonance controls meander initiation and leads to selected meander wavelengths falling in the resonant range.

The picture that emerges from the linear analyses has been recently questioned by Seminara and Tubino (1990) who derived a nonlinear analytical solution valid within the resonant range; preliminar results of the above analysis show that for a relatively wide range of values of meander wavelength and width to depth ratios centered at resonance the linear solution does no longer apply and nonlinear effects lead to a solution which exhibits a sharply different behaviour.

Validation of the above theoretical findings through the existing experimental data is not possible since they refer to different experimental conditions and are not systematic. Thus a further progress in the understanding of meandering initiation and development preliminarly requires a detailed comparison with experimental data obtained through controlled experiments designed so as to cover the range of wavelengths and width to depth ratios close to the resonant values. To this end flume experiments have been carried out in fixed-sidewalls sinuous channels in the laboratory of the Hydraulic Institute of Genoa University. Experimental conditions have been chosen so as to examine bar development as channel wavelength varies within the resonant range, for the same set of flow conditions. Measurements have focussed mainly on bed topography; during experiments bed evolution in each pool has been documented and detailed altimetric surveys have been carried out at the end of each run.

A detailed description of experimental procedure and results is deferred to sections 3 and 4 respectively, and follows a brief review of previous theoretical results that is given in section 2.

Section 5 is devoted to a discussion of results along with some concluding remarks.

2 - THEORETICAL FRAMEWORK

To investigate the development of bed topography in cohesionless sinuous channels a large number of theoretical studies has been developed by various authors (e.g. Engelund, 1974; Gottlieb, 1976; Ikeda et al., 1981; Smith and Mc Lean, 1984; Kitanidis and Kennedy, 1984; Blondeaux and Seminara, 1985; Seminara and Tubino, 1985; Ikeda and Nishimura, 1986; Odgaard, 1986). A detailed discussion of the mathematical framework and the physical basis of these analyses, which are linear and employ both two-dimensional and three-dimensional flow models, can be found in some recent review papers (Parker and Johannesson, 1989; Seminara and Tubino, 1989; Seminara, 1989; Tubino, 1990). However, a

brief description of the main theoretical assumptions and results may help the reader to locate the experimental results reported in this work in the field of the present status of the theoretical research on river meandering.

Since the original work of Langbein and Leopold (1966) the problem has been treated by assuming that the channel axis follows a 'sine-generated curve'. Thus, denoting by R_o^* twice the radius of curvature at the bend apex and by λ_m^* the intrinsic meander wavenumber, the longitudinal distribution of channel curvature can be expressed by

$$\frac{1}{r_o^*} = \frac{1}{R_o^*} \exp[i\lambda_m^* s^*] + c.c. \tag{2.1}$$

where r_o^* is the local radius of curvature of channel axis, c.c. denotes complex conjugate, and a curvilinear orthogonal coordinate system is adopted such that s^* is the intrinsic longitudinal coordinate denoting the arclength and n^* is the transverse coordinate measuring the radial distance from the axis (figure 1).

The mathematical problem, posed by Reynolds equations coupled with sediment continuity equation, can be reduced to a dimensionless form by employing the half width of the channel B^* as the relevant spatial scale of channel topography and referring to a basic uniform flow, with given discharge and slope, such that D_o^* and U_o^* denote the average flow depth and speed respectively. Thus, the solution for the dependent variables (velocity field, flow depth and bed elevation) can be sought in terms of five dimensionless parameters which account for both geometric and hydraulic characteristics:

• the curvature ratio

$$\nu = \frac{B^*}{R_o^*} \tag{2.2}_c$$

• the width ratio

$$\beta = \frac{B^*}{D_o^*} \tag{2.2}_b$$

• the dimensionless meander wavenumber

$$\lambda_m = \lambda_m^* B^* \tag{2.2}_c$$

• the Shields parameter

$$\theta = \frac{\tau_o^*}{(\rho_s - \rho)g d_s^*} \tag{2.2}_d$$

• the roughness parameter

$$d_s = \frac{d_s^*}{D_o^*} \tag{2.2}_e$$

where τ_o^* is bed shear stress associated with the reference uniform flow, ρ_s and ρ are sediment and water density respectively, g is gravitational acceleration, and d_s^* is the diameter of the sediment (modelled as uniform).

Since the timescale of the process under investigation is that associated with bed erosion, which is typically $10^3 - 10^4$ larger than the flow timescale, a quasi-steady approach can be used, whereby the flow field is assumed to adapt instantaneously to changes in the bottom configuration. Moreover, taking advantage of commonly observed features of natural streams, the mathematical problem can be simplified assuming the channel to be wide and its curvature to be small.

The first assumption, namely $\beta \gg 1$, allows one to analyse the flow only in the 'central' region of the channel neglecting the effect of sidewall boundary layers. It also suggests that a two-dimensional depth-averaged flow model can be suitable enough to describe the main features of solution, provided the dispersive effects associated with the helical component of secondary flow, which vanishes when performing the process of depth-averaging, are somewhat accounte for in the model. Following the lead of Engelund (1974), Seminara and Tubino (1989) have incorporated these dispersive effects into a two-dimensional model by decomposing the transverse secondary flow associated with curvature into a helical component with no net flux and a depth-averaged component, and feeding into the 2-D equations for the latter component dispersive (redistribution) terms obtained by the fully three-dimensional solution for the helical component.

The second assumption, namely $\nu \ll 1$, applies conveniently to the stage of incipient meandering and suggests the opportunity to investigate the flow field and bed topography in sinuous channels by means of a suitable perturbation expansion in terms of the curvature ratio ν. Thus, a linear solution of the problem can be determined in the form

$$U = 1 + \nu U_1(n)e_1(s) + c.c. \qquad e_1 = \exp[i\lambda_m s] \tag{2.3}_{c, c}$$

where U is the dimensionless depth-averaged component of longitudinal velocity, scaled by the average speed U_o^*, the longitudinal and transverse coordinate s and n are scaled by the half width B^*, and similar expansions hold for the other variables.

As shown by Blondeaux and Seminara (1985) the solution of the linear depth-averaged problem can be obtained in a closed form without requiring further assumptions on the transverse structure of solution (Engelund, 1974; Ikeda et al., 1981; Odgaard, 1986). A comparison between two- and three-dimensional predictions shows that both models lead to results of comparable accuracy (Seminara and Tubino, 1985, 1989).

The linear model for flow and bed topography in sinuous channels can be incorporated into a bend instability analysis of meander formation following a procedure originally proposed by Ikeda et al. (1981). The procedure is based on the assumption that, bank erosion being a process acting at a larger timescale than that associated with bed development, the steady solution obtained in sinuous channels can be used to build up a model of bend growth in term of the amplification in time of a small-amplitude sinusoidal perturbation of the channel axis. Assuming the original straight channel to be in equilibrium, a perturbation of the channel axis can induce a flow perturbation in the near-bank region able to produce bank erosion (accretion), i.e. bend growth (decay).

Thus, the preferred meander wavenumber selected by bend instability is determined, for given values of θ, d_s and β, from the value of λ_m where the near-bank flow perturbation reaches its peak. As shown by Blondeaux and Seminara (1985), the linear solution exhibits a sharp peak in a fairly wide range of values of wavenumber and width ratio, thus showing a resonant (or quasi-resonant) response of flow and bed topography (figure 2). The peak in the linear response occurs whenever β and λ_m fall in a range close to some resonant values (β_R, λ_R), for a given set of flow and sediment parameters; whitin this range the wavenumber of the forcing curvature is close to the wavenumber of free bars which are steady and not freely growing either in space or in time. Thus curvature provides the forcing for a particular set of natural responses consisting of steady alternate bars.

The resonant behaviour of the linear solution, missed by previous theoretical models where the full coupling between the flow, sediment transport and bed topography was not retained, has been further confirmed within the context of a three-dimensional model (Seminara and Tubino, 1985) and recently recovered also by Johannesson and Parker (1989). Resonance implies that meander wavenumbers selected by bend instability fall within the resonant range, where the bend growth maximizes. When compared with meander wavenumbers observed both in the laboratory and in the field the above prediction appears

fairly satisfactory (Blondeaux and Seminara, 1985, figure 8; Johannesson and Parker, 1989, figure 7). Moreover, for given values of θ and d_s, the resonant values β_R fall in the range physically sensible for meander formation.

However, further theoretical and experimental results suggest that a direct experimental evidence of resonance in sinuous channels in the form predicted by linear models can hardly be expected.

Unless channel curvature exceeds a threshold value, migrating alternate bars may coexist and interact nonlinearly with steady point bars forced by curvature, as observed by Kinoshita and Miwa (1974) and explained theoretically by Tubino and Seminara (1990). The above coexistence, which clearly emerges both in fixed-sidewalls experiments (Gottlieb, 1976) and in free-meandering experiments (Fujita and Muramoto, 1982), is neglected in all the existing linear models, but is responsible for some nonlinear effects that affect the bed topography in sinuous channels (see Seminara and Tubino, 1989). Notice that meandering initiation is invariably found to operate in a channel where migrating free bars have already formed and undergone a finite-amplitude development.

Moreover the linear expansion breaks down when the resonant conditions are exactly met, the solution exhibiting an infinite peak. A recent attempt to remove the above singularity has been introduced by Seminara and Tubino (1990) by means of a nonlinear analysis developed in the range of λ_m and β close to resonant conditions. The above theory shows that within this range nonlinear effects are able to smooth out the infinite peak of the linear solution and shift the maximum response (i.e. the maximum bend growth) towards larger values of λ_m (figure 2). It also appears that as the curvature ratio ν increases, the solution is largely controlled by nonlinear effects in a quite wide range of values of λ_m and β.

Then, for larger values of the curvature ratio ν, when migrating bar-perturbations are suppressed, linear models are no longer significant and a nonlinear analysis is needed to properly describe the bed topography in sinuous channels for values of λ_m and β which cover the range of values usually found in natural meanders. In particular, within this range the correct order of magnitude of solution is $O(\nu^{1/3})$ rather than $O(\nu)$ as in a linear theory. This implies that, even for small values of the curvature ratio ν, the contribution of higher harmonics of bed topography can affect the solution in a non-negligible way and lead to the reproduction of distinctive asymmetries that characterize bed topography in natural meanders. This also implies that, within the above range, the finite amplitude evolution of meander bends is likely to be governed by higher harmonics of flow and bed topography induced by flow nonlinearities.

3 - METHODS

The experiments were carried out on a large indoor platform measuring 29.6 m long and 2.3 m wide. The concrete platform had a longitudinal slope of approximately 0.006 and was overlain by a 2 cm thick veneer of painted wood. Along both sides of the platform was an approximately 0.40 m high railing that supported a carriage used for leveling the bed and measuring bed and water surface topography. The railing had a constant slope of 0.006 that was carefully checked every 1.5 m.

The 0.35 m wide sinuous flume was constructed by attaching 2 m long and 0.15 m high strips of clear polycarbonate to the wooden platform base. Three or four right-angle braces of bent polycarbonate were glued to the 2 m strips. Screws anchored the base of the braces to the platform forming vertical flume walls. The channel centerline followed a sine curve. The trace of each bank in the series of bends was established by plotting on oversize paper at a 1:1 scale. This paper was tacked to the wooden platform and the banks positioned to follow this trace. Once the banks were correctly located the paper was removed and the polycarbonate sealed to the platform with silicon caulk. The caulk gave additional rigidity to the banks. Once the channel was laid out, local variations in width associated with the

banks not standing at right angles to the platform were eliminated by placing additional braces along the banks. The platform was long enough to allow the modeling of more than seven consecutive bends also for the configuration with maximum wavelength.

The flume was filled with a well sorted 0.76 mm quartz sand. The D_{84} of the sediment was 0.87 mm and the D_{16} was 0.65 mm. These percentages correspond to the grain diameter such that 84 or 16 percent were finer.

Flow entered the flume through a 0.35 cm wide notch in the headtank at the upstream end of the flume. A fine mesh screen served as a baffle to surface waves that might have been engendered in the head tank. The inlet pipe's submergence 0.20 m under the free surface minimized disturbance. Two narrow vertical sills, one 0.10 m and one 0.30 m into the flume, prevented inlet scour and thus maintained a constant inlet elevation.

At the downstream end of the flume a tailgate was constructed to maintain the outlet elevation. The tailgate was a plastic strip whose ends were recessed into a wooden enclosure. A ramp with glued-on sand, partially buried in the sand, met the tailgate at the height of the tailgate. This ramp was designed to prevent scour associated with the vertical sill. Above the sill setting the tail elevation a second plate could be lowered to set the height of the free surface. Flow spilled through a tipped trough into the weir box. The weir box was 2.50 m long and .70 m wide and was partitioned into two sections. In the first section, a mesh bag for collecting sediment was placed. The two sections were connected through a slot at the base of the partition. The downstream end of the box had a 90° weir plate. Flow spilled out of the weir box into a large reservoir. A pump submerged in the reservoir returned water to the headtank. A valve along the pipe connecting the pump to the head gate could be turned to vary discharge.

Sediment was fed into the upstream end of the flume with a sediment shaker. The sediment flux out of the shaker was difficult to calibrate so was monitored frequently and adjusted to give the appropriate feed rate set equal to the observed average tailgate flux. The sediment discharge at the end of the flume was measured by collecting the sand in a 0.40 m wide, 0.60 m long and 0.35 m deep mesh draped frame submerged in the weir box. The submerged collector was suspended from an overhead beam and attached to a strain-gage. The cumulative submerged weight was noted every 5 to 10 minutes during the experiment.

The bed of the channel was prepared by using 0.34 m wide scraper that was attached to a carriage that ran along the rails. The bed was smoothed flat to a constant slope of 0.0060 before each experiment. A very low discarge was passed over the bed prior to the experiments to prepare a smooth saturated surface. The bed elevation at 1.0 meter intervals was measured with a point gage attached to the carriage. Once flow was begun the required discharge was achieved by slight adjustments to the valve. The flow discharge was determined with a point gage attached to the weir box.

During experiments the position and maximum depth in each pool was noted periodically and the sediment water interface at the concave bank was traced to document initial bed evolution. After the stable discharge was reached early in the experiment, water surface topography was measured in order to calculate the average depth unperturbed by developing topography. Every 45 minutes later water surface elevation was remeasured. After the pools and bars had become stabilized in position and once water surface slope and discharge were average on the long term indicating equilibrium conditions, the experiment was terminated. This generally took 3-4 hours.

The bed topography was preserved at the end of the experiment as flow waned by carefully placing damp newspapers on bar tops near steep slopes. This prevented dissection of the sideslopes and consequent filling of the pools. The bed topography was measured with a laser scanning device mounted on the carriage. The precision of the scanner was 0.01 mm. The scanner was driven by a stepper motor which stopped every 2 mm across the channel for a measurement. The system was run of an Olivetti M21 computer. The data was later

converted into elevation data for presentation. Traverses of the topography were made every 0.10 to 0.20 m along the channel depending on the channel wavelength; for the configurations studied this corresponds to 36-44 cross-sectional surveys per channel wavelength. Typically two to two-and-a-half wavelengths were scanned. The portion of the channel studied began at least 8 m downstream of the inlet and ended at least 7 m from the flume exit.

4 - RESULTS

To simplify as much as possible the drawing of the channels and the data flow between the intrinsic curvilinear coordinate system (s^*, n^*) and the Cartesian system (x^*, y^*) (see figure 1), we decided to have a purely sinusoidal channel axis given, in nondimensional form, by

$$y = a \sin(\lambda x) \tag{4.1}$$

where x, y and the amplitude a are scaled by the half width B^* and λ denotes the Cartesian wavenumber, normalized by $1/B^*$. The curvature ratio ν retains exactly the same meaning as before while the longitudinal distribution of curvature differs from (2.1) of terms of order $O(a^2\lambda^2)$.

As was outlined in section 2, the number of parameters present in the problem is high, so that many experiments would be needed to clarify the influence of each variable on the phenomenon. We therefore decided to keep the value of ν constant through all the experiments while varying conversely channel wavenumber λ and width ratio β in a range as broad as possible. As mentioned before the slope of the channel and the sand diameter were fixed once and for all to be respectively 0.006 and 0.76 mm.

The choice of ν results from a compromise between the requirement of a small ν due to theory validation purposes and space limitations, and the need for having a relatively high value of ν to minimize the presence of propagating bedforms. On the basis of theoretical estimates of the threshold conditions for migrating bars suppression, the value of ν was therefore chosen to be 0.05; this corresponds to a channel curvature characteristc of natural meanders in the initial-intermediate stage of evolution.

Four different channels were built up with four values of λ and the same width (35 cm). The selected range of wavenumbers (0.15 ÷ 0.3) comprises the resonant range ($\lambda \sim 0.15$ ÷ 0.2 accordingly with a linear theory) and extends towards higher values of λ which are better representative of natural situations. Once λ was fixed, the amplitude a of each channel was determined through the relationship

$$\nu = \frac{a\lambda^2}{2} \tag{4.2}$$

In table 1 the geometric characteristics of the channels are shown (L^* and a^* are the dimensional Cartesian wavelength and amplitude of channel axis respectively).

Four to five experiments were performed in each channel at different flow rates that were chosen so as to produce almost the same set of values of average flow depth (D_o^*) for each configuration. Working in this way we were able to analyse the behaviour of each measured quantity both as a function of β (at λ = const) and of λ (at β = const). An exact reproduction of the same values of β for each experiment was not possible since flow depth was indirectly obtained by setting a prescribed flow discharge; this led to minor variations in the average value of flow depth that resulted in larger differences in the value of β. Thus a small work of interpolation was needed to obtain iso-β curves. It might be useful for the reader to stress here the fact that on these iso-β curves Shields stress θ and roughness parameter d_s are constant too.

Table 1: Geometrical characteristics of channels.

Channel	B* (m)	L* (m)	a* (m)	λ	a	ν
W 15	.175	7.33	0.77	0.15	4.44	0.05
W 20	.175	5.49	0.43	0.20	2.5	0.05
W 25	.175	4.4	0.28	0.25	1.6	0.05
W 30	.175	3.66	0.19	0.30	1.11	0.05

Table 2: Flow and sediment transport parameters.

Run	Q_o^* (l/s)	D_o^* (cm)	β	d_s	θ	Fr
W15Q07	0.76	0.89	19.66	0.085	0.043	0.82
W15Q10	1.01	1.01	17.33	0.075	0.048	0.91
W15Q14	1.42	1.36	12.87	0.056	0.065	0.82
W15Q16	1.56	1.51	11.59	0.050	0.072	0.77
W20Q07	0.76	0.92	19.02	0.083	0.044	0.79
W20Q10	1.01	1.13	15.49	0.067	0.054	0.77
W20Q12	1.23	1.19	14.71	0.064	0.057	0.86
W20Q14	1.42	1.34	13.06	0.057	0.064	0.84
W20Q16	1.56	1.50	11.67	0.051	0.072	0.77
W25Q07	0.76	0.82	21.34	0.093	0.039	0.93
W25Q10	1.01	1.11	15.77	0.068	0.053	0.79
W25Q12	1.21	1.21	14.46	0.063	0.058	0.83
W25Q14	1.39	1.40	12.50	0.054	0.067	0.77
W25Q16	1.56	1.52	11.51	0.050	0.073	0.76
W30Q07	0.76	0.94	18.62	0.081	0.045	0.76
W30Q10	1.01	1.05	16.67	0.072	0.050	0.86
W30Q12	1.20	1.19	14.71	0.064	0.057	0.84
W30Q14	1.42	1.39	12.59	0.055	0.066	0.79
W30Q16	1.56	1.51	11.59	0.050	0.072	0.77

Hydraulic conditions and flow and sediment transport parameters of the experimental runs are summarized in table 2 where Q_o^* and Fr are flow discharge and Froude number respectively.

Flow and sediment characteristics were chosen so as to obtain relatively high values of width ratio, as is typical of natural conditions, and to prevent the occurrence of small-scale sand waves as ripples or dunes. In spite of the relative smallness of Shields stress θ (always less than 0.075) sediment transport was indeed appreciable in all the experiments, the critical value of Shields stress being around 0.035 for the values of particle Reynolds number characteristic of the experiments.

The bed time development was almost similar for all the experiments. After approximately 45 minutes a scour hole appeared, tipically located downstream of the bend apexes. While deepening the scour was seen to migrate downstream or upstream depending on experimental conditions until an equlibrium configuration was reached which remained reasonably unchanged until the end of the experiment. This regime topography was fairly stable with the only exception of small propagating sand waves that were often seen to born within the scour hole and vanished at the downstream end of the pool.

Runs at $\lambda = 0.15$ showed the existence of more pronounced migrating alternate bars, clearly detectable in the straight reaches between two bends, whose amplitude was however smaller than that of fixed bars. It seems that the steady scour hole acts as a source and sink for bar-like disturbances. When the straight reaches between two consecutive bends are long enough, as in the case of runs at $\lambda = 0.15$, bar perturbations emerging out of the apex pools are able to develop finite amplitude fronts which spread on the cross section and form conjugate (alternate) fronts. These fronts propagate in the straight reaches between bends and then interact with steady pools when reaching the bend apexes. The latter interaction strongly modifies the alternate pattern; fronts are considerably slowed down while sediment transport concentrates within the pools along the outer bank forming narrow fronts which are stalled and then sped up again when coming out of the pools.

In run W15Q16 free perturbations showed a relatively large amplitude and were often seen to propagate through the bend keeping a distinct alternate pattern. The characteristic wavenumber of these alternate bars was remarkably higher, namely three times larger than that of the meandering channel.

To grasp the main features of bar topography which emerges from the experiments we considered two measured quantities to be representative of each experiments: the former is the value η of the maximum scour with respect to the average (initial) bed, made non-dimensional with the uniform flow depth D_o^*, the latter is the position of the scour hole with respect to the bend apex. Both quantities were averaged over the series of bends within the scanned reach. Position of scour hole is given in terms of the dimensionless quantity $\delta = \Delta x^*/(L^*/2)$, where Δx^* is the longitudinal distance of the scour hole from the bend apex, so that δ ranges from -0.5 (scour hole L/4 upstream of bend apex) to 0.5 (scour hole L/4 downstream of bend apex).

Experimental findings are shown in figures 3_{a-b} and 4_{a-b} where η and δ are given as functions of channel wavenumber λ and width ratio β. It may be noticed from figure 3_b that η exhibits a relative maximum that moves towards larger values of λ as β increases. Moreover figures 4_{a-b} show a fairly regular behaviour for δ which is decreasing with β (the scour hole moving upstream as discharge decreases) and increasing with λ (the scour hole moving downstream as wavelength decreases).

The maximum scour η and phase lag δ are given in table 3 along with the values of the difference between maximum a minimum bed elevation within a wavelength (H_b) and the maximum flow depth within the pools (D), both scaled by the uniform flow depth D_o^* and again averaged over the scanned reach. Notice that the maximum scour contributes to about 3/4 of the total difference between maximum and minimum bed elevation, the ratio η/H_b being fairly constant in the experiments and falling in the range 0.71÷.80.

The output of the scanner device is a large ($\Delta x^* = 10 \div 20$ cm, $\Delta y^* = 2$ mm) cartesian arrays of bed elevations with respect to some constant slope level. The data sets were then moved to the (s^*, n^*) space where a linear interpolation was performed on a regular grid. Back to the (x^*, y^*) space this means that all nodes now lie on sections normal to the channel axis. In figure 5 the measured bed topography for run W25Q14 is shown. Contours are plotted every 0.5 cm. Thinner lines refer to pools while thicker lines refer to shallow areas. The topography shows an almost regular pattern. Both relative position and amplitude of the scours are fairly constant along the channel.

Table 3: Experimental results.

Run	λ	β	η	H_b	D	δ
W15Q07	0.15	19.66	3.34	4.42	5.36	-0.25
W15Q10	0.15	17.33	3.08	4.34	4.97	-0.23
W15Q14	0.15	12.87	2.71	3.59	4.06	-0.09
W15Q16	0.15	11.59	-	-	3.72	-0.08
W20Q07	0.20	19.02	3.66	4.74	5.16	-0.10
W20Q10	0.20	15.49	3.73	4.74	5.00	-0.01
W20Q12	0.20	14.71	3.41	4.40	4.55	0.02
W20Q14	0.20	13.06	3.34	4.33	4.21	0.06
W20Q16	0.20	11.67	2.78	3.72	3.97	0.09
W25Q07	0.25	21.34	4.41	5.49	6.54	0.13
W25Q10	0.25	15.77	3.58	4.54	4.77	0.15
W25Q12	0.25	14.46	3.28	4.26	4.74	0.18
W25Q14	0.25	12.50	2.91	3.90	4.06	0.19
W25Q16	0.25	11.51	2.69	3.61	3.77	0.19
W30Q07	0.30	18.62	3.69	4.71	5.32	0.14
W30Q10	0.30	16.67	3.77	4.94	5.22	0.24
W30Q12	0.30	14.71	3.16	4.12	4.53	0.23
W30Q14	0.30	12.59	2.96	3.91	4.27	0.25
W30Q16	0.30	11.59	2.61	3.55	3.86	0.27

The results were then averaged superimposing all the bend apexes (mirrored when needed). This procedure leads to a one-meander averaged bed topography in which the small differences between different bends are filtered out together with small-scale disturbances arising from propagating bedforms.

The procedure of interpolation and averaging tends to smooth artificially bed profiles, even though the fine spatial-grid used ensures for small damping. Whenever the measurements were seen to be sensitive to this post processing, like in the determination of the position of the maximum scour hole, a local extra measurement was made to increase accuracy.

In figures 6_{a-d} four averaged bed topographies are plotted for the runs W20Q07, W20Q16, W30Q07 and W30Q16. As can be easily recognized, the scour hole moves downstream for increasing discharges (Q07 $\rightarrow$ Q10) and increasing λ (W20 $\rightarrow$ W30). The scour always extends on a smaller area than deposition does, being, on the other hand, of larger amplitude.

Figure 7 shows the same plot for the run W15Q10 which exhibits a double scour hole, along the outer bank, a tipical feature of all runs at the smallest wavenumber (i.e. longest wavelength and amplitude of the channel axis).

The last operation performed on the bottom data was a 2D Fourier analysis that allowed us to extract some information on the structure of the topography in the Fourier space, to identify the most relevant harmonics and their relative importance. To perform the Fourier analysis data were mirrored along one of the banks because the fundamental harmonic in the transverse direction is twice the width of the channel.

In figure 8 a spectrum of the amplitudes of bed harmonics as a percentage of the total is shown. Not surprisingly the spectrum exhibits a peak on the 1-1 harmonic which is by far the most important and, together with 2-2 and 0-2 harmonics, gains one half of the total "energy". The first index refers to the harmonics in the s direction (1 means the fundamental, 2 the harmonic with half the wavelength of the fundamental and so on) while the second

index refers to the n direction so that, for example, the 0-2 har ...onic is constant along the longitudinal direction and has a wavelength in the transverse direction equal to the width of the channel.

In figure 9 the behaviour of the fundamental harmonic η_{11} as a function of λ is shown with β as a parameter. As the width ratio increases the role of the fundamental becomes less important, this effect being more evident at lower values of channel wavenumber.

5 - CONCLUDING REMARKS

Experimental data have been obtained through detailed investigations of bar topography in sinuous channels characterized by different wavelengths, for given flow conditions. The steady topography obtained at the end of experiments shows a fairly regular pattern along the channel in all runs except for those run with maximum wavelength ($\lambda = 0.15$) where detection of steady topography has been somewhat complicated by the presence of propagating bars superimposed on steady point bars.

However, within the range of values of channel wavelength examined in the experiments, both the intensity of maximum scour and the position of scour holes with respect to bend apex appear to follow a consistent trend. The picture that emerges from experimental data allows one to draw some conclusive answers on the suitability of theoretical models which have been employed to predict steady bar topography in meandering channels. In particular, results of experiments strongly support the idea that within the range of wavelength characteristic of natural meanders the prediction of bed topography based on a linear solution does no longer apply since it appears that resonance does not operate in the form predicted by a linear analysis. Rather than exhibiting a sharp peak within the resonant range, the bed response is found to follow a more smoothed trend, which still exhibits a maximum for values of meander wavenumber which are typically larger than those predicted by linear theory. The maximum response shifts towards higher wavenumbers as channel width ratio increases.

A quantitative comparison between theoretical predictions and experimental data is stil in progress. However it is of interest to stress here that preliminar theoretical findings derived within the context of a nonlinear resonant solution (Seminara and Tubino, 1990) appear to be qualitatively confirmed by these data. A glance at figure 2 shows that the nonlinear solution seems to pick up at least the overall trend that emerges from laboratory investigations and predicts a maximum response within the range found in the experiments. Moreover downstream shift of scour holes with increasing meander wavenumber and flow discharge is also reproduced by the theoretical solution, which however seems to overestimate experimental findings.

A further point deserves some additional discussion. A bend stability theory predicts maximum bend growth when both β and λ are close to some resonant values. Results of the detailed investigations on free-meandering made by Fujita and Muramoto (1982) indirectly support these theoretical findings since a clear tendency to meandering was detected in the experiments when channel widening has proceeded enough for the width ratio to have reached a nearly constant value falling within the resonant range. Results of present experiments do not allow to ascertain whether the bed response exhibits a maximum as β increases, since in all the experiments different values of width ratio have been examined by varying the average flow depth, thus leading to consequent variations of Shields and roughness parameters. Therefore. a complete answer on the behaviour of bar topography within the resonant range will require a further set of experiments properly designed so as to examine separately the effect of width ratio variations.

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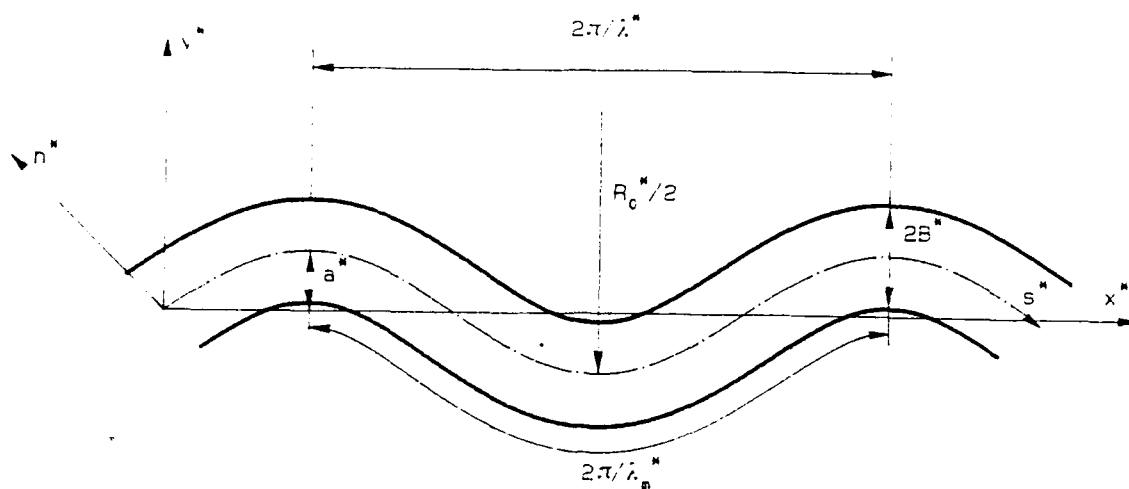
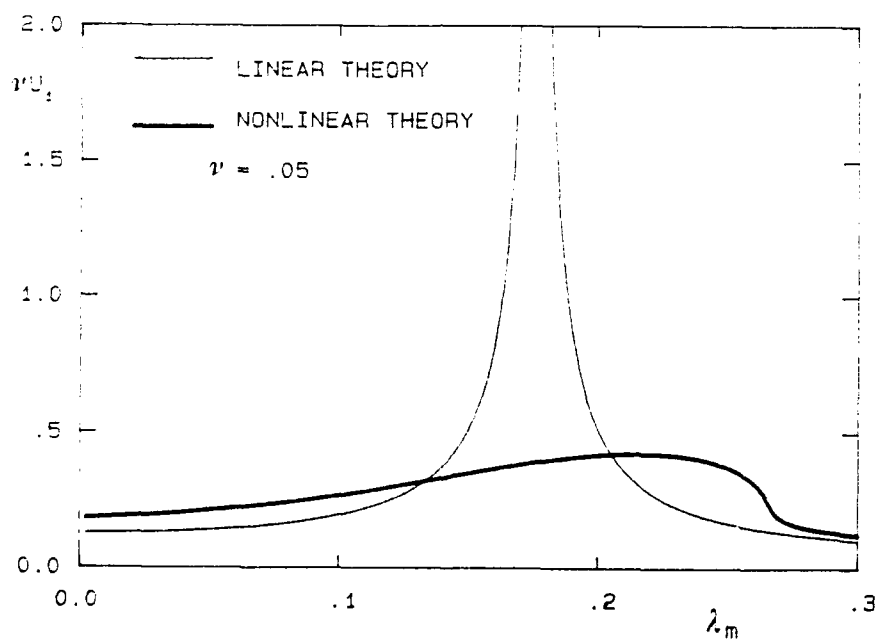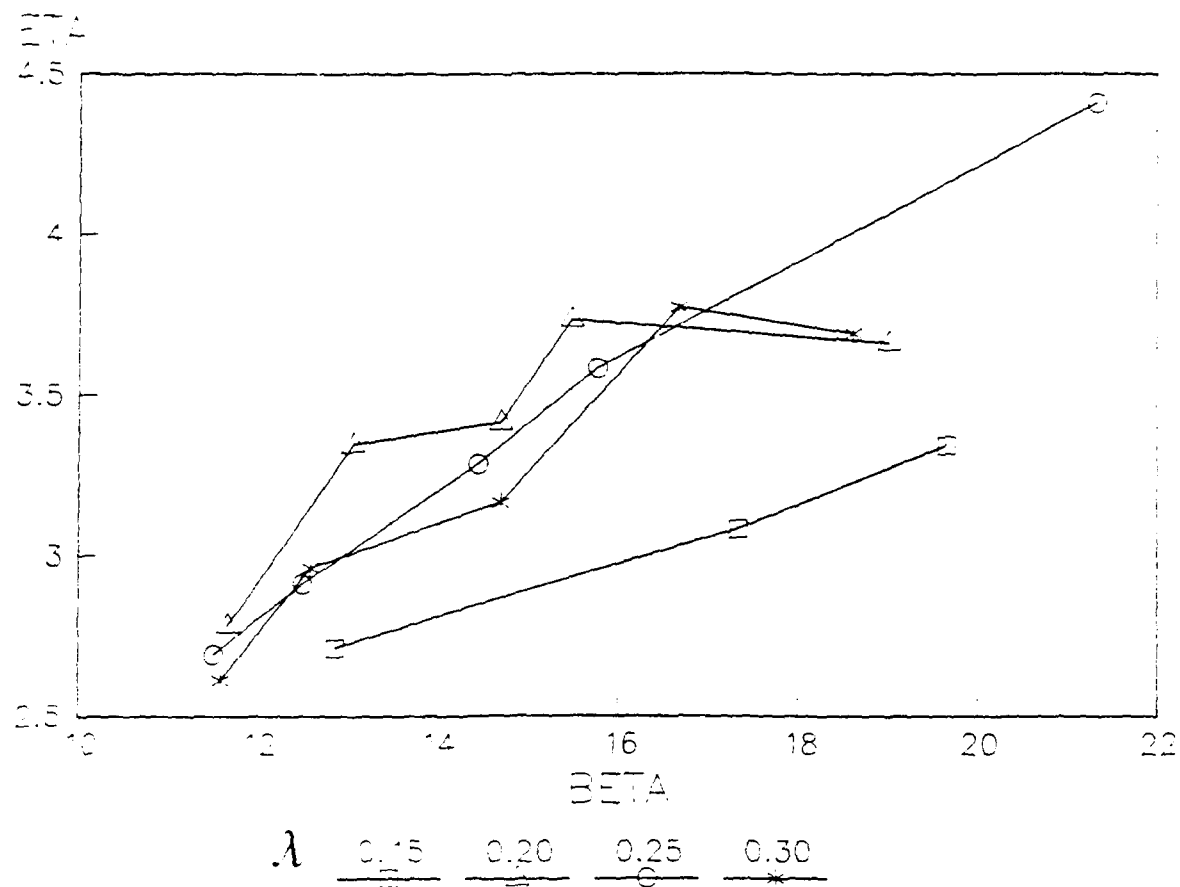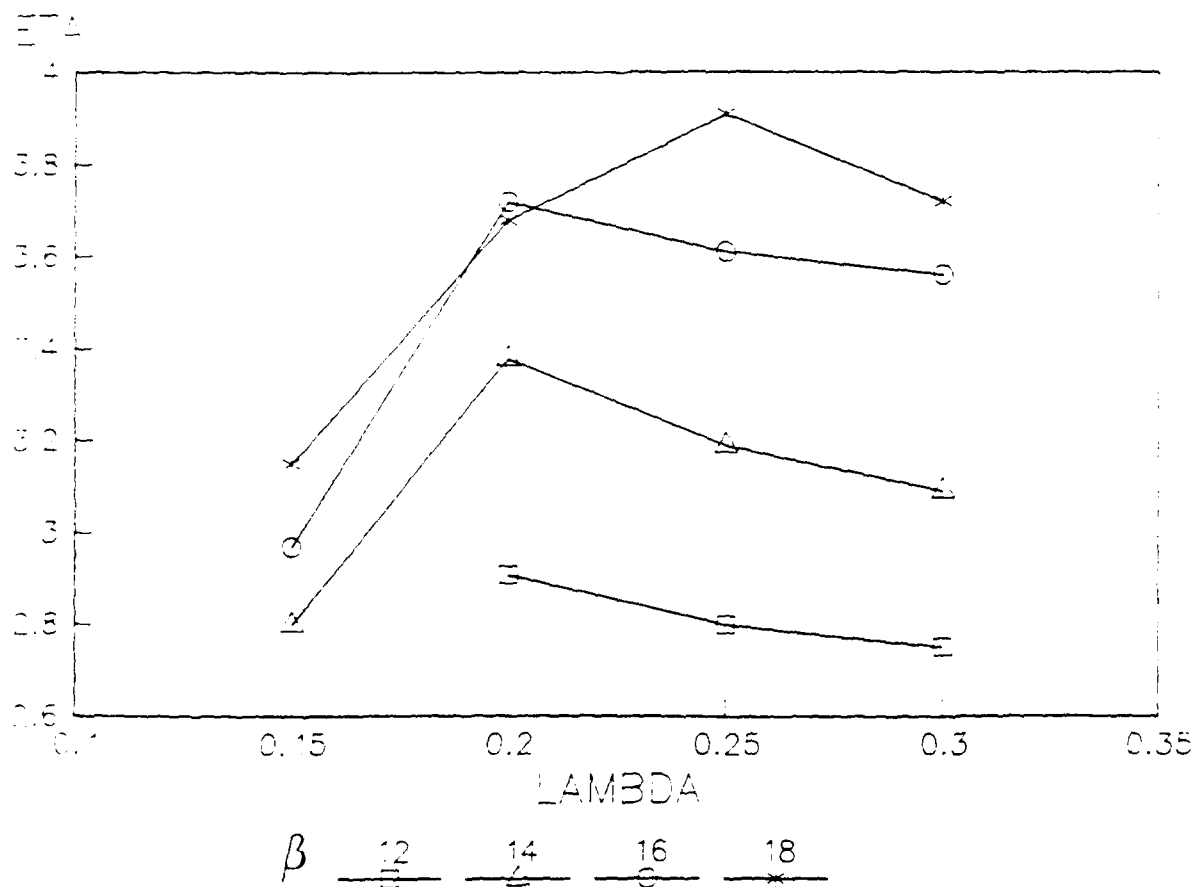Figure 1 - Geometrical sketch and notation.

Figure 2 - The amplitude of the linear perturbation of longitudinal velocity ($\nu\, U_1$) is plotted versus the intrinsic meander wavenumber λ_m and compared with theoretical findings derived within the context of a weakly nonlinear analysis ($\theta = 0.06$, $d_s = 0.06$, $\beta = 13$).
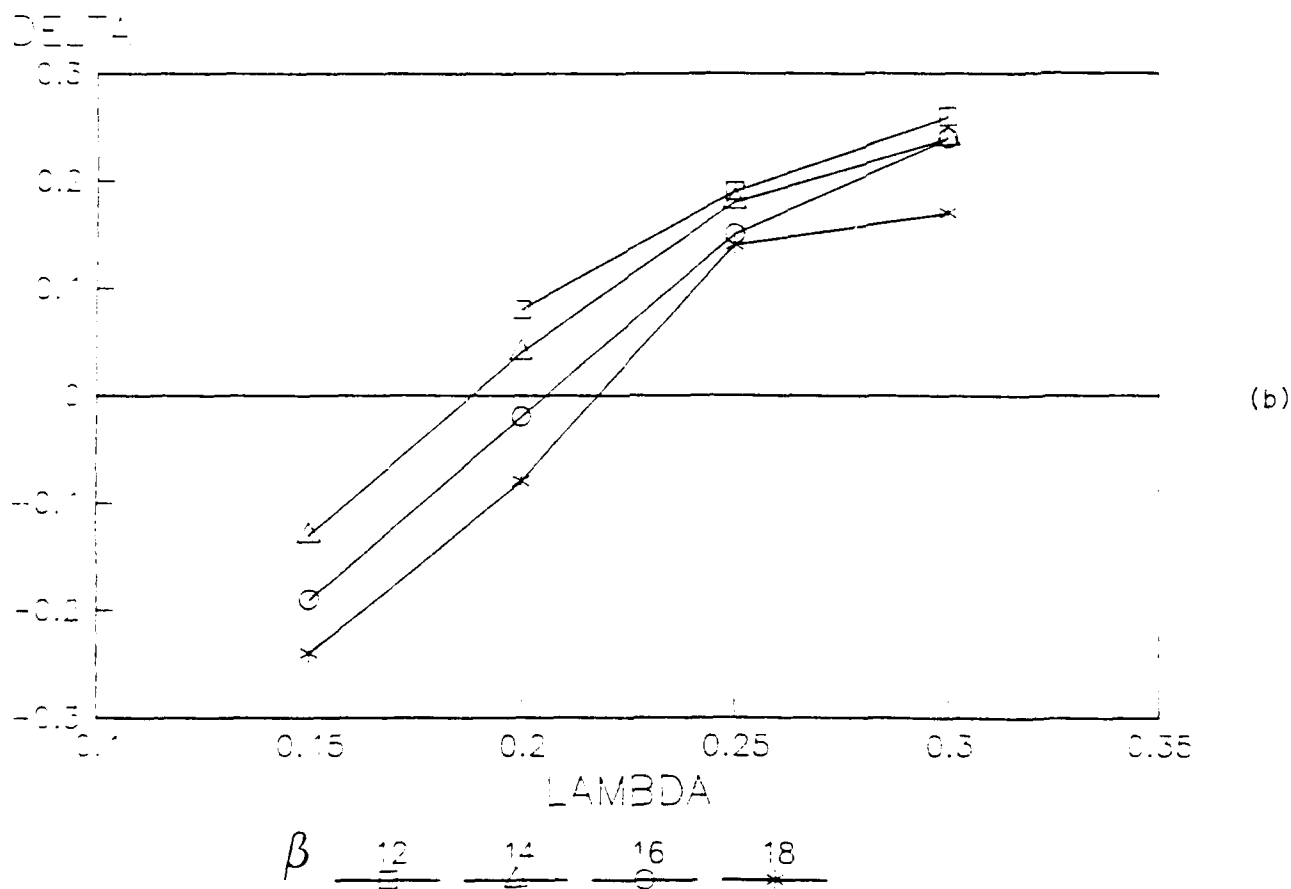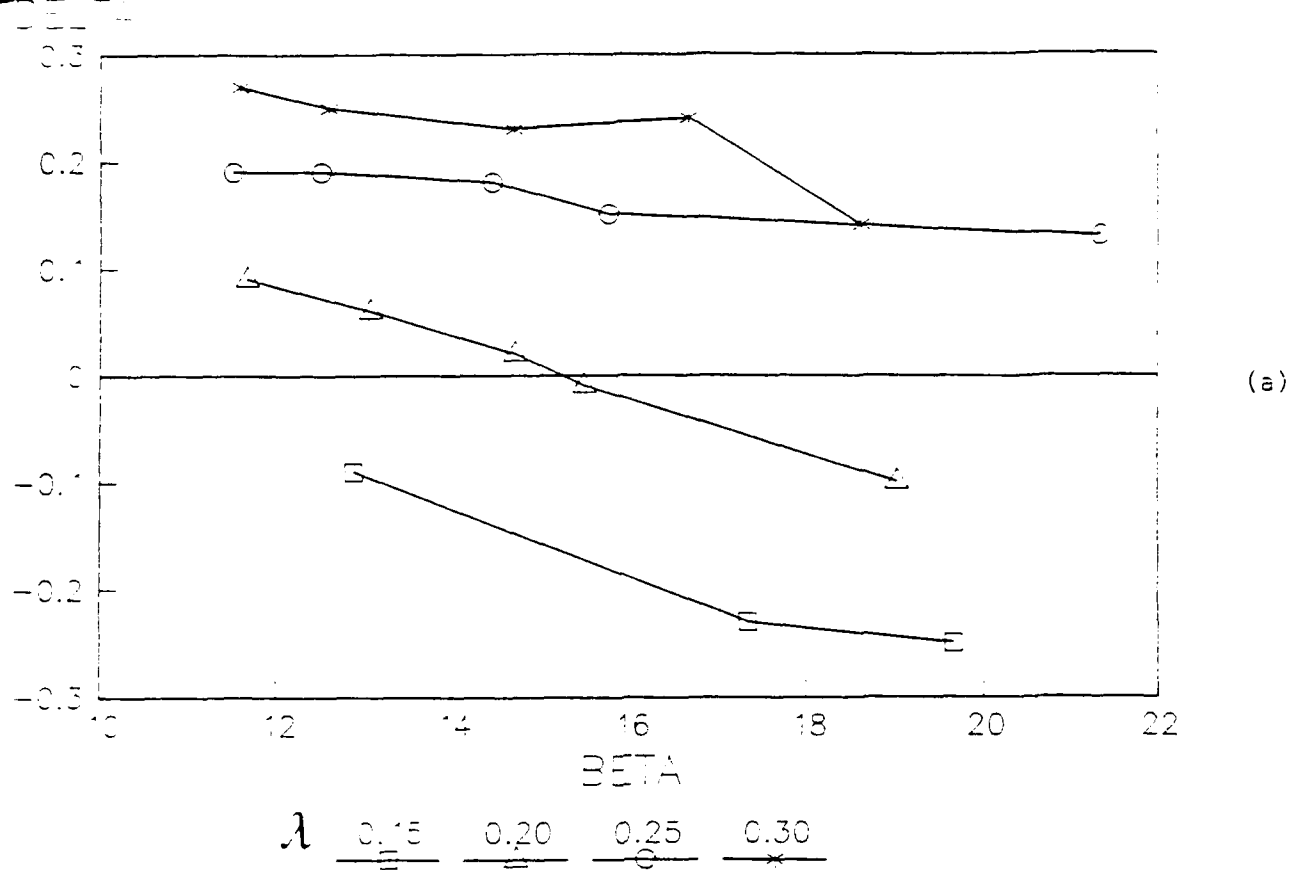
(a)

(b)

Figures 3 (a-b) - Dimensionless maximum scour η as a function of channel width ratio β (a) and wavenumber λ (b).

Figures 4 (a-b) - Phase-lag of scour hole with respect to bend apex as a function of channel width ratio β (a) and wavenumber λ (b).
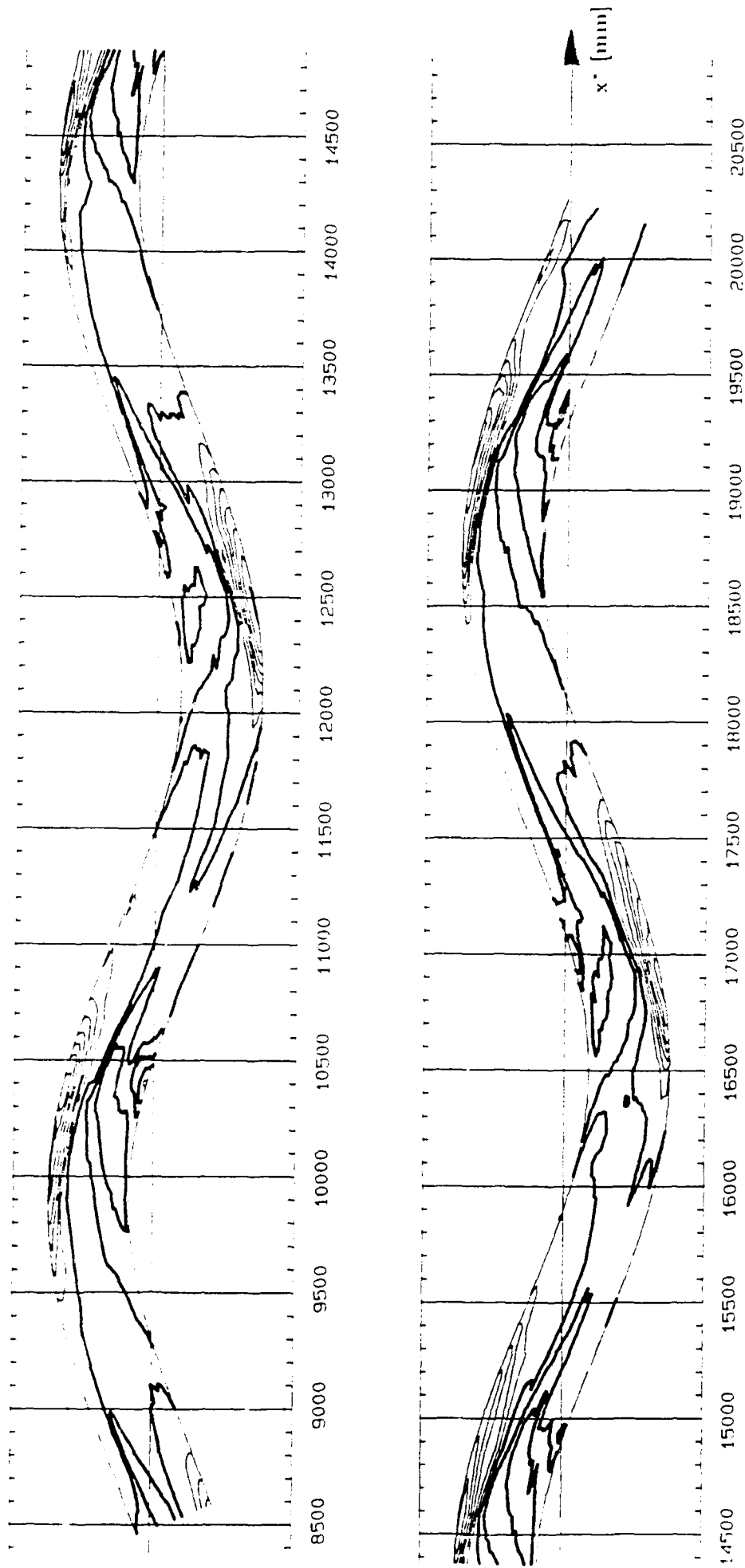
Figure 5 - Measured bed topography for run W25Q14. Contour intervals 0.5 cm. Thicker and thinner lines refer to deposition and scouring areas respectively.

(a)

(b)

(c)

(d)

Figures 6 (a-d) – Averaged bed topographies for runs W20Q07 (a), W20Q16 (b), W30Q07 (c), W30Q16 (d). (Contour lines as in figure 5).
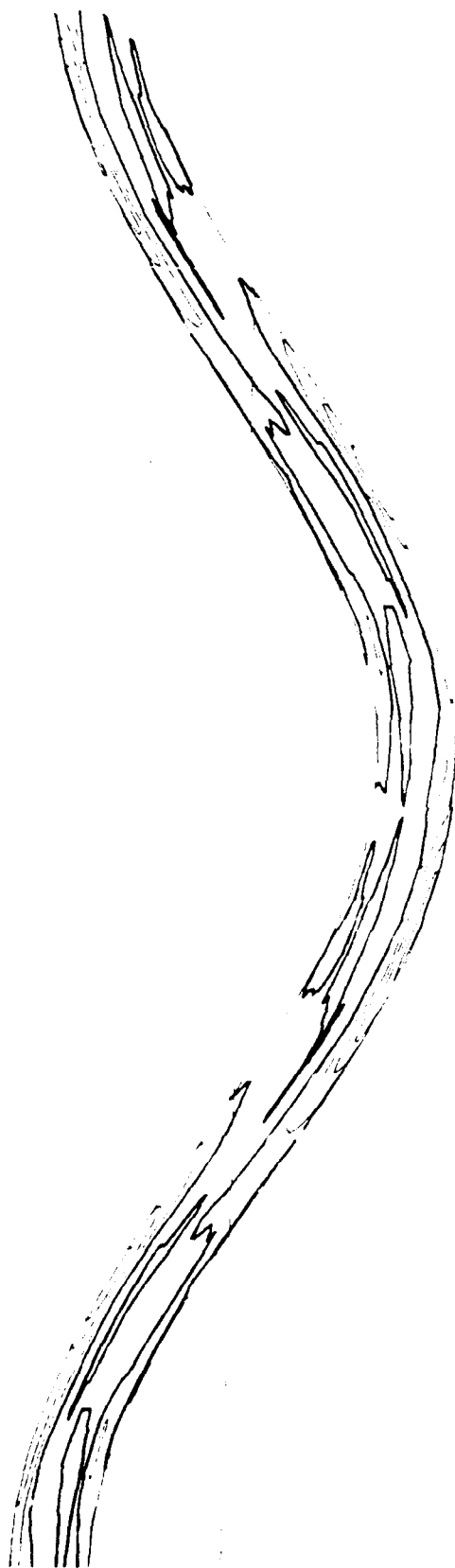
Figure 7 - Average bed topography for run W15Q10. (Contour lines as in figure 5).
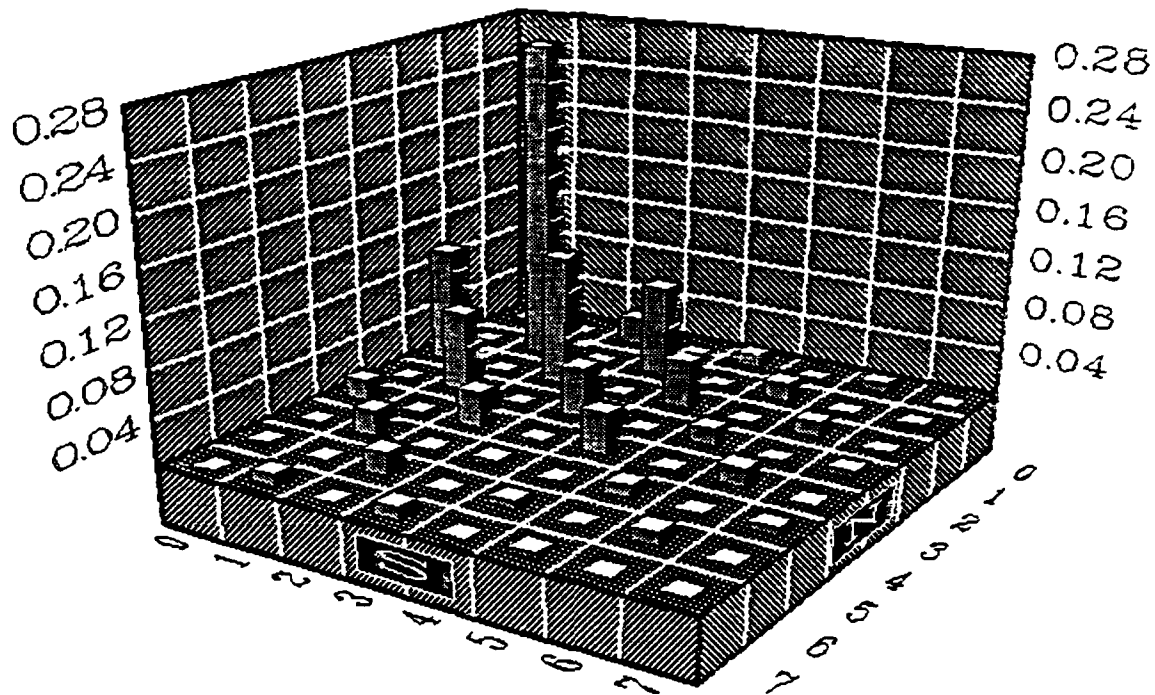
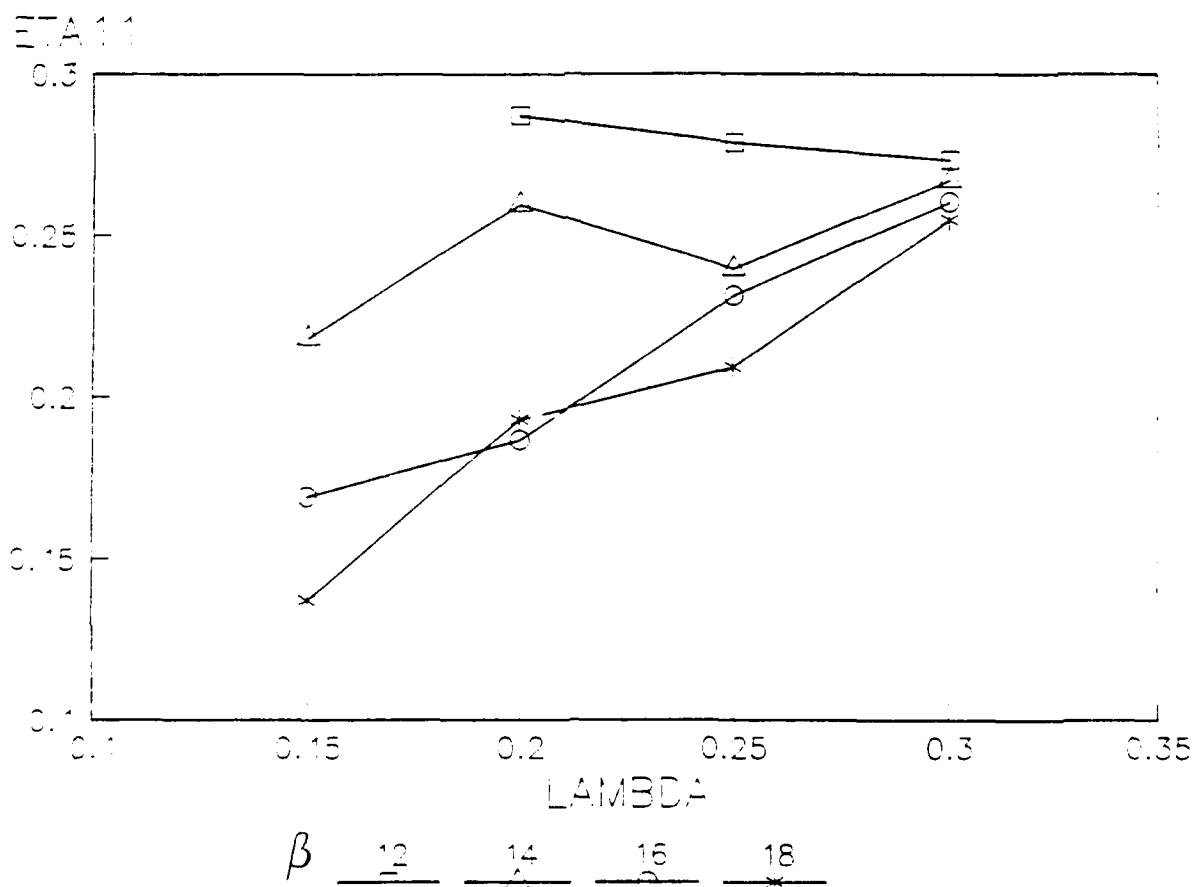Figure 8 - 2D Fourier spectrum of bed elevation for run W25Q14.

Figure 9 - The 1-1 harmonic of bed elevation (η_{11}) as a function of wavenumber λ for given
values of channel width ratio β.

The Sediment Size That Determines Channel Morphology

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Abstract

Studies of twelve gravel-bar streams in the mountains of Colorado and
Wyoming show that the bulk or largest volume of bedload is in the sand size,
much smaller than the size seen in the bed material and on the bars. The coarse
fraction, though small in volume of annual transport, determines and comprises
the major features of the channel morphology. High discharges carry bedload
only slightly more coarse than do low discharges, though the transport rate of
high discharge is very much larger than that at low flow.

In gravel-bed streams the bed material is much coarser than the sediment
that is largest in amount over a period of time. The coarsest few percent of the
total volume make up the bars, the riffles, and the surface material on the bed.
The largest clasts as well as the largest volume of the annual bedload are carried
at discharges near the bankfull stage.

The Sediment Size That Determines Channel Morphology

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A growing interest in gravel-bed streams has led to an expanded data base involving bedload transport rates determined by Helley-Smith sampler and associated sediment size distribution. There are also a few data sets in which the total bedload was caught in a settling basin or trap. The latter data include a measure of the large mobile clasts that exceed the size of the mouth of a sampler.

In the snow runoff seasons of 1988 and 1989 detailed measurements were made in several gravel streams in the mountains of Colorado. This unusually complete set of data on bedload and sediment size distribution allows several questions of interest to be discussed. Some characteristics of the streams studied are listed in Table 1.

Size Distribution of Bed Material and Trapped Sediment

The East Fork River near Boulder, Wyoming in the vicinity of the bedload trap is a few miles downstream from the place where the river leaves the mountains, flows through moraines, and thence meanders in a valley between terraces of glacial outwash. But soon after entering this valley zone, it is joined by a tributary, Muddy Creek, that is contributing sand washed out of glacial material. The sediment load from Muddy Creek is enhanced in recent times by irrigation return flow. The bedload trap and studies of the data have been published (Leopold and Emmett, 1976 and 1977; Emmett, 1980).

Table 1

Channel Characteristics at Locations Where
Sediment Samples Were Taken

Stream	Elevation (m)	Drainage Area (km^2)	Bankfull Discharge (m^3/s)	Channel Width (m)	Slope	D$_{84}$ of Bed Material (mm)
East Fork near Boulder, WY	2,200	466	20	15	0.0007	20
Poudre Pass Creek near Chambers Lake, CO	3,050		9.6	20	0.0062	150
Lower Trap Creek, near Chambers Lake	3,000	12.2	4.5	5	0.018	70
Upper Trap Creek, CO near Chambers Lake	3,050	9.8	4.5	9	0.060	152
SFK Cache La Poudre near Nederland, CO	2,340	228	11.3	14	0.008	147
Little Beaver Creek near Pingree Park	2,426	32	1.8	7	0.026	158
Left Hand Creek near Lyons, CO	2,288	134	5.9	9	0.035	120
Middle Boulder Creek near Nederland, CO	2,560	76	11	14	0.015	90
Goose Creek #1 near Cheesman, CO	2,304	209	3.8	7 ·	0.015	120
East Fork Encampment Creek, WY	2,921	9.1		3.3	0.027	170
Coon Creek, WY	2,906	16.8		5.1	0.023	220

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The sediment derived from the basin upstream of Muddy Creek is primarily gravel, which is prominently displayed in the riffles. The streambed in pools is generally covered with sand, derived for the most part from Muddy Creek. The bedload, then, consists principally of sand, but the gravel of the riffles also is moved. Painted rocks of gravel size placed on gravel riffles were carried away.

The major morphological features, riffles, central and point bars are composed of and formed by the coarse part of the total bedload. The bulk of the bedload is of finer material, principally sand.

The bedload trap on the East Fork River consisting of an open slot across the full width of the channel, 14.6 m, caught all the bedload. The size distribution of the trapped bedload is shown in Figure 1. Data for Figure 1 are in Emmett's Table 2 (1980). The data represent a composite of samples taken during 31 days in 1976. A single sample representing a given discharge was then the composite of about 40 scoops taken during a period of five to eight hours, or 80 kg of sediment.

The 31 days of sampling in 1976 extended through a whole snow-melt season beginning on May 18, when the discharge was 9.87 m^3/s, and ending June 21, when the discharge was 9.53 m^3/s. The peak discharge in this season was on June 5, when the discharge was 22.4 m^3/s. The bankfull discharge is about 20 m^3/s which in the annual flood series has a recurrence interval of 1.5 years. This snow-melt period of measurement in 1976 included discharges from 0.47 to 1.12 times the bankfull, a typical range in normal runoff years that includes the bulk of the sediment transport during a year.

Note in Figure 1 that the D_{50} caught in the conveyor-belt trap was 1.13 mm, and that 60 percent of the total weight caught in the trap was a size between 0.5 and 2.5 mm.

The bed material of the East Fork River is described by a composite of 232 individual samples collected at 29 cross sections in a 200-m reach near the bedload trap. The median particle size of each cross section varied from 0.6 mm to 25.4 mm. The largest median values occurred at the two riffles included in the sampled reach. A map showing the cross sections and the bed-material size in each is shown in Figure 5 of Emmett (1980).

The size distribution of the bed material is plotted in Figure 1 so that the bed material may be compared with the sediment caught in the conveyor-belt trap. The D_{50} of the bed material is 1.25 mm. Sixty percent of the total weight is made up of sizes between 0.5 and 12 mm.

The difference between the trapped sediment and the bed material is in the size classes larger than the median. The distribution of the finer 50 percent by weight is nearly the same for the trapped sediment and the bed material, as can be seen in the near coincidence of the two curves of Figure 1 in the range of 0 to 50 percent by weight.

Another installation where the total bedload was trapped is constructed on two gravel-bed streams in the Medicine Bow area of Wyoming, and measured by the U.S. Forest Service (Wilcox, memorandum of 1989).

On each of Coon Creek and East Fork Encampment Creek (see Table 1), a large wooden box or enclosure was constructed to trap all incoming bedload. Each year, the sediment was excavated by machine and the volume and size distribution measured. Transport rate of bedload was measured during storm periods with a Helley-Smith sampler.

In the years 1986—1988, the sediment removed was 5.5 tons, 9.2 tons, and 12.8 tons, respectively, at Encampment Creek. In the years 1983—1986 and 1988, the totals were 220, 75, 60, 150, and 50 tons, respectively, at Coon Creek. This amounted to an annual yield of 0.92 tonnes/km^2/year from Encampment Creek and 6.0 tonnes/km^2/year from Coon Creek.

These unique measurement data are summarized in Figures 2 and 3. In these graphs the size distribution of four materials is plotted, (1) the bulk of the sediment excavated from the box trap labelled "pond," (2) bedload samples taken during storm events with a Helley-Smith sampler, (3) bulk sample of bed material, and (4) bed material measured by pebble count sampling.

Both Encampment Creek and Coon Creek data are similar in certain respects despite the difference in annual sediment yield. The size distribution of the bulk material collected in the box traps was nearly identical to that obtained in the Helley-Smith sampler.

In the largest particle size, above 64 mm, the sampler under-represented the actual load. In the box trap, the sediment caught included a few percent by weight of sizes of 128, 256, and 512 mm that clearly could not go into the 10 cm mouth of the sampler.

In all three of these installations where the total bedload was trapped, the bulk of the load was of sand size despite the fact that the bed material was of much larger size. In the East Fork the riffles are gravel, not sand. D_{84} of the bed material, 20 mm, is typical of what one sees on the riffles and bars.

In Coon Creek and Encampment Creek the bed material is gravel having a D_{84} of 220 and 120 mm, respectively. Yet, 70 percent of the material caught in the box trap was less than 6 mm. The D_{50} of the trapped material was between 1.5 and 2 mm. Thus, the bulk of bedload in these gravel streams is sand yet the streambed is obviously gravel and cobbles.

This conclusion applies not only to the three streams where a trap caught all of the bedload, but also to the other gravel rivers discussed. Table 2 shows the sediment size category comparing the largest percentage by weight of the bedload sampled in the eight mountain streams studied. Note that in all these gravel streams, the size category of either 1 or 2 mm comprised from 15 to 33 percent of the weight of bedload caught during the runoff season. The Colorado streams are gravel bed with D_{84} of bed material from 150 to 200 mm.

All the streams listed in Table 2 can be classified as gravel-bed channels in that the bars, the riffles, and the point bars are predominantly composed of gravel. In all of these the bed material or surface layer of clasts is somewhat coarser than clasts immediately below the surface. The ubiquitous condition of gravel-bed channels has unfortunately been called armoring or paved, implying the presence of a pavement. Such terms suggest that the surface layer or bed material does not move. Measurements show that even the D_{84} of the bed

Table 2

Season Totals of Bedload Samples
Caught in Helley-Smith Sampler, 1989

Stream	Number of Days	Total Weight Caught (g)	Total Weight Per Day of Sampling (g)	Largest Rock Caught (mm)	Size Category[*] With Largest % of Weight	
					mm	% of Total Weight in Category
Goose Creek #1	17	40,319	2,371	30	1.0	31
Goose Creek #2	10	3,348	334	64	1.0	23
Goose Creek #4	16	8,578	536	41	1.0	23
Left Hand	56	211,297	3,773	113	1.0	33
Middle Boulder	20	32,792	1,639	98	2.0	15
SFK Cache La Poudre	40	11,196	279	43	1.0	27
Little Beaver	40	10,389	259	45	1.0	30
Upper Trap	22	13,841	629	136	2.0	19

*Sizes listed are passing through, not held on.
 Sieve sizes were 0.5, 1, 2, 4, 5, 6, 8, 11.2, 16, 22, 32, and 45 mm.
 Size category of 1 mm is from .5 to 1 mm.

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material moves in discharges below or at bankfull, but in small quantities. Thus, the occurrence of large clasts larger than 100 mm in the material caught in the box traps or ponds confirms the mobility of bed material.

Size Distribution of the Bed Material

In both the Encampment and Coon Creeks bed material was measured in two ways, by a volume taken to the laboratory for drying, sieving and weighing, and by pebble counting.

In both these streams there is a marked difference in sorting of the bulk of the load, mostly sand, and the gravel material. Note in Figures 2 and 3 that the box trap material labeled "pond" is poorly sorted in that there are equal percentages by weight in categories of size from 0.5 mm to 10 mm. In contrast, bed material gravel has a sharp maximum of about 100 mm with smaller percentages by weight of both smaller and larger sizes.

This difference in sorting appears to be general. Figure 4 presents the size distribution of bed material and that caught in the Helley-Smith sampler for three of the mountain gravel bed streams in Colorado. In these data, the same finding is seen for a rather straight cumulative curve describing the bedload caught and the markedly S-shaped curve for bed material. Such curves are available for most of the streams listed in Table 1 and they show similar tendencies.

For many of these streams, size distributions of gravel on channel bars and in the subarmor or subpavement are available. The subarmor nearly always has a curve that lies between the bed material and the bedload caught in the sampler, that is finer than bed material but coarser than the caught bedload.

Implications for Channel Morphology

The largest portion of the bedload in these gravel bed channels is not in the gravel size but in the sand size. Yet the visual impression of the channels is formed by the coarser fraction that makes up not only the bed, but also the point bars and the pool-riffle sequence. Thus, the principal morphologic features of the channel are made of material representing only a few percent of the total annual bedload. The median size of the bed material in these data is equalled or exceeded in only about three percent of the total bedload moved.

Discharges that Move the Bulk of the Bedload

The measurement points of streams in Table 1 are all located in proximity to a gaging station for which a flow duration curve was plotted. A bedload rating curve constructed from Helley-Smith sampling is also available. Applying the percentage of time various discharges were experienced to the associated bedload transport rate, the total bedload carried by each discharge category was computed. The computation was carried out for each bedload sampling point.

A typical relation obtained is shown in Figure 5 for Left Hand Creek, drainage area 134 km^2, D$_{84}$ of bed material 120 mm. The upper graph shows the percentage of time various discharges occur. The data are shown in the form of a cumulative curve typically used for a flow duration curve, but also as a histogram of non-cumulated percentages. The most common category was about 0.1 m^3/s that occurred 24 percent of the time. The average discharge was equalled or exceeded 20 percent of the time, a common characteristic for many rivers.

The bottom graph in Figure 5 shows the percentage of the annual or long term volume of both water and bedload carried by various categories of discharge. The largest volume of water is carried in the category of discharge 2.8-4.5 m³/s, or 0.47—0.76 times bankfull discharge.

The maximum bedload is carried in the discharge category 4.5-7.3 m³/s or 0.76—1.2 times bankfull. At discharges both smaller and larger, less bedload is transported. The discharge carrying the maximum sediment load has been called the effective discharge. In a study of some gravel-bed streams in Colorado, Andrews (1984) found that the effective discharge was approximately equal to the bankfull discharge. The example above is in agreement with Andrews' findings.

Sediment Size Distribution As Discharge Changes

One might expect the size of bedload moved to increase proportionally with increase in discharge. The data show that such increase does occur, but its magnitude is rather small. Two examples are given in Figures 6 and 7. In these figures the size distribution of bedload caught in a Helley-Smith sampler is shown for a large and a small discharge. In Figure 6, at the East Fork in Wyoming, the five-fold increase in discharge resulted in a change of D_{50} from 0.7 to 1.4 mm or a two-fold increase. In Figure 7, at Goose Creek #1, a change in discharge from 0.76 to 4.1 m³/s resulted in a change of D_{50} from 0.55 to 0.75 mm. In all cases studied, low and high discharges both had a wide distribution of bedload size from sand to small gravel, but the increase in size with discharge was not very large.

Concluding Discussion

Gravel-bed streams have a bedload smaller in grain size than what is seen on the channel bed. Sand is the size of the major volume of bedload in gravel-bed streams. However, the major morphologic features of gravel-bed streams are composed of gravel, not the sand that makes up the bulk of bedload. Channel bed material, point bars, and riffles to the pool-riffle sequence, are made up of gravel that moves at low transport rate, the clasts of which move only occasionally, and for short distances.

Size distribution is characterized by better sorting in the coarse bed material than is the sandy bulk of the bedload. Size in any category is only slightly larger in high than in low discharges. Thus, it appears that the gravel-bed channels, the coarse fraction, moved only occasionally and moved only short distances in any one season, makes up the principal morphologic features of the channel.

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FIGURE CAPTIONS

Figure 1.   Size distribution of bed material and of sediment caught in the bedload trap, East Fork River, Wyoming. The bed material data are the average of 232 samples. Sediment caught in the conveyor-belt trap, transport weighted, is an average of 52 samples caught in 31 days during 1976 (Emmett, 1980, Table 2). Upper diagram, percent by weight of sediment of different size classes, not cumulated. Lower diagram, same as upper diagram, but cumulated; the ordinate is percent by weight equal to or less than each size class.

Figure 2.   Size distribution of bedload in Coon Creek, Medicine Bow National Forest, Wyoming, showing four types of measurements. Bedload in box-pond is nearly identical to that caught in Helley-Smith sampler, but the latter did not catch the clasts greater than 64 mm. Bed material in bulk sample is nearly the same as size determined by pebble count.

Figure 3.   Size distribution of bedload in E. Fork Encampment Creek, Medicine Bow National Forest, Wyoming, showing four types of measurements. As in Coon Creek, sediment caught in box-pond is nearly the same as that in Helley-Smith sampler except for large clasts. Bed material and pebble count data are nearly the same.

Figure 4.   Size distribution of bedload in three mountain streams in Colorado. Bedload caught in Helley-Smith sampler is compared with bed material of the channel.

Figure 5.   Relation of discharge to the percent of time and percent of volume of water in a Colorado mountain stream. The relation of discharge to volume of sediment is shown for bedload. For both relations, data are expressed as a histogram, and the same data is expressed as cumulated values, equal to or more than...

Figure 6.   Bedload size distribution during the day of lowest and highest discharge in the year 1975 at the bedload trap, East Fork, Wyoming.

Figure 7.   Bedload size distribution determined by Helley-Smith sampler at the time of lowest and highest discharge in 1989, Goose Creek #1, Pike National Forest, Colorado.

EAST FORK

o - - BED MATERIAL

x — CAUGHT IN TRAP

FIG. 1

COON CREEK

SEDIMENT SIZE IN MM

◇ — — HELLEY-SMITH   ×——× BED MATERIAL

+——— POND   •—••— PEBBLE COUNT

FIG. 2

ENCAMPMENT CR.

PERCENT BY WEIGHT

SEDIMENT SIZE IN MM

Σ PERCENT BY WEIGHT

◇ — — HELLEY-SMITH   ×—× BED MATERIAL
+ —— POND   •—··— PEBBLE COUNT

FIG. 3

FIG. 4

LEFT HAND CREEK, COLO.

FIG 5

EAST FORK

— ×— 6/2/75 Q= 5.82 CMS
— o— 6/8/75 Q=25.6

FIG. 6

GOOSE CK. #1

5/13/89 .76 CMS

6/29/89 4.1

FIG. 7

SPATIAL PATTERNS OF BEDLOAD TRANSPORT AND CHANNEL CHANGE IN BRAIDED AND NEAR-BRAIDED RIVERS

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ABSTRACT

Better understanding of channel change in low-sinuosity gravel-bed rivers requires attention to within-reach spatial patterns of nonuniform flow and bedload transport. Field techniques for measuring such patterns are discussed and results illustrated. Point sampling of bedload is the weakest link in this approach. Mean transport rat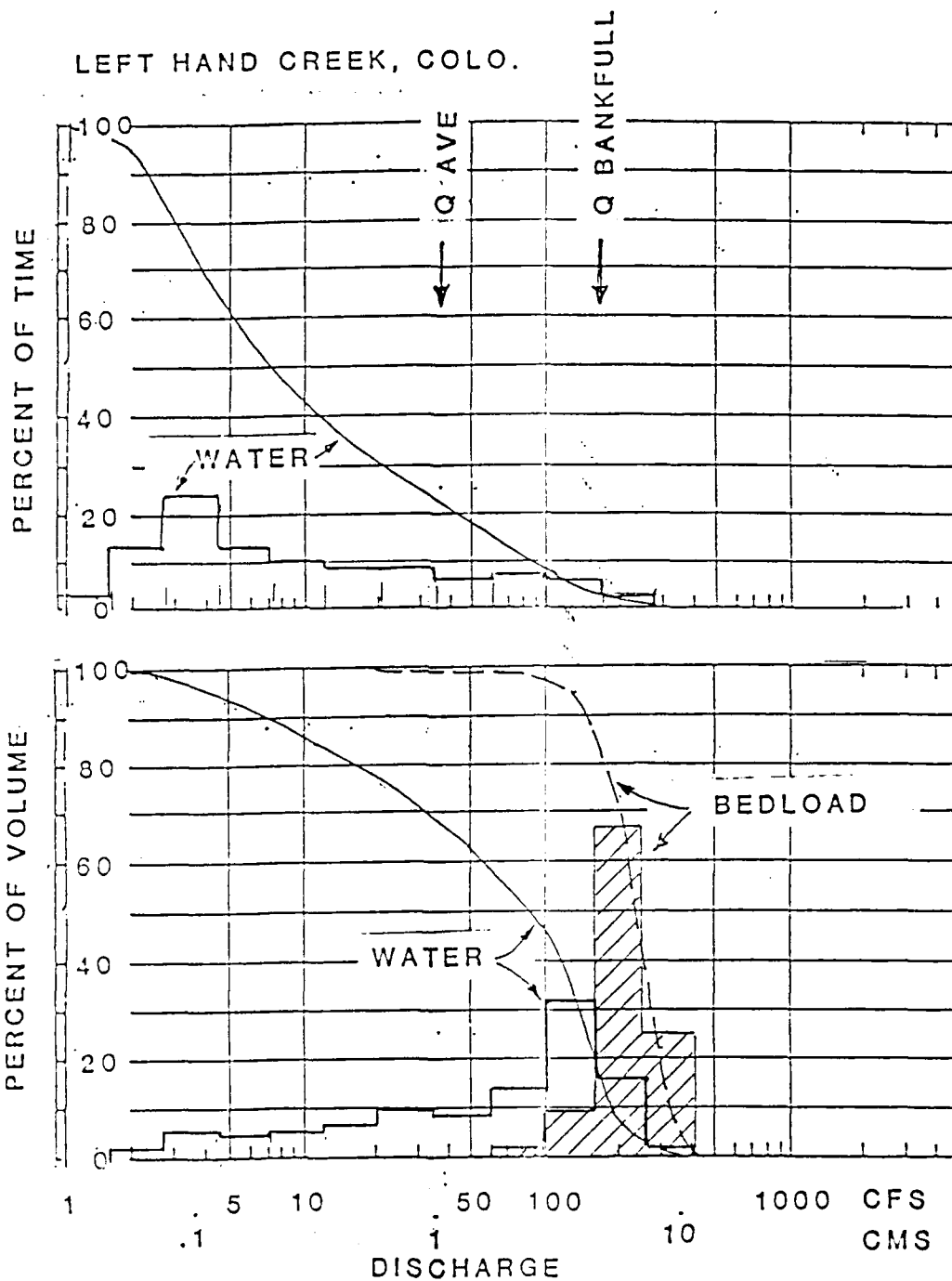es and streamwise variations in transport can alternatively be estimated from measured channel change using within-reach sediment budgeting.

1 INTRODUCTION

The last decade has seen an upsurge of research on the dynamics of gravel-bed rivers. Laboratory, field, and mathematical approaches have all been adopted, but each usually in a reductionist framework in which particular aspects of the basic physics are isolated as they operate at point scale or in the idealised context of uniform steady flow. This strategy has yielded greatly improved understanding of the basic processes, but arguably has not yet led to a comparable advance in understanding of the behaviour of gravel-bed rivers at the reach scale.

The difficulty in applying at the *reach* scale new understanding of hydraulics and gravel transport at the *point* scale lies in the nonuniformity of flow over lengths greater than a few channel widths. Allied to this are lateral and longitudinal variations in bed texture and bedload transport, leading to the possibility that erosion may occur in one part of a reach at the same time as deposition elsewhere. These different aspects of nonuniformity are of course interrrelated in a feedback system (Figure 1). The existing irregular channel configuration (three-dimensional form together with bed roughness)
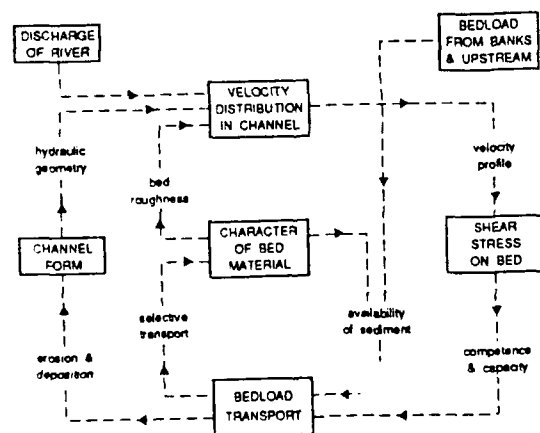
Figure 1: System diagram of the interrelationship of channel configuration, flow, and bedload transport in gravel-bed rivers (modified from Ashworth & Ferguson 1986).

controls the nonuniform pattern of flow through the reach at any given discharge. Since discharge is variable the flow is unsteady as well as nonuniform, and its pattern of nonuniformity may vary with stage. The spatially and temporally variable stresses exerted on the bed combine with the availability of sediment, whether in the bed or supplied from local bank erosion or upstream, to determine the rates of transport of different sizes of bedload at different places and times. Since this transport is nonuniform and unsteady, erosion and deposition alter the initial channel geometry, and selective entrainment or deposition may alter the texture of the bed. In braided rivers, channel change may also alter the proportions of total discharge flowing through individual channels. The general character of the river may remain similar in the absence of environmental change or human intervention, but the details of individual reaches are liable to change during competent flows.

To geomorphologists and sedimentologists the challenge is to explain how channel configurations develop through channel change, and how channel change relates to nonuniform processes. Such an understanding also promises better engineering predictions of channel change in specific circumstances. For meandering rivers, there has long been a qualitative understanding and quantitative theory has now been developed to the point that predictions can be made of facies patterns and planform evolution (e.g. Parker & Andrews 1985, 1986). But for braided rivers there are no standard quantitative models for predicting flow patterns or rates of channel change, and even the mode of channel change may be hard to foresee (e.g. Mosley 1987, p.823-4). The few field observations that have been published suggest that similar-looking initial configurations can evolve in different manners (e.g. Church & Jones 1982 Figure 4), and that the same reach can change in essentially opposite ways on different days according to peak flow levels (Ashworth & Ferguson 1986).

In principle it should be possible to model numerically the short-term development of a braided reach. This would involve iterating around the feedback system of Figure 1, using a 2-D finite representation of the vertically-averaged flow to drive a sediment transport and budget model, as proposed by Nelson & Smith (1989) for meandering channels. However, published numerical models for nonuniform stream flow cannot cope with rapid streamwise changes in width, nor with bifurcations and confluences of flow. There are also differences of opinion on the parameterisation of the transport threshold, the transport function, and the lateral and vertical sorting of bed material.

Once the modelling approach is extended to braided channels it will be necessary to validate it by observations. An empirical approach may also give direct insight into how braided rivers work. We have been attempting this for several years. It requires process measurements as well as observations of channel change during periods of competent flow, and the measurements must be spatially distributed. Many practical problems must be overcome to achieve this. This workshop contribution summarises some of our findings, but our main intention is to pass on our experience in making measurements in active gravel-bed rivers and relating process measurements to channel change.

2 PROBLEMS IN OBTAINING SPATIALLY DISTRIBUTED MEASUREMENTS

2.1 Choice of study area

We argued earlier that most gravel-bed rivers are nonuniform and undergo channel change, but it does not follow that one river is as good as another for an observational study. The need for spatially distributed measurements creates serious practical problems which constrain which rivers can usefully be studied.

The first problem is scale. Froude-scaled laboratory channels give insight into typical modes of behaviour and whole-river sediment transport (e.g. Ashmore 1982), but are so shallow and rapidly changing that spatially distributed measurements are hard to make. Prototype braided rivers, in contrast, tend to be too big and fast for easy or safe measurement by wading, and too rapidly shifting for footbridges to survive long. Davoran & Mosley (1986) have shown the potential of jetboats as measurement platforms, but the necessary equipment and expertise are rarely available. In our field research we have opted for wading (using a wetsuit when necessary) as the simplest and most versatile means of repeating measurements at many points within a reach, and have therefore restricted our studies to small rivers, with discharges of about 5-20 m^3s^{-1} in marginally active conditions.

A second problem relates to hydrological regime. Making measurements at enough points within a reach to gain an idea of spatial patterns of hydraulics and sediment transport can take several hours, over which there must not be significant change in discharge. This rules out floods caused by rainfall events in small catchments. We have worked instead in rivers with partly or wholly nival or glacial regimes in order to take advantage of gradually varying high flows at predictable times of day and year. The rivers concerned are the Lyngsdalselva, a proglacial braided stream in arctic Norway; the Feshie, a near-braided channel in the Scottish Highlands; the White River, a locally braided glacier-fed stream in the northwest US; and the Sunwapta, a proglacial braided river in the Canadian Rocky Mountains.

In each case we selected short study reaches with well developed examples of what we and others regard as key morphological elements of the prevailing channel pattern: a pair of alternate bars and associated pools and riffles in the Feshie, and an X- or Y-shaped braid confluence and bifurcation in the three braided rivers. Simple examples of these basic elements, fairly symmetric and without complications, are even harder to find than simple, regular meander loops; they are therefore characteristic rather than typical, but if we cannot understand how they work, there is little hope of understanding more complicated situations.

2.2 Quantifying channel configuration and change

Channel plan form in braided rivers is highly stage dependent, as well as altering over time through erosion and deposition. Mapping water lines on different dates, whether from aerial photos or by ground survey, is not a good guide to channel changes because it confounds stage effects with erosion and deposition. Moreover, mapping cannot reveal scour or fill within the wetted perimeter of any channel. It is therefore essential to make repeated surveys of cross sections, in our case by levelling during the morning/midday low of the diurnal flow cycle. Echo sounding should be possible in larger rivers. The attainable accuracy is limited by the irregular roughness of the bed, but inferred depths of erosion or deposition between surveys should be accurate to within one bed particle diameter. The endpoints of each cross section are marked by wooden posts or metal rods between which can be stretched a tape measure, and all surveys are tied in to the same benchmark. Graphical presentation of channel changes is easier if the sections form a simple rectangular grid, but this may be inappropriate in curved or bifurcating channels.

The other aspect of the channel configuration that must be quantified is the texture of the bed, since this influences both the flow resistance and the availability of different size fractions of bedload. Flow resistance in gravel-bed rivers is best quantified from a size-by-number (i.e. pebble count) grain-size distribution, either using some multiple of the D_{50}, D_{84}, or D_{90} percentile (e.g. Bray 1980) or using the full distribution (Wiberg & Smith 1987). Sediment availability is normally quantified by a size-by-weight (i.e. bulk sieving) distribution, and is commonly complicated by a difference between relatively coarse surface layer and relatively finer subsurface sediment. Bulk sieving to the standards proposed by Church et al. (1987) is time consuming even for a single site, and much more so when spatial patterns and temporal changes are of interest. We have tended therefore to use the quicker pebble counting approach to quantify spatial and temporal variations in surface texture, and collect and sieve large bulk samples of surface and subsurface sediment only at the start, end, or both of a study and from a few locations within a reach. Size distributions obtained in these different ways are intercomparable to some extent so long as they are truncated at common endpoints.

2.3 Flow measurements

In a braided river a knowledge of total discharge, e.g. from a gauging station at an undivided section upstream or downstream, is of limited value for a reach study because the split of total discharge between individual channels can alter considerably from day to day. A manual or automated record of flow stage within the study reach gives a better guide to day-to-day variations, and can be supplemented by velocity-area or salt-dilution gauging if discharge values really are required.

The surface streamline pattern within a reach can be visualised by observing the paths of floats introduced at different lateral positions, and quantified by compass bearings of surface flow direction at different stations along cross sections. Flow direction near the bed may differ from surface direction, especially in confluences where a twin-cell secondary circulation is thought to exist

(Ashmore 1982). It can be measured in clear water by attaching a streamer to a mechanical current meter, or in turbid water by using a two-component electromagnetic current meter.

In our studies we have regarded bed shear stress, τ, as the key flow variable for comparison with bedload transport rates. The local value of the main alternative (specific stream power) is most easily estimated as the product of shear stress and mean velocity, so poses the same measurement problems. The formula

$$\tau = \rho grs \tag{1}$$

(where ρ is water density, g the gravity acceleration, r the hydraulic radius of the channel and s the water surface slope) is often used for estimating shear stress in gravel-bed rivers but can be inappropriate for several reasons.

(1) It gives only the mean stress averaged over the wetted perimeter, whereas the local stress will tend to increase from banks to talweg. Some workers (e.g. Davoran & Mosley 1986) have therefore replaced r by the local flow depth d in equation (1).

(2) Shear stress is only proportional to the width-averaged or local depth-slope product in *uniform* flow. In channels with streamwise variation in depth and/or width, and therefore convective acceleration and deceleration of flow, the momentum equation contains extra terms which do not necessarily cancel out (Dietrich & Whiting 1989).

(3) The shear stress estimated from the energy gradient is the total stress balancing both skin and form resistance to flow, whereas the effective stress for bedload transport is generally regarded as the skin-drag component alone. In gravel-bed rivers with high relative roughness (ratio of characteristic grainsize to flow depth) as much as half the total stress may be associated with the form drag of large protruding clasts (Wiberg & Smith 1987).

(4) At a practical level, accurate measurement of water surface slope can be difficult in steep shallow rivers with lateral slopes over bars, hydraulic jumps in the lee of the largest clasts, standing waves around survey staffs and wading rods, and temporal variations as pools and riffles are drowned out.

Our preferred alternative to the depth-slope method is to measure local velocity profiles and estimate the bed shear stress by fitting the logarithmic "law of the wall". In 1984-7 (Lyngsdalselva, Feshie, and White River) we used an array of current meters on a single wading rod in order to measure a four-point velocity profile in 30 seconds. In practice, replication was often necessary because in glacial rivers the bearings of some makes of current meter are intermittently clogged by suspended sediment. In 1989 (Sunwapta River) we therefore switched to using a single current meter on a topsetting rod which allowed the height to be altered without removing and replacing the wading rod, and thus without any change in the precise location and zero height of the velocity profile. This is slower than using an array of meters, but by assembling a larger team of fieldworkers we were able to measure as many profiles per hour.

The "law of the wall" for a turbulent boundary layer states that velocity u varies with height z as

$$u/u* = (1/\kappa)\ln(z/z_0) \tag{2}$$

where $\kappa = 0.4$ is Von Karman's constant, $u* = (\tau/\rho)^{0.5}$ is the shear velocity, and z_0 is the roughness height of the bed. If z is accurately determined, as in the field procedures we have used (but not when a wading rod is removed and replaced each time z is altered), ordinary least squares regression of u on $\ln(z)$ can be used to fit equation (2). This can be done in the field with a good calculator. If the best fit intercept and slope are a and b, and ρ is taken as 1000 kg m^{-3}, the shear stress τ and roughness height z_0 are given by

$$\tau = 160b^2 \tag{3}$$

$$z_0 = \exp(-a/b) \tag{4}$$

Integration of the logarithmic profile over the full flow depth d shows that the mean velocity $<u>$ occurs at height d/e and can thus be estimated as

$$<u> = a - b + b.\ln(d) \tag{5}$$

The standard errors s_a, s_b of the regression coefficients a, b can be used to calculate standard errors and confidence intervals for the estimated shear stress and roughness height (Wilkinson 1984). In particular, it is obvious from (3) that

$$s_\tau/\tau = 2 s_b/b \tag{6}$$

Others things being equal, a greater number and range of measurement heights will yield a lower value of s_b and a more precise estimate of the shear stress. In 1989, with 6-10 points per profile, our shear stress estimates had standard errors averaging about 10%, compared to upwards of 20% for our earlier 3- or 4-point profiles. There are, however, problems in extending the measurements throughout the flow depth. Measurements very close to the bed are very sensitive to the zero height used, i.e. the positioning of the wading rod on the rough bed. In our experience it is impossible to apply zero plane displacements in an objective way to overcome this problem, even though it is often easy to see from curvature in the logarithmic plot (or abnormally low or high estimated roughness heights) that a problem exists. Zero plane error in z decreases in relative terms as z increases, so measurements above the height of the largest clasts are not significantly affected; they are also far less vulnerable to distortion by local jet or wake effects near large clasts, and to physical blocking of the current meter by bedload.

Measurements far from the bed may also be misleading. It is known that velocity profiles can be distorted by secondary circulation, convective acceleration, and streamwise changes in bed roughness. Many authors assume that, even without these complications, velocity profiles in open channels are logarithmic only in the bottom 10% or 20% of the depth (e.g. Bathurst 1982). However, other authorities dispute this (e.g. Yalin 1977 p.30-33) and some detailed flume studies (e.g. Cardoso et al. 1989) have found minimal deviation from a logarithmic profile almost to the surface. We have generally found little departure except in confluences (presumably affected by secondary currents or underflow/overflow), though fitting the log law to shallow bar-head profiles can give surprisingly high estimates of shear stress, perhaps because of acceleration or roughness effects.

A more general problem with using the full profile is that a logarithmic outer profile may be distorted near the bed by the form drag of protruding clasts, as noted in a flume study by Nowell & Church (1979) and quantified theoretically by Wiberg & Smith (1987). Heathershaw & Langhorne (1988) appealed to this effect to explain dog-leg logarithmic plots for profiles in tidal currents over an irregular gravel bed. Using only the outer profile in this situation gives a total stress that is relevant to the entrainment of the largest clasts but will overestimate the effective stress on smaller particles; retaining the near-bed points will give a compromise estimate.

Weighing up these conflicting problems, we consider it best to measure at as many heights as conveniently possible up to at least half depth, but look critically at the logarithmic plot and be prepared to omit suspect points especially near the bed. The estimated roughness height can be a useful guide: a value much higher or lower than at adjacent locations may reflect a zero plane displacement or other problem in the measurements, and will tend to be associated with over- or under-estimation respectively of the shear stress since both τ and z_0 increase with the regression coefficent b (in contrast, the mean velocity $<u>$ is much less sensitive to errors in the velocity profile). A typical value of z_0 for a poorly sorted gravel bed is around 0.1 D_{84} or 0.2 D_{50} (e.g. Bray 1980, Heathershaw & Langhorne 1988, Ferguson et al. 1989); we regard values of z_0/D_{50} outside the range 0.1-0.5 as suspicious.

If one is prepared to accept an empirical or theoretical scaling of z_0 in terms of the bed grain size distribution, a rapid estimate of shear stress is possible using a single velocity measurement (Dietrich & Whiting 1989). This does however require the assumption of a simple logarithmic velocity profile at least up to the height of the single measurement. Another alternative which we are investigating is to measure the turbulent Reynolds stress close to the bed using a two-component electromagnetic current meter.

2.4 Bedload transport measurements

In the last gravel-bed rivers workshop Davies (1987) expressed the view that "present methods are unable to obtain accurate and comprehensive bed load data by point sampling or sediment budgeting". Little has happened since then to alter this pessimistic view, though as discussed below we see more merit in sediment budgeting than Davies did.

The best-known problem with direct point sampling of bedload is the short-term variability in transport rates. Some of the evidence for this variability (e.g. Hubbell 1987) relates to flume experiments at high transport rates with migrating bedforms that are not always present in rivers, but the situation is no better at lower transport rates when entrainment clearly has a strong stochastic element. Einstein's model for this allows estimation of confidence intervals (McLean & Tassone 1987), but only in terms of the mean hop length and mean particle rest time, neither of which is likely to be known with any accuracy, especially in a spatial study involving single samples at different locations where the underlying mean transport rates are assumed to differ. Sampling for longer at each location evidently reduces the likely error, but means that fewer sites can be sampled within a given period of steady flow. Empirical investigations of the sampling variability of bedload transport under different conditions may provide useful guidelines (K. Prestegaard, pers. comm. 1990).

There are also some more practical, and soluble, problems in bedload sampling. In proglacial rivers with high loads of suspended sediment, the standard 0.25 mm mesh Helley-Smith sampler bag is apt to fill with fines before much gravel has been collected. If the suspended load is not of interest, the standard mesh should be replaced by a coarser one; we use 2 mm woven nylon. A double-size (152 mm) sampler improves the trap efficiency for coarse fractions, but in looking at spatial patterns the sampler must remain easily portable and not require any preparation of the bed at each measurement point. Samplers with tailfins are harder to handle because of the extra drag, and in some patterns the support frame for the tailfin tends to perch on protruding clasts allowing bedload to pass beneath the orifice.

If only the total bedload transport rate is of interest, not its grainsize distribution, samples can be weighed while still damp (with subsequent correction to dry weight) and returned immediately to the river; it is usually possible also to extract and measure the coarsest particle as an index of flow competence. If fractional transport rates and/or percentiles of the size distribution are required, it is necessary to bag and label the samples immediately for subsequent drying, sieving and weighing. In our work in meltwater rivers we have generally been able to do this the morning after sampling, drying samples in the sun and using a battery-powered top pan balance for weighing. If dimensionless transport rates or shear stresses are to be calculated a value will be needed for sediment density, so this should be determined, e.g. by weighing a selection of pebbles and the water they displace. In this paper we refer mainly to mass transport rates, denoted by q_s and measured in units such as kg $m^{-1}s^{-1}$.

There are of course several other ways of investigating bedload movement. Traps and tracers of various kinds are good choices for some purposes, but appear less useful than portable samplers if the aim is to map spatial patterns of bedload. Indirect estimation of bedload transport from channel change is discussed later.

3 METHODS FOR ANALYSING SPATIAL PATTERNS

Once spatially-distributed measurements of flow strength, bedload transport, and channel configuration and change have been made in a particular study reach it is necessary to interrelate the spatial patterns of each component of the system. Perhaps the most obvious way to do this is to plot the flow and transport measurements on maps, or against distance downstream if measurements have been restricted to just one or two transverse positions at each cross section. In Ashworth & Ferguson (1986) we used maps and verbal descriptions of the streamwise trends in shear stress and bedload transport rate to explain observed channel changes in the Lyngsdalselva, but on the basis of very limited data along the talweg only. We present below some new examples with fuller data, and discuss problems with this approach.
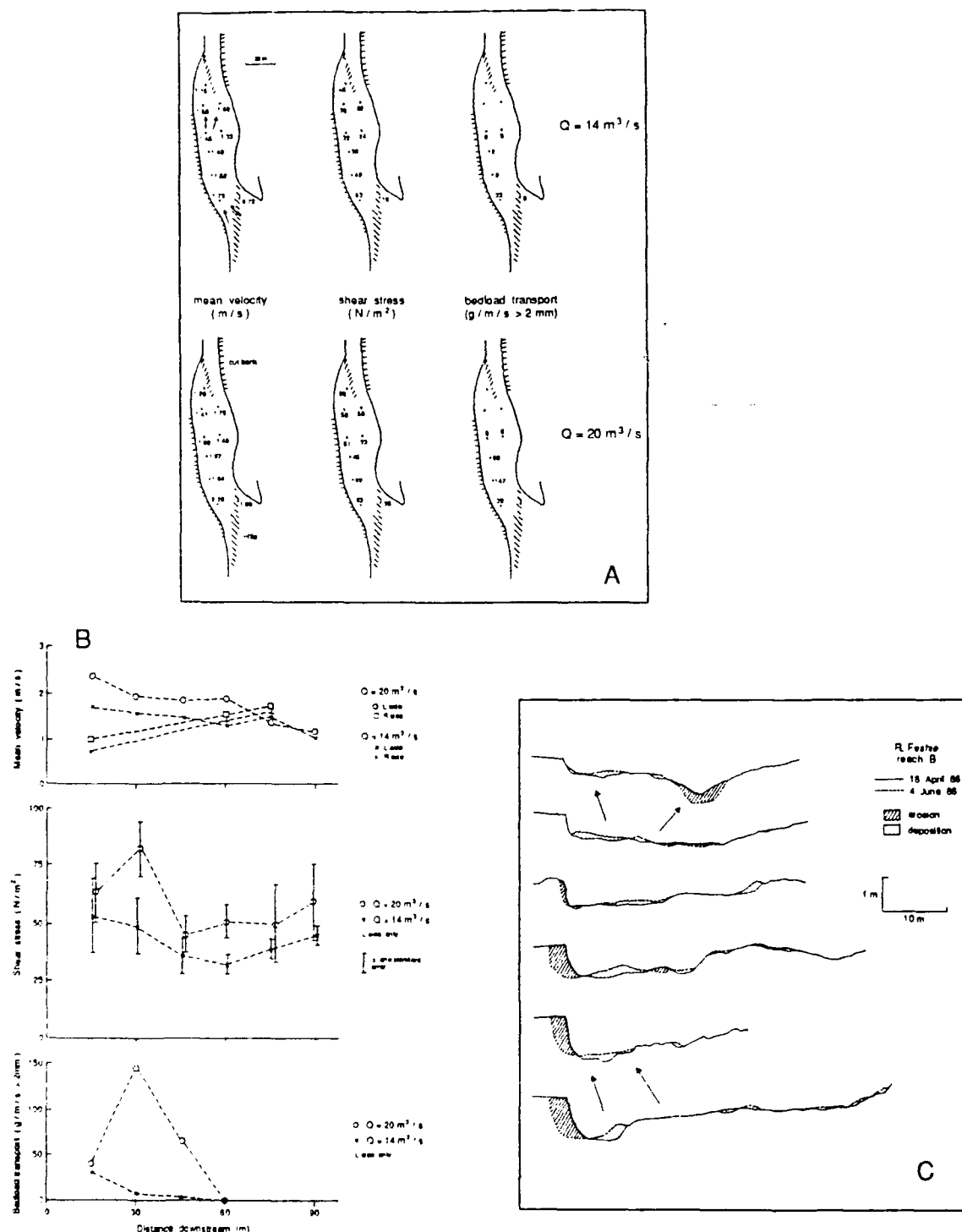
3.1 River Feshie, 1986

This study was made on a single day (2 May 1986) of diurnally varying snowmelt flow, with a peak discharge just greater than bankfull in the alternate-bar reach concerned. The general character of the river has been described before (Ferguson & Werritty 1983) and the process measurements were amongst those analysed by Ashworth & Ferguson (1989). They were made by two people over three hours, using tension from a rope to support the person wading since flow depth and mean velocity were relatively high (up to 0.8 m and 2.4 m s^{-1}). Measurements were made at one or two lateral positions on each of six cross sections about 15 m apart along the reach, and were repeated at two different flow stages.

The patterns of $<u>$, τ, and q_s are represented by numbers at appropriate locations on a map of the reach in Figure 2a. This form of mapping retains the maximum amount of information but does not give an immediate visual impression of the spatial patterns of the variables. On close inspection it can be seen that velocity and shear stress are high in the left-bank pool head, decrease down the pool along this bank towards the next bar front, but increase again towards the next pool head diagonally rightwards. The 40% increase in discharge during the afternoon was accompanied by an increase in velocity and stress at most locations, but no clear change in their spatial pattern. The plots of velocity and stress against distance along the reach in Figure 2b show this more clearly, and emphasise the out-of-phase pattern on opposite sides of the channel in this alternate-bar morphology.

Bedload transport was restricted to the points of higher velocity and stress, indicating that within-bank flows in this river are only just competent. This being so, the relatively small increase in velocity and shear stress as the discharge rose during the day led to a bigger relative increase in bedload transport, and a more pronounced spatial pattern: increasing (implying erosion) along the uppermost part of the pool, but thereafter decreasing (implying deposition) towards the bar front. This indicates a transfer of sediment from the poolhead to the pooltail and bar front. Figure 2c shows that this is just what happened in this reach over the melt season as a whole, and pebble-tracing studies also showed an accumulation of sediment towards the bar front. The process measurements are therefore consistent with observed channel change, and support the suggestions by Ferguson & Werritty (1983) about mechanisms of bar growth in this type of channel.

3.2 White River, 1986

This study involved intensive measurements by three people in this glacier-fed river over two weeks in late July and early August. As in our earlier Norwegian study an X-pattern (confluence and bifurcation) reach was selected, but in this case all parts of the reach were wadable for most or all of the time. Unfortunately the simple appearance of the reach was misleading, for a ribbon of mobile sand ran along the left side of the talweg and had a considerable effect on shear stress, roughness height, and gravel transport. We have discussed elsewhere (Ferguson et al. 1989) these effects and the statistical relationship between shear stress and bedload transport, but not the spatial patterns.

Figure 2: Measured mean velocity, shear stress, and bedload transport rate in an alternate-bar reach of the River Feshie, Scotland, at two discharge levels (Q) on 2 May 1986. In (a), dots mark locations of measurements, numbers by them indicate values, and no number indicates missing value. In (b) the data are plotted against distance down reach, with points close to the left bank distinguished from those further right. Channel change at the same cross-sections is shown in (c) for a six week period within which the highest discharge was on the day of the process measurements.
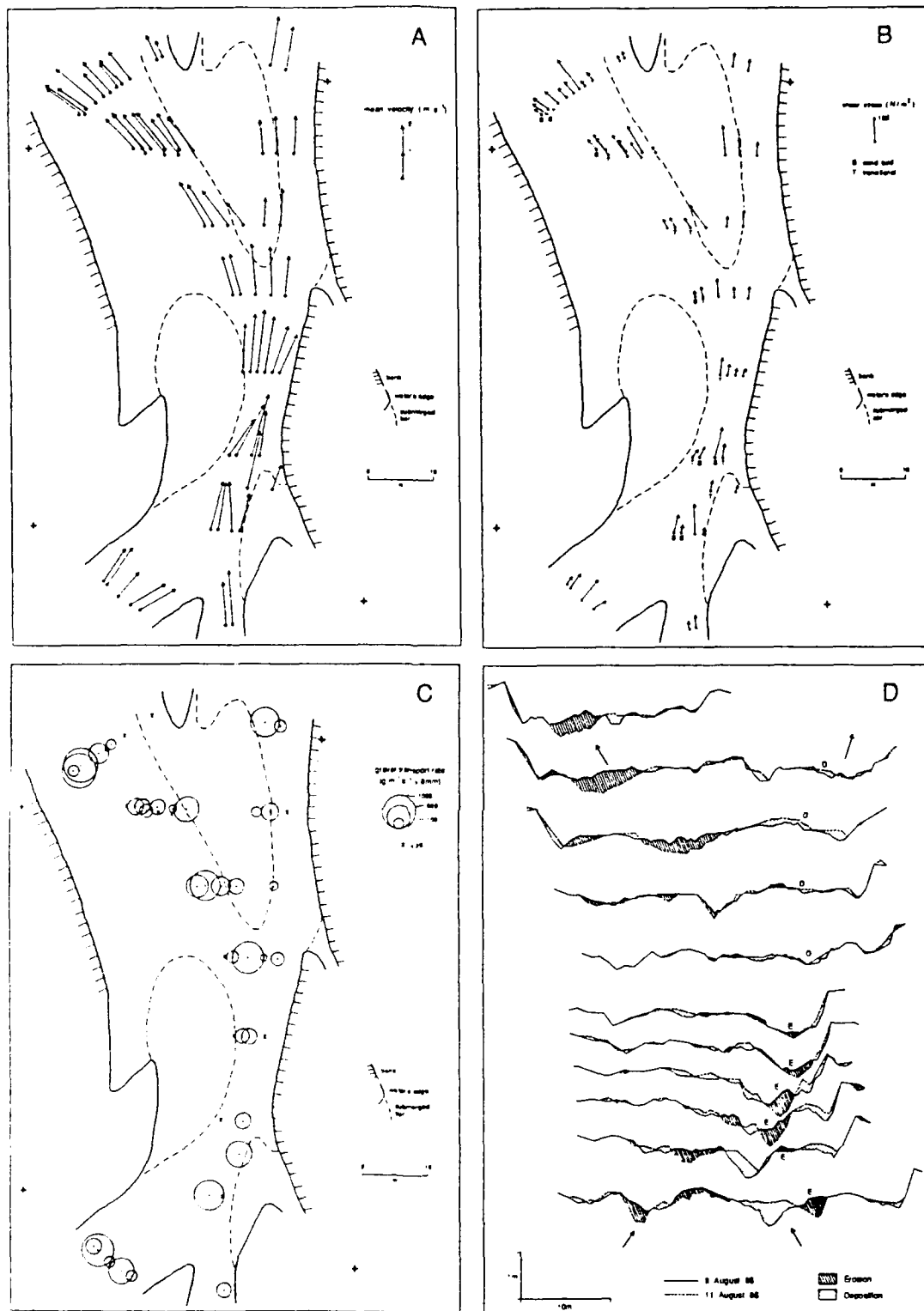
Figure 3: Spatial patterns in a braided reach of the White River, northwest US, during high meltwater flows. Measurements on evenings of 8/9 August 1986 are mapped as follows: (a) magnitude of mean velocity, plotted in direction of surface velocity; (b) shear stress, plotted in direction of surface velocity, with non-gravel bed types indicated and dashed arrow where standard error of stress estimate is large; (c) gravel transport rate. Channel change over 8-11 August is shown in (d), with the main cell of erosion and deposition labelled by E and D.

Figure 3 illustrates the patterns of mean velocity, shear stress, bedload transport rate, and channel change in the last few days of the study, when discharge and activity were greatest. Maps with vector arrows or proportional circles for measured quantities are used for maximum visual impression of spatial patterns. The convergence and acceleration into

and past the confluence, and divergence and deceleration in the ensuing bifurcation zone, are apparent in the velocity map but not clearly reflected in the maps of shear stress or transport rate. This is partly because of the complication introduced by the sand ribbon, but no doubt also because sampling and measurement errors in bedload and shear stress are higher than those in mean velocity. We discuss later some ways of dealing with this noise in the spatial patterns.

The channel change diagram (Figure 3d) is also complicated, but less so if changes originating upstream of the study reach are ignored. The main changes that can be compared with our process measurements are erosion in the confluence (zone E in the diagram) with deposition further downstream on the medial bar and right distributary (zone D), and erosion in the left distributary, presumably with deposition further downstream. Both erosion zones approximately coincide with zones of flow acceleration, as in the marginally-competent meltwater flows of the Norwegian and Scottish case studies.

3.3 Sunwapta River, 1989

This study involved a bigger team than the previous ones, with up to three current meters and three bedload samplers simultaneously in use in an attempt to obtain detailed and reliable spatial patterns. Velocity profiles were measured at up to 11 heights, and bedload was sampled for at least 5 minutes, to reduce sampling error. The reach was a simple chute and bifurcation, with finer gravel (surface D_{50} of 20-30 mm) than in the other rivers studied, and measurements were made over a week of high meltwater flows in July-August. The results will be described more fully elsewhere but an example is given in Figure 4. The divergence of bedload transport and the general pattern of erosion in the constriction, deposition downstream of it, are clearly revealed. But is such a map reliable, even when based on such labour-intensive methods? It undoubtedly contains substantial noise as a result of sampling error. In the final sections of this contribution we therefore attempt to assess this error and compare locally-averaged or totalled rates with independent estimates based on observed channel change.
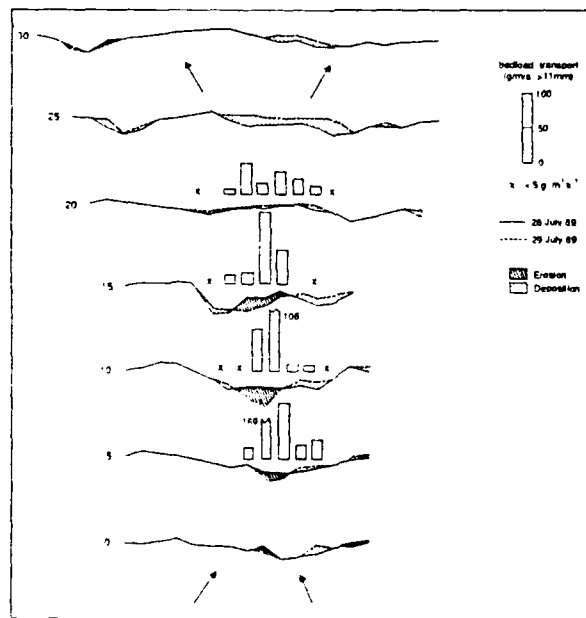
Figure 4: Channel changes during 28-29 July 1989 in a reach of the braided Sunwapta River, Canada, and bedload transport rates sampled during high meltwater flow on the evening of 28 July 1989.

3.4 Local averaging of measurements

A standard way of dealing with sampling error in independent measurements is to average them, since the sampling variance of their mean is less than that of the individual values. In looking at nonuniform patterns of shear stress or bedload transport in gravel-bed rivers we are naturally keen to retain as much spatial detail as possible, but if the sampling noise overwhelms the spatial trend it may be advantageous to do some averaging of the raw measurements.

If measurements have been made at several locations along each cross section of a reach, one possibility is to calculate a mean value for each section. In the case of bedload transport, a total value for the section (Q_S as opposed to q_S) also has physical meaning. Either Q_S or a width-averaged q_S will have greater precision (in percentage terms) than the individual q_S measurements; totals or means can then be plotted against distance downstream in the hope that streamwise trends emerge more reliably than from the raw data.

This is illustrated in Figure 5 using data from two of the case studies discussed above. In Figure 5a the White River bedload measurements of Figure 3c are averaged over zones of one to three adjacent cross sections, the zone size being varied to achieve a minimum of n = 5 samples in each zone. Standard errors of the mean transport rates are estimated in the ordinary way from the sample size and standard deviation. They are not exact, because we are dealing with small and skewed samples from what is likely to be a non-homogeneous population. In this example the spread of individual transport rates is very high and the sample sizes are low, so the zonal means still have standard errors too large for differences between zones to be significant. The averaging thus fails to define an unambiguous spatial pattern.

In Figure 5b the cross-section totals of bedload in the Sunwapta River on four different days, including the data mapped in Figure 4, are plotted against distance downstream. The standard errors are omitted for clarity but are again high, in the range 30%-60% of the estimated totals. Many of the apparent increases or decreases in transport from one section to the next could therefore be artefacts of sampling error, though the differences between dates are more convincing.

We discuss later how the spatially-averaged bedload measurements illustrated in Figure 5 compare with inferences from observed channel changes in the same reaches.
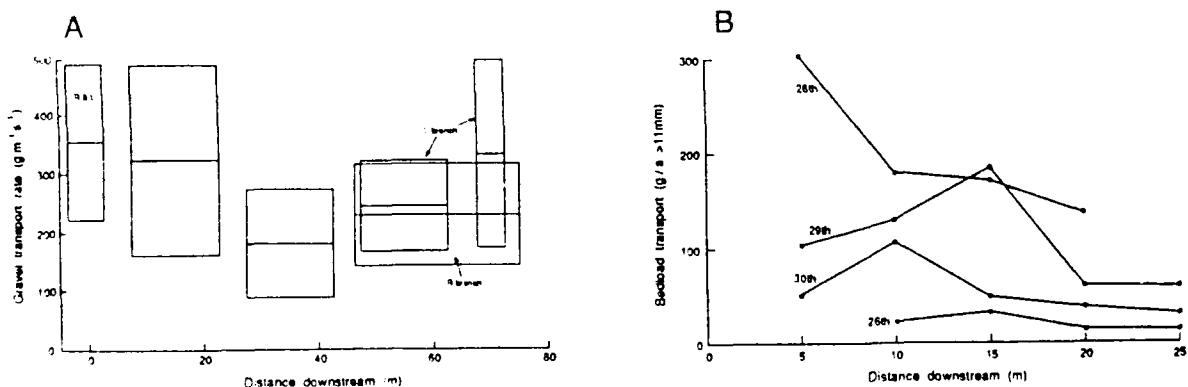
Figure 5: Examples of downstream patterns of local averages and lateral totals of sampled bedload transport rates: (a) White River, 8/9 August 1986, (b) Sunwapta River, 28 July 1989. In (a), each box extends one standard error above and below the mean transport rate for the cross-sections concerned.

3.5 Inferring bedload transport rates from channel change

The possibility of estimating a spatially and temporally averaged mean sediment transport rate from observed channel changes over a period has been applied many times to bedform migration. Neill (1987) extended it to meander migration over several years, and Church et. al. (1987) and McLean (1990) have used it in multi-year comparisons of long reaches of large, low-sinuosity rivers. Our application is at far shorter time and space scales but uses the same principle. The general idea is that the volume V of sediment removed, divided by the average distance L that it travels before being deposited on the next point bar and by the time interval T over which V is measured, gives the mean volumetric flux through the reach. This method will underestimate the true mean transport rate if alternate scour and fill cause net channel change between surveys to fall short of gross change. Our White River and Sunwapta investigations show that channel change at a point can reverse within a few days, and this is supported by Laronne & Duncan's (1989) work using scour chains in a New Zealand river. Underestimation for this reason is likely to increase with the time interval T between surveys. Another source of underestimation is that some bedload may travel nonstop through the reach; this is decreasingly likely as reach length L increases, and the available evidence from pebble tracing studies by several researchers is that few pebbles travel further than the next riffle during one period of competent flow.

In braided channels the loci of erosion and deposition are less predictable than in meandering streams, and even over quite short timespans (e.g. 3 days in Figure 3d) there may be a complicated mosaic of net erosion and net deposition through the combined effects of channel scour, fill, migration and avulsion. Resurveys over short periods may however reveal distinct zones of erosion and deposition which can be matched up as transport cells in the same way as in a meander bend. As already illustrated by our case studies, just-competent flows tend to move sediment from pools and confluences to bars and bifurcations. If in such a situation the volumes of erosion and deposition are similar, so that sediment can be inferred to move from one to the other with a well defined mean travel distance, a reasonable estimate of the mean transport rate should be possible.

It is necessary first to compute volumes of erosion or deposition between successive surveys. This is done by superimposing the plotted surveys at times t_1 and t_2 for cross-sections $i = 1, 2, ..$ at distances xi along the channel, delineating the zone of (say) erosion (as labelled by E in Figure 3d), and determining the width wi and mean depth di of erosion at each cross-section. The total volume of erosion within the zone is then obtained by summing increments between adjacent sections, calculated as follows on the assumption that w and d vary linearly with distance:

$$Vi = (x_i - x_{i-1})(2w_i d_i + w_i d_{i-1} + w_{i-1}d_i + 2w_{i-1}d_{i-1})/6 \qquad (7)$$

Deposition volumes are quantified in the same way. Equation (7) simplifies if either w or w/d is constant between adjacent sections. If the sections are not parallel, appropriate corrections can be made to x and w. A degree of subjectivity is involved in deciding at just what distance w and d go to zero at the ends of a zone, but the volume increments here are small.

Once a volume V_e of erosion over a time interval T has been matched up with a similar volume V_d of deposition a short way downstream, the mean transport rate q_s in $m^3 m^{-1} day^{-1}$ or similar units can be estimated as

$$q_s = (V_e/A_e).(L/T) \qquad (8)$$

where A_e is the plan area of erosion and L is the mean travel distance of particles moving from the erosion zone to the deposition zone. L can be estimated as the difference in position along the channel of the centroids of V_e and V_d, calculated as $\Sigma_i(x_i w_i d_i)/\Sigma_i(w_i d_i)$. Conversion of the transport rate estimated using equation (8) to mass units such as $kg\ m^{-1}s^{-1}$, as for point samples, requires multiplication by an assumed bulk density and by the fraction of the interval T between surveys during which active transport is thought to have occurred.

We have applied this method to the main erosion/deposition cell in our White River study reach (Figure 3d, zones E and D). Surveyed changes over 3 days amounted to 19.6 m^3 of erosion from an

area of 131 m^2, and 18.3 m^3 of deposition at a centroid-to-centroid distance of 49 m downstream. Substitution of these values in equation (8) gives a mean transport rate of 2.4 m^{3}m^{-1}day^{-1}, or about 0.18 kg m^{-1}s^{-1} assuming a bulk density of 1600 kg m^{-3} and 6 hours per day of active transport. This agrees well with the mean of 0.17± 0.06 kg m^{-1}s^{-1} (> 2 mm) for 12 Helley-Smith samples within the erosion/deposition cell in the relevant period. A similar calculation from the 4.0 m^3 of erosion and 4.1 m^3 of deposition during the day of maximum change in the Sunwapta River (Figure 4) gives an estimated mean transport rate of 1.5 m3m^{-1}day^{-1} or 0.11 kg m^{-1}s^{-1}. This is in reasonable agreement with the mean of 0.08± 0.03 kg m^{-1}s^{-1} (> 2 mm) based on Helley-Smith sampling. On the evidence of these comparisons, it appears that space-time averages of bedload transport can be estimated from channel change at least as reliably as by direct sampling.

3.6 Within-reach sediment budgeting

Whether or not clear erosion/deposition cells can be identified, surveyed channel changes at a series of cross-sections within a reach can be used to infer the rate of change of bedload transport along the channel by using the sediment continuity equation. In finite difference form this is

$$\Delta q_s/\Delta x = -V/(w.\Delta x.T) \qquad (9a)$$

or

$$\Delta Q_s/\Delta x = -V/(\Delta x.T) \qquad (9b)$$

where V is volume of erosion (negative) or deposition (positive) over channel length Δx and time T, w is width of erosion or deposition, q_s is volumetric transport rate per unit width and time and Q_s is totalled over w and averaged over T. If V and w are known for each of a series of cross-sections, the transport rate need only be known at one section for equation (9) to define its value at all other sections in the reach. If no direct measurements of bedload transport are available all that can be done is to set a lower bound by the requirement of non-negative transport at all sections, as done by Griffiths, (1979) in a pioneering application of this approach to surveyed changes over many km and years in a New Zealand river. In some circumstances a downstream boundary condition of zero gravel transport can be assumed, as in McLean's (1990) budget for the lowermost gravel-bed reach of the Fraser River, Canada, using surveyed changes over 32 years at sections 2 km apart.

Our application of the sediment budget approach is at a quite different scale (steps of 1-3 days and 5-10 m) and the nearest we have to a boundary condition is the mean transport rate in our spatially-distributed Helley-Smith samples. We turn again to the White River and Sunwapta case studies to illustrate the method. Changes over 3 days in the long profile of a 5 m wide talweg strip in the White River are shown in Figure 6a. The pattern of channel change implies a downstream decrease in transport rate into the aggrading head of the confluence zone, an increase past sections 20 and 25 which scoured during this period, then a decrease along the aggrading right distributary but an increase along the degrading left distributary. The volumes of erosion or deposition between successive pairs of sections were estimated using equation (8) and converted using equation (9) to changes in mean transport rate along the channel. The inferred downstream variation in mean transport rate is plotted in Figure 6b, with the general level of transport matched to the 0.27 kg m^{-1}s^{-1} mean of the 47 Helley-Smith samples collected from various points along this strip on the evenings of 8 and 9 August. As already noted the means of directly-sampled transport rates in the six zones distinguished in Figures 5a and 6b have large standard errors, so that downstream changes are not defined unambiguously. They can best be described as not inconsistent with the predictions from channel change. The predicted curve in Figure 6b lies within one standard error of each of the six zone means, and the locations of the highest predicted and measured mean transport rates (at either end of the reach) agree, but the lowest measured mean transport rate is not at the predicted location.

A similar comparison of cross-section total transport rates (Q_s) in the Sunwapta, as estimated from point sampling (Figures 4 and 5b) and from channel change in a 7 m wide central strip, is illustrated in Figures 6c and 6d. The channel aggraded throughout the reach during 22-28 July, implying a downstream decrease in transport rate, but during 28-29 July and 29-30 July there was proximal erosion and distal deposition, which implies a maximum transport rate in mid-reach. The direct samples agree with these qualitative patterns on the 26th and 29th, but not on the 28th. The quantitative agreement is good on the 26th, and fair on the 28th and 29th taking into account the large standard errors of the Helley-Smith estimates of Q_s.
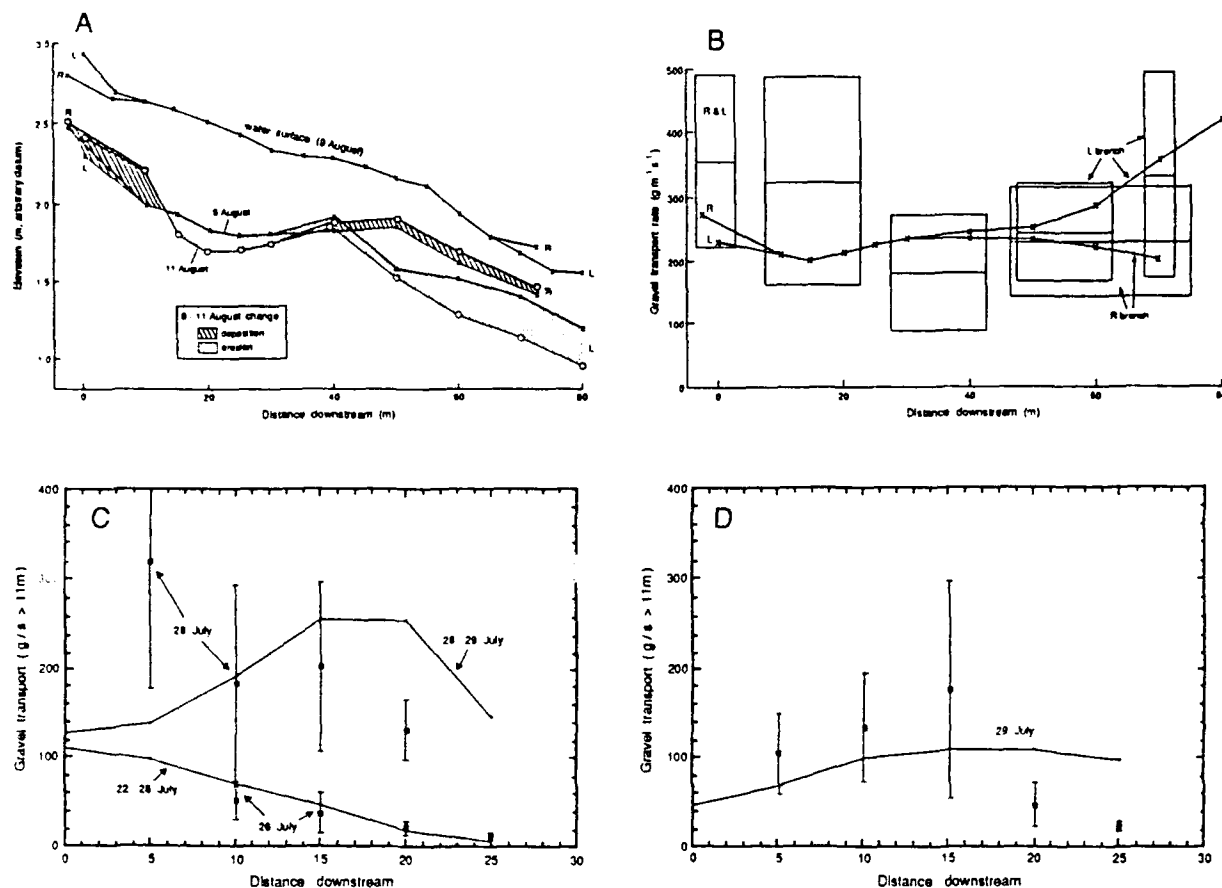
Figure 6: Comparisons of downstream variation in mean bedload transport rate as estimated by point sampling and from channel change: (a) aggradation and degradation of 5 m wide talweg strip of the White River; (b) mean transport rates inferred from (a) (continuous lines) compared with those measured directly (boxes as in Figure 5a); (c and d) width-integrated transport rates in the Sunwapta River as inferred from aggradation or degradation of 7 m wide central strip (continuous lines) and from point sampling (symbols with standard error bars).

We feel that these comparisons cast further doubt on the reliability of Helley-Smith sampling of bedload, rather than invalidating the idea of inferring downstream changes in bedload transport from intensive surveys of channel change. A further possibility in future studies would be to estimate Q_S at a baseline cross section by intensive Helley-Smith sampling across the active width, then use frequent resurveys and equation (9) to calculate transport rates progressively further upstream and downstream. This should give more reliable estimates of cross-section average transport rates along the reach than would be obtained by the same total sampling effort distributed between all the sections, but at the expense of losing information on lateral variations in transport and on grainsize distributions of transport away from the baseline section.

4 CONCLUSIONS

We have described some of the measurement techniques and methods of data presentation and analysis that we have used in the last few years in an attempt to understand channel change in braided rivers in terms of nonuniform processes. Our experience and results show that spatially distributed measurement is feasible, but not easy, and that gaining accurate and reliable measurements is even more difficult. The weakest link is sampling of bedload. The numerical examples discussed here suggest that the relatively easy and accurate measurement of channel change provides a partial alternative, since local mean values of bedload transport can be inferred from volumes of erosion or deposition. However, this is at the price of losing detail of spatial patterns in bedload transport.

5 ACKNOWLEDGEMENTS

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1

BEDLOAD TRANSPORT PATH AND GRAVEL BAR FORMATION

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Gravel Bar Formation and Reworking

New, central bars are formed when bed areas are inundated and, with few exceptions (Hein 1974; Smith, 1974), cannot be observed in the field due to the turbidity of the flow. Bar reworking also occurs at flow stages below bankful, when the river bed can be seen, and because the changes are often sufficiently slow to clarify the mechanisms of reworking. Accordingly, most of our understanding concerning bar formation derives from flume studies where suspended sediment has been excluded (Leopold and Wolman, 1957; Ashmore, 1982; 1985; 1987, Hoey, 1989), whereas most of the observations on bar reworking are based on field studies (Fahnestock, 1963; Hitchcock, 1977; Werritty and Ferguson, 1980; Ferguson and Werritty, 1983; Ashworth and Ferguson, 1986; Ashworth, 1987). The difficulty of observing the formation of protoype bars explains why some hypotheses concerning their formation have been highly speculative (Rundle, 1985).

Central bars may be formed by deposition at sites of flow divergence. These may be a new 'nucleus' of coarser gravel (Leopold and Wolman, 1957) or an advancing single or multiple sediment lobes (Ashmore, 1982). Such depositional units have been coined a variety of terms (Smith, 1978; Church and Jones, 1982), but are now often termed 'unit bars' (Smith, 1974). Central bars may also appear due to erosion of existing bed areas. Three erosional mechanisms have been proposed (Ashmore, 1985). Central bars may be formed during high stages of flow due to meander cutoff (Lewin, 1976; Werritty and Ferguson, 1980; Ferguson and Werritty, 1983) and due to alternate bar cutoff (Ashmore, 1982). A second erosional mechanism is dissection of gravel sheets at failing stage (Hein, 1974), and a third mechanism is the formation of miniature 'protochannels', which have been observed to develop on a steep, plane bed (Moss, Walker and Hutka, 1980; Moss, Green and Hutka, 1982).

The conditions favouring one or more of the above mentioned mechanisms of central bar formation have not been determined. The 'nucleus' bar can be formed due to the initial deposition of coarser bed-material, favouring it for gravel rather than sand bed streams. Its preferred formation downstream of a confluence has been observed in a model (Ashmore, 1982) and it is suggested from painted gravel transport paths in two Scottish wandering, gravel bed rivers (Ashworth, 1987). Recent field documentation demonstrates that flow divergence, lower flow velocities and lower shear stresses indeed occur simultaneously downstream of a confluence (Ashworth, 1987). Although advancing sediment lobes or gravel sheets have been observed in the field and laboratory, it is yet to be demonstrated whether they are as common in a wide, gravel bed rivers as they are in some flumes relating 1-2 confluences (Ashmore, 1982; Davies and Lee, 1988).

Reworking of gravel bars involves primarily bank erosion resulting in a single, supposedly more sinuous pattern (Werritty and Ferguson, 19870; Ashworth, 1987). This erosion occurs with concomittant deposition of similar quantities of bed material in the opposite bank of wandering gravel bed rivers, an observation particularly well documented in reach D of the Dubhaig (Ashworth, 1987). Reworking also includes channel 'switching' and resultant infilling of previous anabrances as well as the deposition of new gravel sheets (Werritty and Ferguson, 1980).

Gravel tracers may be used to infer modes of bar formation. For instance, Brewster (1986, in Ashworth, 1987) showed that a bar head of the River Feshie may develop by receiving sediment derived far upstream, rather than from the immediate surroundings. Bed material appears to be prteferentially deposited on the stoss of point bars (reaches A and C of the Dubhaig) and of longitudinal (compound?) bars (reaches B of the Dubhaig and Feshie; see, Ashworth, 1987). The use of painted gravel is, however, difficult in very active braided rivers due to burial of most of the tracers (Ferguson and Ashworth, 1986).

Additional prototype observations are required to evaluate the relative roles of these mechanisms of bar formation and reworking. The purpose of this study is to determine modes of bar reworking and formation in a wide, braided river, and to compare them to those in a lower, narrower alternate bar reach. Specifically, we assume that bed material transport paths should be strongly controlled by flow divergence and increased friction around an emerging nucleus or a point bar, whereas paths would in large part be parallel to flow direction if sediment accumulation occurs around advancing lobes or sheets..

Study Area

The North Branch Ashburton River drains 300 km^2 of steep greywacke terrain in the Southern Alps, South Island of New Zealand. The lower catchment, from the gorge to the confluence with the South Branch, is an 80 km^2 segment of the Canterbury Plains, comprising a sequence of Pleistocene and recent alluvial lobes or fans and terraces of gravel and sand (Mabin, 1980). Mean annual precipitation in the North Branch Catchment is 1400 mm. Estimated mean annual and 50 yr flood peaks at the gorge are 180 and 570 m^3 s^{-1}.

There are three types of channels in the North Branch: a mostly rock-confined, shallow alluvial channel upstream of the gorge, a wide, braided channel between the gorge and Thompson's Track Bridge and a narrower, alternate bar channel extending 21 km to the confluence with the South Branch.

An upper, braided reach and a lower, alternate bar reach were selected (Fig.1). The upper reach is located between the Rangitata Diversion Race (RDR) and Thompson's Track Bridge. It is about 220m wide, wider than most other reaches of the river. Based on older aerial photographs we have determined that willow planting has confined it from an earlier width of about 500 m. Mapping based on more recent aerial photographs shows that this reach has well developed longitudinal, medial bars, often with merely 2 primary anabraches that are completely detached from the banks during low flow conditions (Fig. 2a). The river bed comprises active bed areas and a series of sequentially higher and older surfaces or bar coplexes covered by dense grass and shrub colonies (Laronne and Duncan, 1989a). These surfaces are dissected by a series of distributary channels that are dry during base flow conditions. The length of reach surveyed was 700 m with an average bed slope of 1.15 percent.

The lower, narrower study reach lies between Thompson's Track Bridge and Shearer's crossing, within Bland's Reach, which has been aggrading since 1937. A wedge of bed material 4 m thick has been accumulating in this reach at an average rate of 5.8 cm yr^{-1} and as much as 11.4 cm yr^{-1}. The river bed is comprised of alternate bars, it is about 85 m wide, straight, and is confined by willows and high banks for floodplain protection (Fig. 2b). In relation to their original widths, the lower reach has been confined slightly less than the upper reach. The length of reach surveyed for this study was 750 m with an average bed slope of 0.73 percent.

Methods.

The South Canterbury Catchment Board's Old Weir recording site at the gorge provided continuous water level and discharge data. In addition, three temporary installations of 3 m Foxboros supplied gauge records at the Rangitata Diversion Race above the upper reach, at Thompson's Track Bridge below it and at Digby's Bridge further down river (Fig. 1). Several stage staffs were also used. Flow velocity was measured using Gurley current meters at 0.6 flow depth by wading or jet boating. The location along the cross section was measured using a

tight tag line or by surveying. Water surface elevation was determined while measuring flow velocity during two instances using a tachometer or an electronic distance meter (EDM).

The morphoplogical characteristics and bed changes from the gorge to the confluence with the South Branch were documented along 40 km in 7 aerial photography strips taken between 9 November, 1985 to 3 April, 1986. Photographs were taken from a light aircraft at ca. 1500 m above sea level using 35 mm cameras with a focal length of 55 mm (35 mm in one instance). Photographs were near vertical with 5-70 percent overlap. Average scale, calculated from distances between benchmarks, is accurate to +/- 5 percent. The enlarged photographs were used to prepare 1:1000 maps of the study reaches.

Altogether 17 cross sections in the upper reach and 24 in the lower reach were surveyed with a tachometer or EDM before event 1, as well as after events 1,2,3 and 6. The survey parties paid particlar attention to significant breaks in slope, to channel banks, to the average bed elevation between or on top of bed particles and to survey all the minor channels. The survey data was transferred either manually or else directly from the EDM electronic field 'notebook' to a programme that allows drawing the cross sections as well as calculation of volume changes between sections (Ibbitt and Duncan, 1985).

Surface texture was determined using a number by number method (Wolman, 1954) described by Mosley (1982), also identifying particles in ranges 2-8 mm and < 2mm.. Surface samples consisted of 100 particles, except bar transects which involved a double sample size. In addition, 200 samples of subsurface bed material were taken from scour chain holes (see below). Each sample weighed about 40 kg and was taken from a depth of 0.2-0.8 m. Boulders (perhaps > D_{95}) were unfortunately discarded in the upper reach because they were dangerously large to deal with during the operation with the excavator. Samples from the upper and lower reaches were separately combined in two ca. 4 T heaps, mixed, sampled as required for such coarse material (Church, et al., 1987) and sieved separately.

Magnetic tracer particles, approximately in the D_{50}-D_{95} size range of the subsurface, were used to evaluate the direction and quantity of bedload (Laronne et al., 1986) and its relationship to bedform development. The original method (Hassan, et al., 1985) was slightly changed to enable detection to a depth of 0.9 m below the surface. This was accomplished by inserting two magnets into every particle. The tracers were sprayed with abrasion resistant yellow paint, numbered on both sides, weighed and their size was measured. Particles were randomly chosen and placed at scour chain positions 30 cm below the surface as well as several on the surface. A total of 534 tracers were placed in the upper reach and 436 in the lower reach.

On location, the depth of burial was measured from the surface to the top of the tracer and location was determined by surveying or tape. Following recovery, the particles were placed in the position where they had been found.

Bed material size.

Spatial variations in the texture of surface material are inherent to gravel beds (e.g., Laronne and Carson, 1976; Mosley and Tindale, 1985; Church et al., 1987). We identified five large, distinct textural assemblages: the surface of main anabranches, the average bar surface or bar transect, patches of coarser and finer sediment on bar surfaces and the subsurface (Table 1).

The median particle size on bars (D_{50}) is 25 percent smaller in the narrow reach, or 30.6 percent smaller when comparing the median from a distribution that excludes sand. D_{84} and D_{95} decrease even more between the reaches, implying selective deposition and some wear downstream. The surface of all the active channels (i.e., excluding the plugged and infilled ones) is coarser than that of adjacent bar areas as it is defined .... than 85 mm in the lower reach and smaller than 70 mm in the upper reach. This conforms to many observations of coarse surface layers termed pavement or armour (e.g., Church, 1972; Profitt, 1980; Parker and Klingeman, 1982; Gomez 1984). The banks of anabranches are often as coarse or coarser than the respective armoured beds. The average texture of the bar surface is similar to that of the subsurface. The median of the subsurface appears to be similar in both reaches, but this is misleading because excluded boulders were only present in the subsurface of the upper reach.

Table 1. _Bed material size statistics (mm) from different bed areas._

	Including sand upper reach	Including sand lower reach	Excluding sand upper reach	Excluding sand lower reach
Bar traverse				
D_{50}	28	21	34.3	23.8
D_{84}	74	52	93.7	49.9
D_{95}	137	74	142	81.6
Coarser patches				
D_{50}	34	39	36.8	38.3
D_{84}	82	69	78.8	71.5
D_{95}	128	95	128	104
Main channel				
D_{50}	65	39	64.9	54.9
D_{84}	119	74	119	68.6
D_{95}	181	99	181	99
Finer patches				
D_{50}	23	17	23.4	17.1
D_{84}	47	37	47.8	35.0
D_{95}	74	56	76.1	49.5
Subsurface				
D_{50}	21*	21	25*	25
D_{84}	64*	64	75*	69
D_{95}	147*	112	181*	~104

*Had boulders been included, the entire distribution would have shifted considerably; D_{95} (incl. sand) of the subsurface may have been ~250 mm.

Flow events and hydraulics

Table 2 summarises characteristics of flow events 1-6. Flow event 2 consisted of a discharge of 6.5 m^3 s-1 plus a 24 m^3 s-1 constant inflow from the RDR, which discharges into the river below the Old Weir site. The RDR discharged water into the N. Branch about weekly, usually less than 10 m^3 s-1. There was a 24 percent transmission loss between the study reaches during the first, smallest event but the losses were insignificant in all succeeding events (Laronne and Duncan, 1990). The first three events were relatively minor, event 3 being a compound hydrograph comprised of two individual, almost equal rises within three days. Based on the flood frequency curve of the river (Laronne and Duncan, 1990), events 1, 2 and 3 recur on average more than five times per year, and the recurrence intervals of events 4-6 are 6 months, 4 months and 3 years, respectively.

The hydraulic character of flow in the study reaches, as well as in an intermediate reach at Thompson's Track Bridge, are presented in Table 3. Additional hydraulic measurements are presented elsewhere (Laronne and Duncan, 1990). Flow direction is not listed in the table, but critical observation of flow direction at the water surface and 0.6 depth by the gauging parties showed that in all instances except event 1, flow direction in the lower reach was essentialy

downchannel, parallel to the general alignment of the reach, and it was parallel to anabranch alignments in the upper reach. The upper reach contained many areas that were neither boatable (too shallow) nor wadeable (too fast) during event 5, an experience shared by Davoren and Mosley (1986), so that additional attempts to gauge it were relinquished. We attempted to gauge the lower reach also during the larger event 6, but standing waves 0.9 m high and shallow depth on bars allowed only partial gauging.

Table 2. *Hydrograph characteristicxs of North Branch Ashburton River at Old Weir.*

Flow event	Date	Peak discharge $(m^3 s^{-1})$	Time to peak (hr)	Bedload Duration[c] (hr)
1	2.12.85	18	16	37
2	11-12.12.85[a]	31	1	14
3	22.12.85	25	13	14
	25-26.12.85	23	38	74
4	16-17.02.86	58	14	47
5	25.02.86	47	15	66
6	13-14.03.86	173[b]	32	155

[a] At the upper study reach, flow due to release from RDR.

[b] At least 207 $m^3 s^{-1}$ at Thompson's Track Bridge.

[c] Duration of flow above 12 $m^3 s^{-1}$, which is the estimated threshold discharge for bedload transport in the lower reach; see Laronne and Duncan (1990) for details.

The water width in the upper reach included many small channels, which complicates analysis of average hydraulic values. It demonstrates the inapplicability of using average sectional values to multiple thread rivers, where only part of a section may be actively transporting sediment (Carson, 1986). For instance, in the upper reach during event 1 the hydraulic radius, computed for all the channels, increased only slightly from the bar head and bar centre (U1 and U4) to the bar tail (U6), where a deep confluence scour pool contained most of the discharge. The depth at the confluence pool was 2.3 times larger than that at the anabrach thalwegs further upstream and flow velocity was also considerably larger, differences masked in average sectional values. Self evidently, the average unit power and shear stress in the pool were considerably larger than the values in Table 3. A similar situation occurred during event 2, at which time the scour pool had undergone incipient filling.

The number of channels conveying water did not alter appreciably as long as bars were not submerged. Because both reaches have been confined, the number of channels decreased to a single, submerged bed during small events (such as event 3) in the lower reach and during larger ones in the upper reach. Hence, the extent of braiding is not constant in confined reaches as it is in natural, braided ones (Mosley, 1982). Average and maximum hydraulic values during events 1, 2 and 5 increased with increase in discharge at both reaches. These responded in a different manner: depth increased much more in the narrow reach whereas velocity increased more in the upper reach. At Thompsons Track Bridge the response was intermediate, noting that the data for the bridge derives from the large event. A marked decrease occured in the spatial gradients of maximum depth, velocity and Fr within the upper reach as discharge increased. These gradients were attenuated in the lower reach during event 1 due to transmission losses.

Table 2. *Hydraulic character of selected North Branch Ashburton River measured cross sections at the study reaches and Thompson's Track bridge.*

section	# of active channels	Q (m3/s)	w (m)	R (m)	d_max (m)	d_umax (m)	u (m/s)	u_max (m/s)	Fr (-)	Fr_umax (-)	Fr_rmax (-)	Ω (W/m)	ω_w (W/m²)	ω (cm/s)	τ (N/m²)
U1	4	18.3	77.0	0.25	0.65	0.5	0.93	1.6	0.59	0.81	0.81	200	26.8	1.09	28.2
U4	3	19.0	76.1	0.25	0.64	0.5	1.11	1.8	0.71	0.81	0.81	210	28.2	1.15	28.2
U6	4	17.8	43.9*	0.31	1.50	1.5	1.29	2.6	0.74	0.68	1.00	200	45.7	1.50	35.0
U2	2*	31.7	35.0*	0.52	1.00	1.0	1.72	2.6	0.76	0.88	0.88	350	102	2.00	58.6
U4	3	31.4	77.9	0.32	0.67	0.6	1.25	2.4	0.71	0.88	0.83	350	45.5	1.45	36.1
U6	6	30.1	86.7	0.25	1.30	1.3	1.38	2.9	0.88	0.81	0.86	330	39.2	1.60	28.2
U2	3	51.0	71.4	0.37	1.17	0.7	1.94	1.7	1.02	1.41	1.41	570	80.6	2.22	41.7
TT1	1	207	99.5	0.72	1.48	0.6	2.86	4.4	1.08	1.18	1.90	1800	182	2.51	62.8
TT3	2	40.7	65.6	0.46	0.80	0.7	1.33	2.2	0.38	0.84	1.01	350	54.2	1.20	40.2
L2	3	12.1	34.1	0.29	0.70	0.7	1.20	.8	0.71	0.69	0.69	87	25.4	0.89	20.8
L4	1	14.1	16.4	0.54	0.70	0.7	1.57	2.1	0.73	0.71	0.77	100	61.6	1.16	38.7
L6	1	14.3	40.5	0.43	0.75	0.5	0.82	.7	0.65	0.70	0.70	100	25.3	0.60	30.8
L2	2	30.1	37.2	0.41	0.76	0.7	1.33	2.3	0.42	0.88	0.91	210	24.7	0.62	29.4
L4	3	30.0	50.5	0.37	0.85	0.8	1.22	2.0	0.70	0.76	0.99	210	35.5	0.98	26.5
L6	1	34.2	36.0	0.32	1.60	1.6	1.22	2.1	0.69	0.81	1.14	240	28.5	0.91	22.9
L6	5	16.9p	10.8+	0.30	0.75+		1.37		0.80			120	29.7	1.01	21.5
L6	6	45.6p	16.0+	0.81	1.20	1.2	1.49	1.7	0.53	0.50	0.51	320	90.7	1.14	58.0

The symbols at the head of the table denote the following hydraulic parameters:

* excluding LB minor channels; Q = water discharge; p denotes partial cross section; w = water surface width   R = hydraulic radius averaged for all channels in a section; d_{max} = maximum depth; d_{umax} = depth @ maximum velocity; u = average flow velocity; u_{max} = maximum flow velocity; Fr = $u/(g d)^{0.5}$ cross sectional averaged Froude number; Fr_{umax} = Froude number at vertical where average velocity is highest; Fr_{rmax} = cross sectional maximum Froude number ; Ω = γQS power per unit length $ω_w$ = γQS/w power per unit width or bed area; ω = Ω /γRv = uS unit power; τ = γRS average shear stress at section; S = average longitudinal bed slope (see Ironne and Duncan (1990) for comparison of S with the local   water surface, energy and bed slopes).

River bed changes
The upper reach

Fig. 3 incorporates a series of 7 maps of the upper reach. The discharge at the Old Weir is noted in each map in order to evaluate the extent of braiding without reference to base flow conditions. During event 1 only small changes occurred: the RB channel between U2 and U3 slightly eroded its right bank, erosion which persisted during event 2, continued in event 3 and was the causal mechanism for the partial avulsion: the formation of the channel in the far RB during event 4. This secondary anabrach deepened in event 5. The small channel realignment at U2-U3 during event 1 brought forth a slight left hand course of the RB main channel at U3-U4 during event 2, ultimately causing RB erosion at U4-U5 and the cutoff at U6 during event 3. The LB main anabranch was slightly plugged at U2 after event 2, and remained fossilised thereafter until it was almost completely plugged at U1-U2 during event 5 and infilled during event 6.

The previous sequence of events shows that small amounts of bed reworking and channel alignment may ultimately cause major changes in location of braids and extent of braiding. Comparing Fig. 3a,b,f and g (note that flow discharge was similar in Fig. 3b and 3f) demonstrates what appear to be two general changes of the braided bed. Some anabranches are abandoned, at times they are plugged at their upper or lower edge, but they are not infilled as additional, rather small events (1-5) recur. It follows that the bed becomes more braided (by adding new anabranches which cut across bars of various sizes and heights) as long as massive overbank flow does not take place. During smaller events channels are realigned and appear to retain the general low sinuousity of 1.1. Overbank flow during a major event and rapid avulsion of the main channel brought about activation of most of the bed, infilling of previous anabranches and the formation of new ones (Fig. 3g-f).

The succession of events leading to bar reworking is also depicted in repeat cross sectional views (Fig.4a-c) of U2, U4 and U6. At U2 the LB main anabranch remained essentially unchanged while the RB anabrach actively erodeed its banks during events 1-3, with total reworking of the river bed, including massive scour and fill during events 4-6, essentially event 6 (fig. 4a). A similar succession of events occurs at mid bar (U4, Fig. 4b). The confluence scour hole at the bar tail was still present after event 1 (Fig. 4c), but the cutoff had begun by eroding the slight protrusion to the right hand of the hole. Most of the hole has filled by the end of event 2 while a similar volume of reworked bank had been eroded in the new channel after event 3. Erosion of high, protruding bar tops and massive deposition was also prevalent at this site during event 6. In summary, it appears that reworking occurred during all the events but new bars were formed only during the large event, when flow occurred on bar tops rather than in anabranches with armoured beds and banks.

The preferred location of deposition sites is demonstrated by areas of fill in plan and sectional views, but also from tracer transport. Tracer location in the upper reach is depicted after each event in Fig. 5. Almost all tracers had not moved during events 1-5 on the higher bar platforms. Some particles were transported considerable distances in the LB channel during event 1 (Fig. 5a). They were apparently unstable originally because of protrusion and lack of interlocking (Laronne and Carson, 1976; Fenton and Abbott, 1977; Brayshaw et al., 1983). The RB anabranch could not be searched due to relatively deep flow and the short interval between events 1 and 2. The lines shown in Fig. 5b are not tracer trajectories but simply connect tracer location sites before and after event 2. In fact, the central bar was not overtopped at all during events 1-5. so that it is evident that tracer trajectories must be parallel to and within the anabranches. In the LB branch tracers moved 50-100 m. and most of them remained immobile during events 3-4 and moved an additional distance of 50-100 m during event 5. In the RB branch distances of transport were much larger and, in fact, no tracers were found between sections 1-4 because all had moved in the channel to the stoss, tail of the bar. They were preferentially deposited at the bar head located downstream from U6. During event 3 some of the tracers at the bar head moved slightly to form a bulge or plug at the upstream end of the LB channel. Insufficient evidence is available on transport during event 4 and 5, but it is apparent

that flow velocities were sufficient to remold the protected, LB side of the river bed which had been inactive hitherto. This conclusion is also borne out by the material derived from scour chains. During event 6 transport distances were very large and, in fact, most of the tracers were lost downstream or under a fill exceeding 1 m. It is apparent, though, that the previous longitudinal bar had been eroded, and bed material from the bar platform had been moved downstream more than twice the length of the large, longitudinal bars. The shorter distances of tracer transport near the outside banks demonstrate a parallel downriver movement.

The lower reach

Fig. 6 incorporates a series of 7 maps of the lower reach. Although event 1 had a low peak discharge, the hydrograph rise and fall were prolonged, allowing considerable change to take place in the lower reach (Fig. 6a and b). The main channel at L1-L2 upstream of the RB alternate bar became plugged at this location, a process which culminated in channel infilling during events 2 and 3 (Fig. 6c and d). The main body of the alternate bar was only sliughtly molded because flow depth was minimal on the bar platform. Nevertheless, the platform was mobilised during the succeeding, artificial event because peak flow was greater even though the time base for bedload transport was short (see Table 2). During event 3 both channels were filled at L1-L4 (Fig. 6d) and two new alternate bar, significantly longer than the previous ones, were formed during event 4. The entire bed was remolded creating brand new bars not only during the largest event but also during event 5 (Fig 6f-g).

The sectional views of the lower reach offer just as convincing evidence (Fig. 7 a-c). At L2, the alternate bar head, the channels filled as much as 1 m and some reworking also occurred on the bar tops (Fig. 7a). Event 2 was capable of remolding the entire bed, with massive erosion and deposition of 0.5 thick sediment wedges during event 3, and total channel change during the succeeding events. Minor changes took place during the first event at the centre of the alternate bar platform due to some bank protection and low peak discharge, more during the second when a central channel had incised into the bar, considerably in the outer bar areas during event 3 and totally during the next events (Fig. 7b). The bar tail (section L6, Fig. 7c) behaved in a manner similar to the bar head but bed activity was mirrored to the opposite bank.

Tracer location after events 2, 3, and 6 are depicted in Fig. 8a-c. Only few tracers were found after event 4, so these locations were omitted herein. The short interval of time between events 1 and 2 precluded locating tracers in this reach. Although the lines in Fig. 8a are not all transport trajectories, the shorter ones at L1 and downstrean of L6 are assumed to be trajectories on bar platforms. Particles originating at the RB of L2 and L3 appear to have moved along the curved channel and deposited at the top and centre of the next alternate bar. After event 3 many tracers were located. The lines depicted in Fig. 8b involve transport during events 1-3. It is evident that many particles were transported along lines approximately parallel with the general alignment of the river, crossing a part or all the upper alternate bar to be deposited anywhere along the route. The data from event 4-6 is very meagre, but shows that distances of bedload transport were considerable.

Summary

The evidence presented in this paper reinforces previous suggestions that small events rework bars but may also bring about large changes such as cutoffs, avulsion of the main anabranch and channel switching (Werritty and Ferguson, 1980; Ferguson and Werritty, 1983; Ashworth and Ferguson, 1986; Ashworth, 1987). Additionally, it is apparent that new bars may be formed and that the entire bed may be molded by events of small magnitude provided that bar platforms are sufficiently submerged to allow higher velocities to develop on the platforms. It does not, though, point towards decreased aggradation within such confined reaches (Carson and Griffiths, 1987; Davies and Lee, 1988; Laronne and Duncan, 1990). Flow direction is not listed in the table, but critical observation of flow direction at the water surface and 0.6 depth by the gauging parties shows that in all instances except event 1, flow direction in the lower reach was essentialy downchannel, parallel to the general alignment of the reach, and it was parallel to anabranch alignments in the upper reach. The upper reach contained many areas that were neither

boatable (too shallow) nor wadeable (too fast) during event 5, so that additional attempts to gauge it were relinquished. We attempted to gauge the lower reach also during the larger event 6, but standing waves 0.9 m high and shallow depth on bars allowed only partial gauging., 198?; Laronne and Duncan, 1990). It is also apparent that the lack of a very coarse armour in the bed and banks of the inner channels enables large channel changes to occur during small events.

Bedload moves totally within anabranches and may diverge on bar heads during smaller events. It appears that most bed material is transported on top of bar platforms parallel to the river banks during formation of new bars. This favours a mechanism of bar (both alternate, point-type and medial bar) formation during high stages of flow due to advance of sheets or lobes. The occurrence of some gravel sheets a-la-Hein (1974) in the upper reach after event 6 and the particular character of massive deposition of wide, tabular matrix and somewhat clast supported sheets in the lower reach (Laronne and Duncan, 1989b) also lend support to the notion of sheet or lobe advance.

Self evidently, the formation of bars is directly interrelated with hydraulics and sediment transport. The actual changes in bed elevation derived from scour chains, the sedimentology of the deposited bed material as well as bedload sampling undertaken in this project will be presented elsewhere.

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Figure Captions

Fig. 1 The North Branch Ashburton River Basin, showing the location of the study reaches and gauging locations.

Fig. 2 Near vertical aerial photographs of the upper study reach (a) and the lower study reach (b).

Fig. 3 Maps of the upper, wide study reach pre events 1-7 (a-g). The symbols denote identical features in all maps. Observe the location of cross section U1-U6. Map 3a corresponds to Fig. 2a.

Fig. 4 Resurveys of cross sections at the head of a longitudinal bar at U2 (a), at its centre at U4 (b) and at the downstream end at U6 (c), upper study reach. Vertical exaggeration is 55. Black areas represent scour; dotted ones represent net fill.

Fig. 5 Tracer locations in the upper reach after events 1-6 (a-f).

Fig. 6 Maps of the lower, narrow study reach pre events 1-7 (a-g). The symbols denote identical features in all maps. Observe the location of cross sections L1-L2. Map 5a corresponds to Fig. 2b.

Fig. 7 Resurveys of cross sections at the head of an alternate bar at L2 (a), at its centre at L3 (b) and at the downstream end at L6 (c), lower study reach. Vertical exaggeration is 26.5. Black areas represent scour; dotted ones represent net fill.

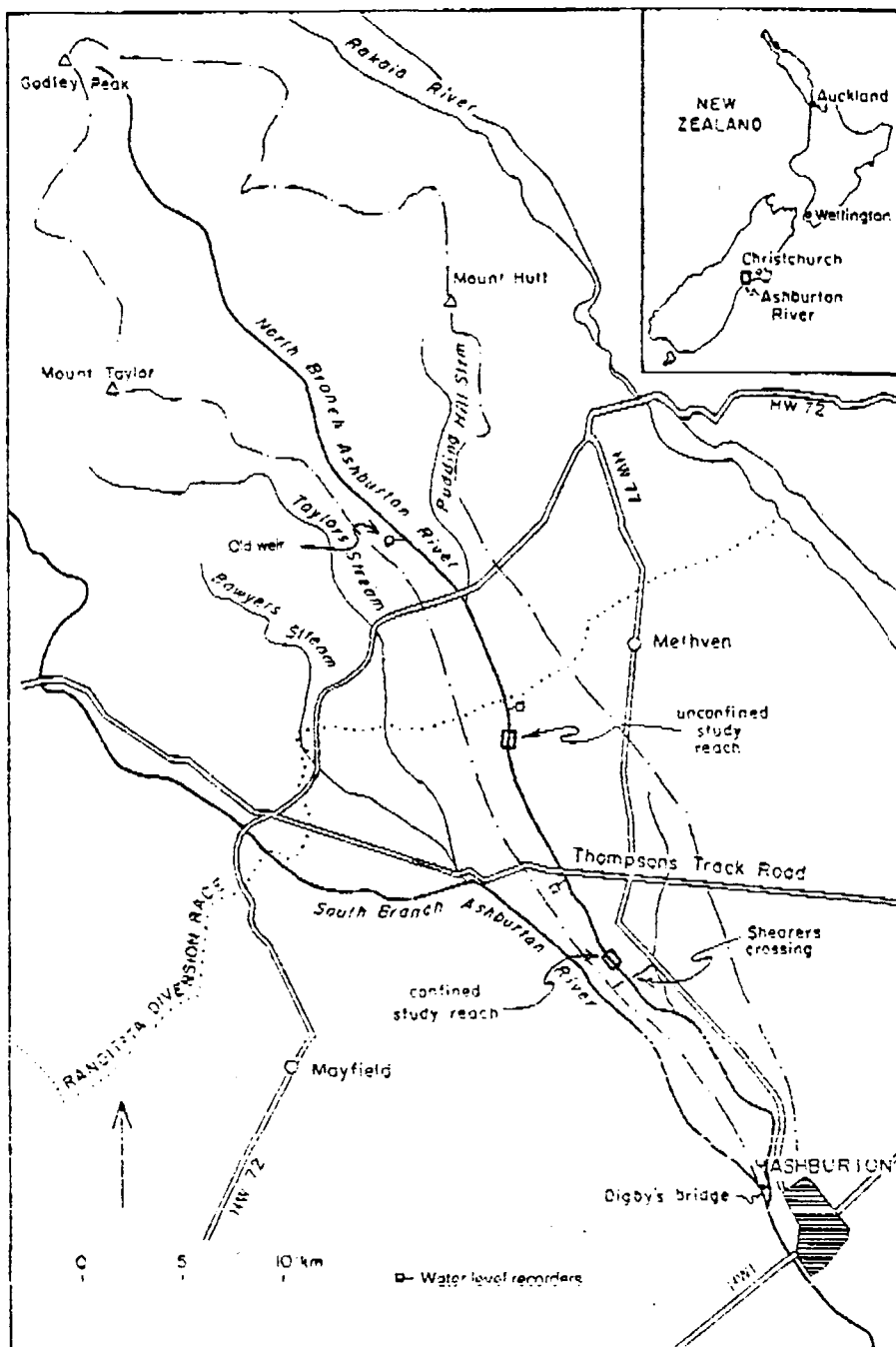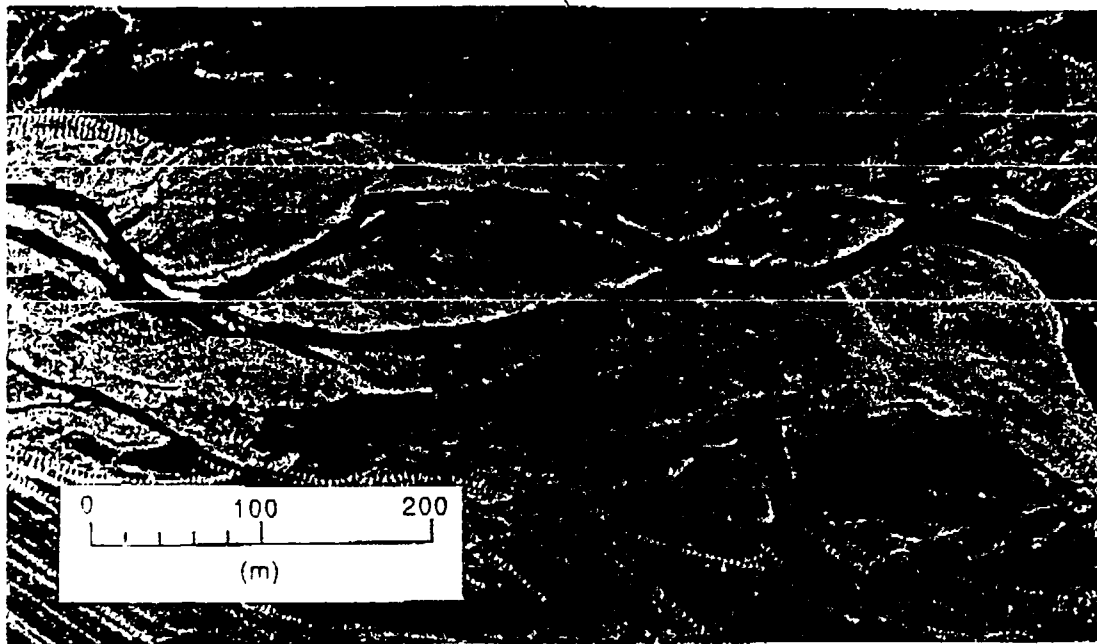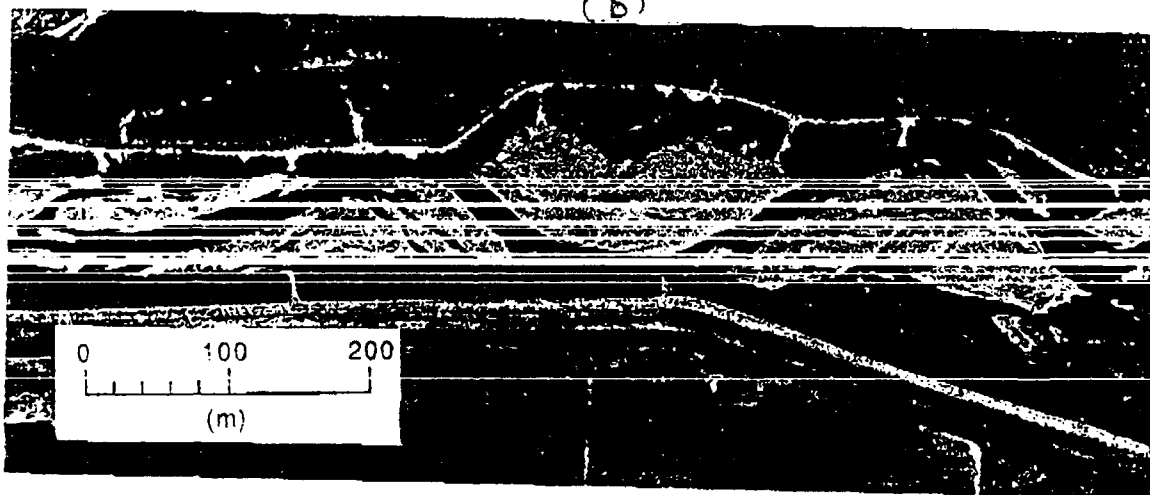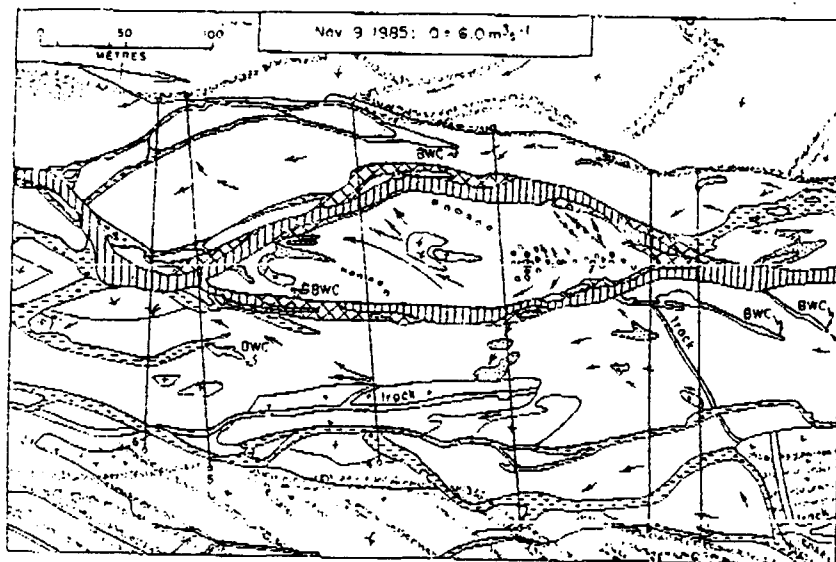Fig. 8 Tracer locations in the lower reach after events 2,3 and 6 (a-c).

Fig. 1

Fig 2

(a.)

← FLOW

(b)

← FLOW

Fig 3

(a) Nov 9 1985; Q = 6.0 m³s⁻¹

(b) Dec. 9 1985; Q = 9.2 m³s⁻¹

(c) Dec. 19 1985; Q = 6.6 m³s⁻¹

KEY
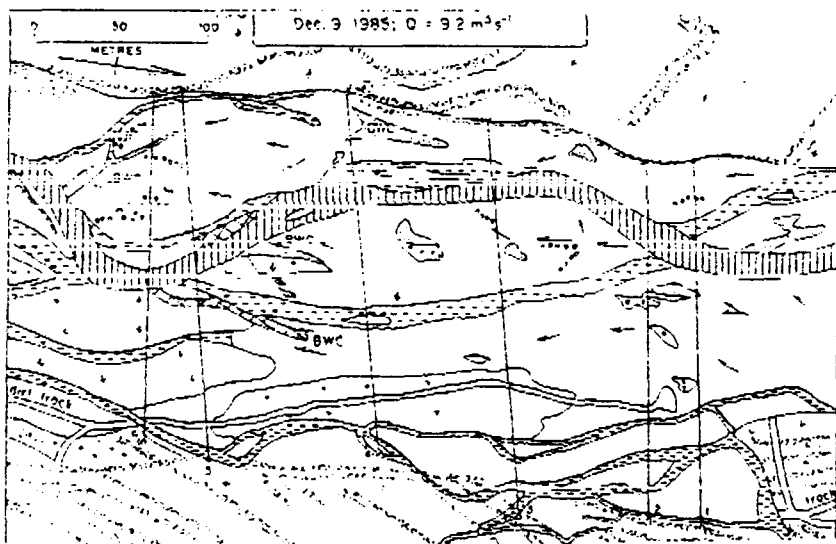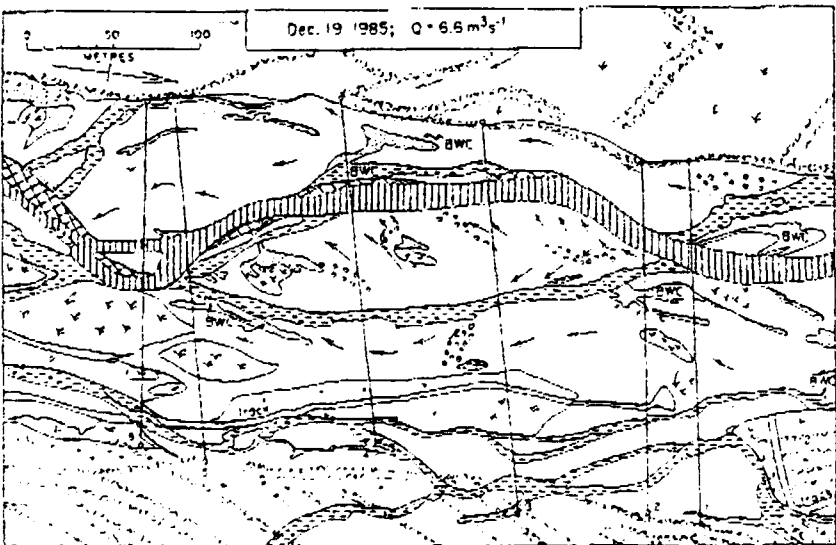
	main channel
	minor channel
	shallow water
BWC	back water channel
	sand
	coarser gravel
	flow direction
	steep bank
	granule-pebble delta
	scour hole
	scour trench
	ground truth marks
	scrub

(c)   Mar 5 1986, Q = 8.6 m³ s⁻¹

(b)   Apr 1 1986, Q = 11.5 m³ s⁻¹

(d)   Jun 3, 1986, Q = 18 m³ s⁻¹

(f)   Feb 2, 1986, Q = 9.3 m³ s⁻¹

Fig 3

Fig 4

Distance (m)

Scour
Fill
Post event 3
Post event 6

(a) UNCONFINED REACH POST FLOW 1 EVENT — Particle movement downstream

(b) UNCONFINED REACH POST FLOW 2 EVENT — Particle movement downstream

(c) UNCONFINED REACH POST FLOW 3 EVENT — Particle movement downstream

Fig 5

UNCONFINED REACH
POST FLOW 4

Particle movement downstream

(d)

UNCONFINED REACH
POST FLOW 5

Particle movement downstream

(e)

Fig 5

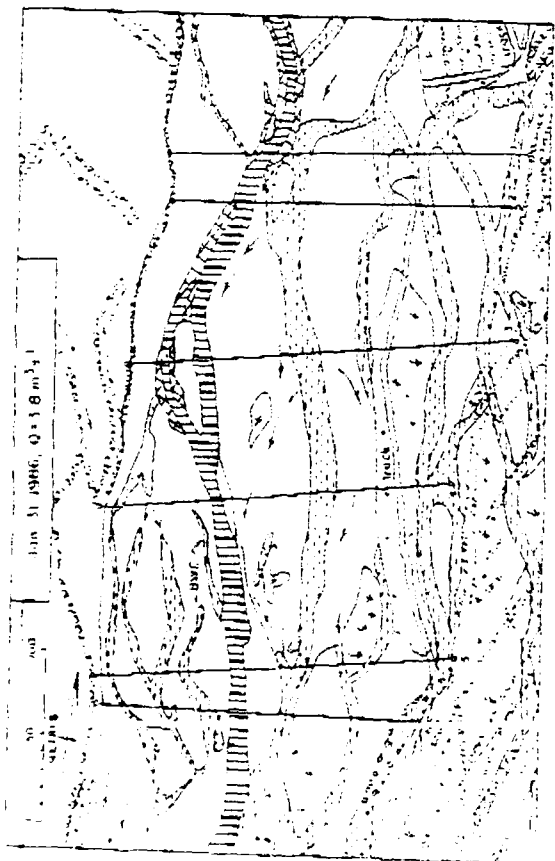UNCONFINED REACH
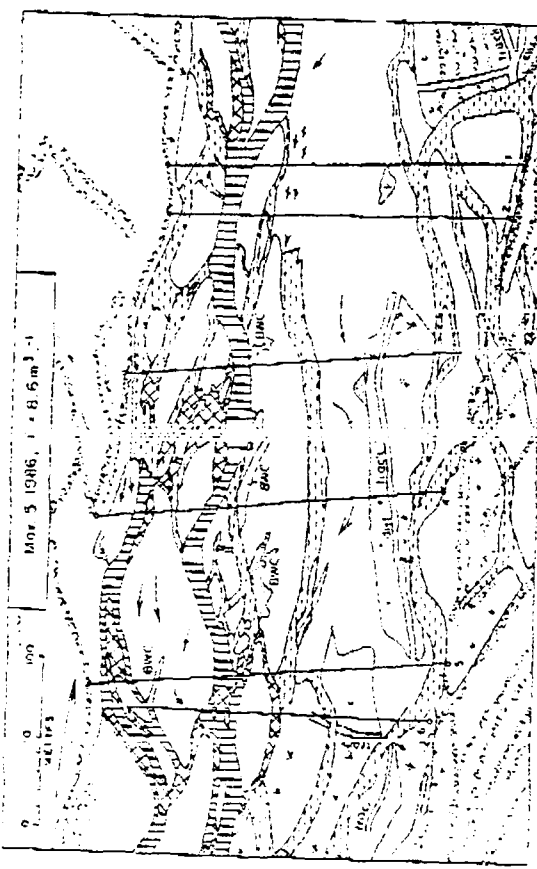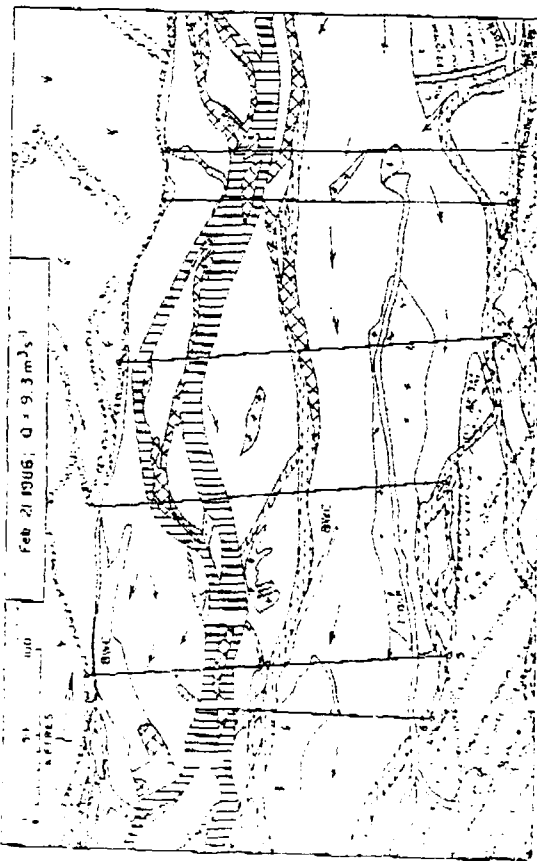POST FLOW 6

Particle movement downstream

(f)

Fig 6

Key
main channel
minor channel
shallow water
back-water channel
sand
coarse gravel
flow direction
steep bank
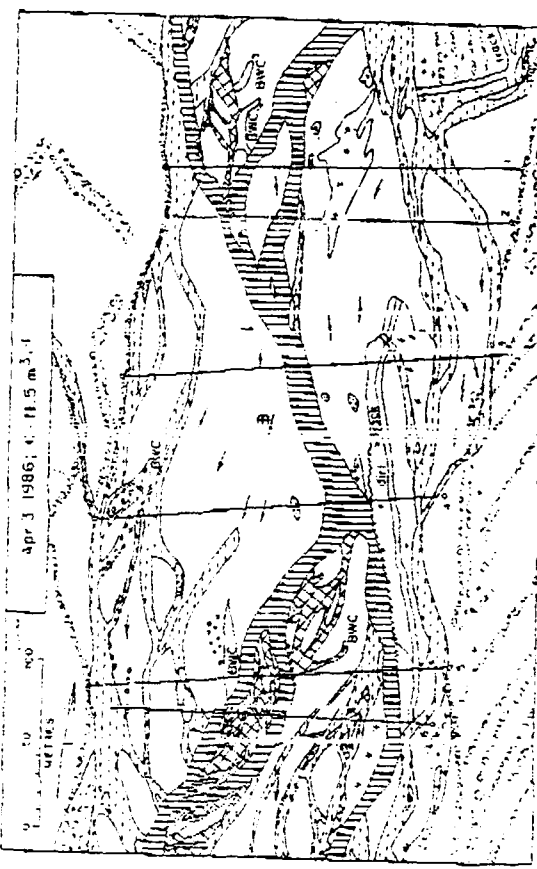granule-pebble delta
scour hole
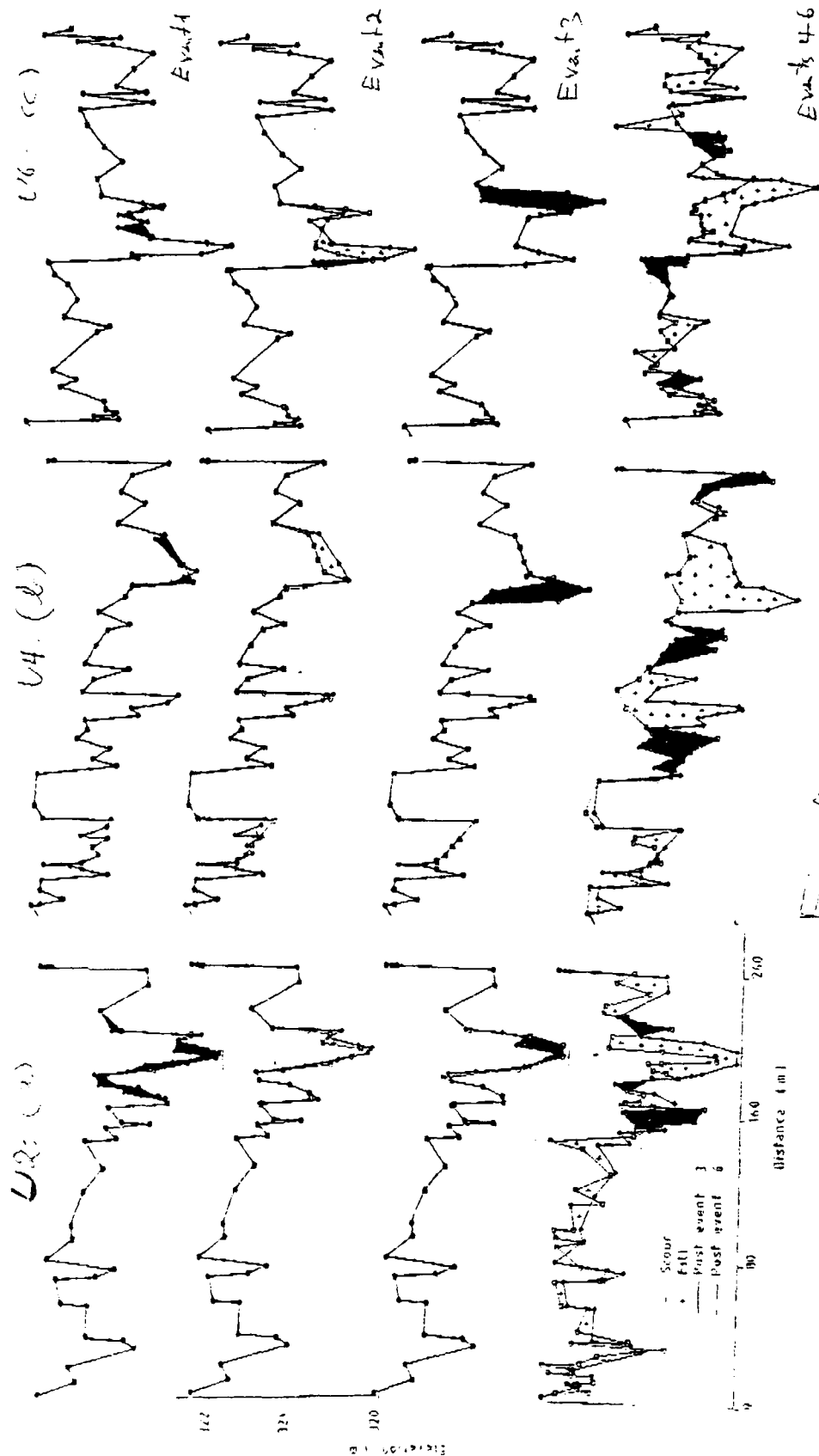scour bench
ground truth marks
scrub

(a)

(b)

(c)

FIG 6

(c) Dos top: 97

L3:(b) low value

L2: (a) buried

Eval 1

Eval 2

Eval 3

Eval 4 & 5

Distance (m)

Fig. 7

Particle movement downstream

WILLOWS

WILLOWS

0   50
METRES

CONFINED REACH
POST FLOW 2

(b)

Particle movement downstream

WILLOWS

0   50
METRES

CONFINED REACH
POST FLOW 3

(c)

Particle movement downstream

WILLOWS

WILLOWS

Distance downstream from origin

0   50
METRES

CONFINED REACH
POST FLOW 6

Fig 3

DOWNSTREAM FINING IN A GRAVEL-BED RIVER IN S. POLAND: LITHOLOGIC CONTROLS AND THE ROLE OF ABRASION

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INTRODUCTION

It has been known for over a century that the bed material of rivers generally becomes finer in a downstream direction. Explanations for this phenomenon have focussed on abrasion and selective transport as the processes causing this pattern, but their relative importance in different field situations has been the subject of considerable debate amongst geomorphologists.

One of the first analyses of downstream fining was that on the Rhine by Sternberg (1875) who observed that grain size changes in bed material followed an exponential pattern of the general form:

$$W / W_0 = e^{-as}$$

where W_0 is the initial weight of a characteristic particle at distance zero, and W is the characteristic weight at a distance s downstream; 'a' being a coefficient of weight reduction. This result was attributed to the process of abrasion as the bed material is transported downstream, a conclusion further confirmed experimentally by tumbling mill experiments by Daubree (1879). Schoklitch (1933) and Krumbein (1941). Kuenen (1956), however, observing that tumbling mills did not adequately reproduce the abrasion processes operating in rivers, developed a circular flume in which the flow was driven by paddles. He identified a number of discrete processes involved in size reduction (splitting, crushing, chipping, cracking and grinding) and generally confirmed an exponential decrease in size. The impact of experimental conditions was also clearly demonstrated by Kuenen's results in which much lower abrasion coefficients were obtained on a concrete bed than on a pebble bed. But the overall pat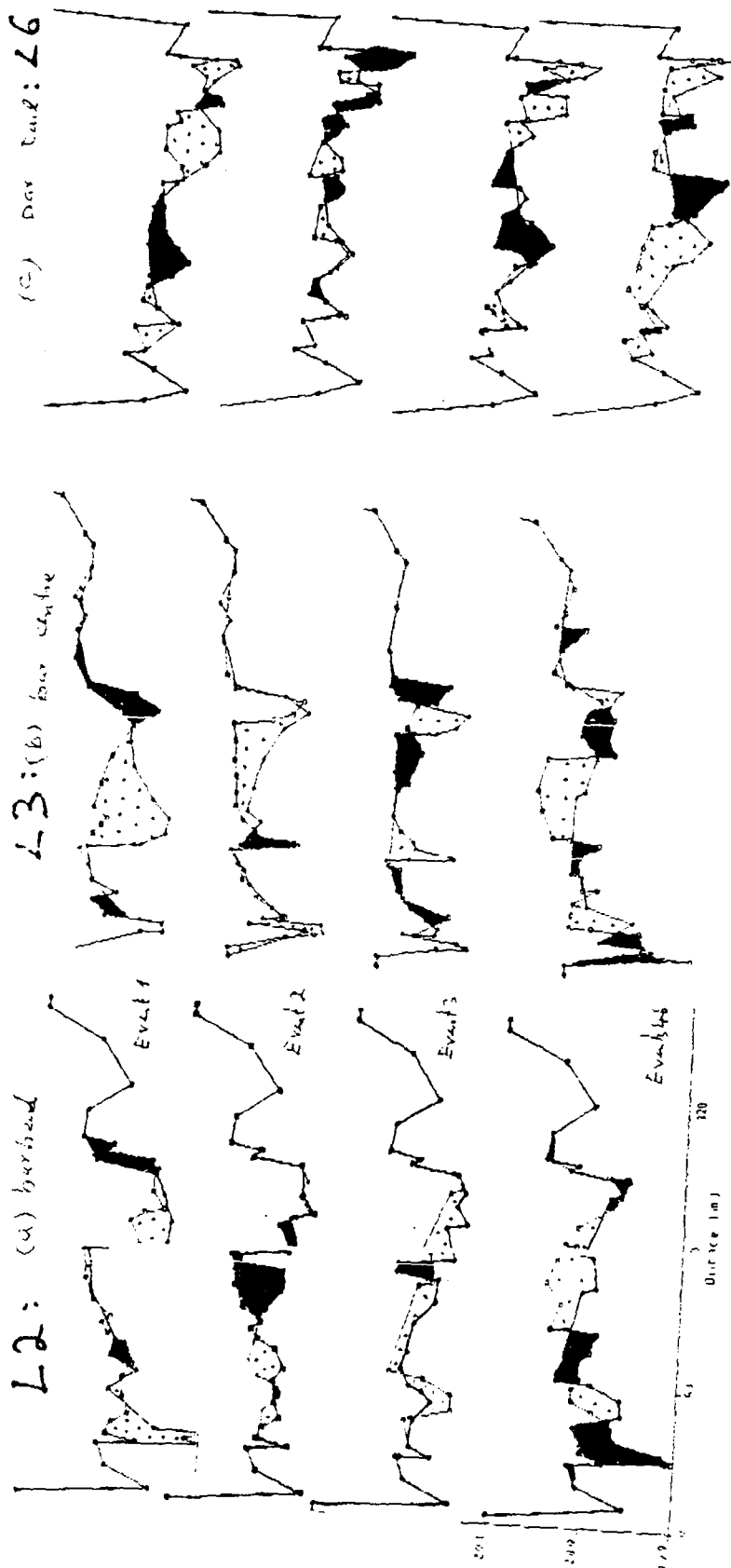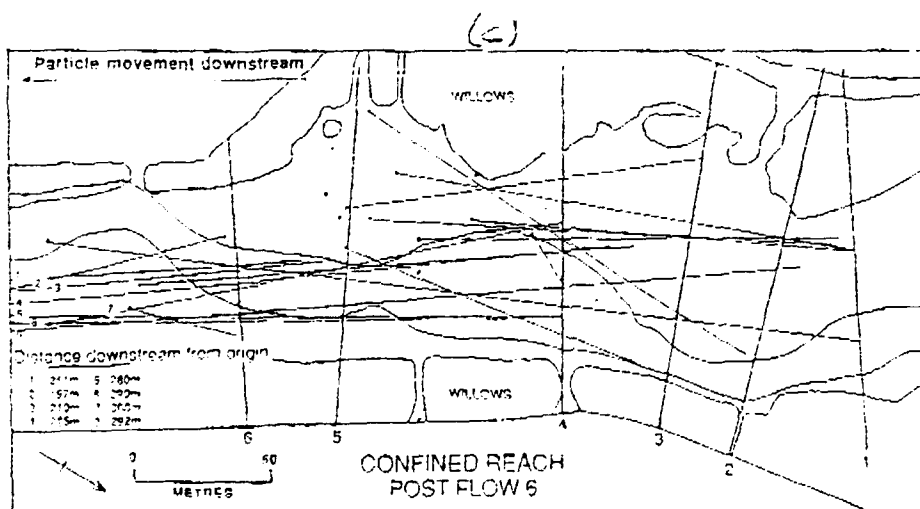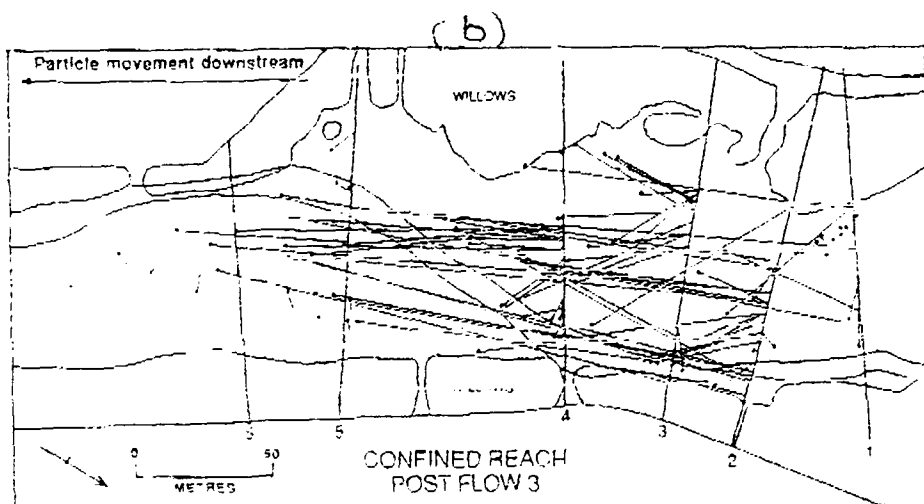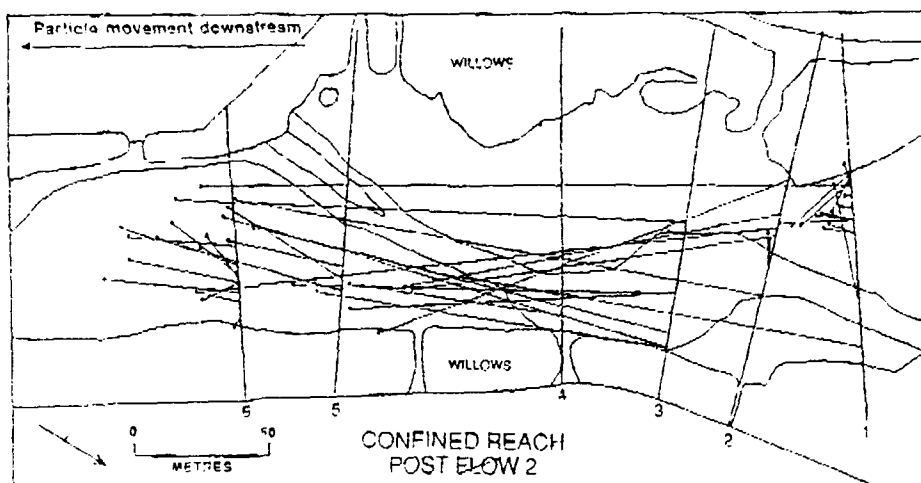tern of results from a variety of experiments with the circular flume demonstrated that the rates of size reduction observed in natural rivers were much greater than could be attributed solely to abrasion effects.

As a result of these findings, selective transport, or hydraulic sorting, began to be accorded the dominant role in determining the rate at which downstream fining of bed material occurs. Thus Bradley *et al* (1972) claimed that approximately 90% of the size reduction in Knik River gravels in Alaska could be attributed to selective transport and only 10% directly to abrasion. Other field-based studies by Plumley (1948) on terrace gravels in the Black Hills and Unrug (1957) working on the bed material of the River Dunajec in southern Poland have also identified rates of downstream fining which are not in accord with Sternberg's abrasion-based exponential "law". However, further complexities in the processes which determine downstream fining have also been recorded and these have accorded a greater significance to abrasion processes. For example, Bradley (1970) noted an order of magnitude difference in the attrition coefficients for fresh and

weathered granite when subjected to laboratory abrasion tests. Schumm and Stevens (1973) have also argued that even Kuenen's work may underestimate the role of abrasion because of attrition that occurs as stones vibrate at flow velocities below the entrainment threshold. Shaw and Keliarhals are prepared to go even further and tentatively claim for rivers in North Saskatchewan that "no contribution to downstream change in grain size is made by differential transport" (1982, p. 39).

It is clear from the above that the relative roles of abrasion and selective transport in determining the downstream fining of bed material in rivers are still imperfectly understood. One of the major problems has been the failure to identify of a field situations where the two processes can unambiguously be separated from each other. This study, based on a river in Southern Poland, attempts seeks to achieve this separation and thereby assess the role of abrasion on contrasting lithologies quite independently of the effects of selective transport.

STUDY AREA

The study area comprises a reach of c.20km along the Czarny Dunajec and its tributaries in S. Poland. The River Dunajec rises in a series of headwaters which drain north from the Tatras Mountains (fig. 1). Each of these tributaries successively drains the crystalline basement (mainly granites, schists and quartzites) and then successively Mesozoic limestones and dolomites and Tertiary flysch (shales locally interbedded with massive sandstones). In the case of the Czarny Dunajec the two main tributaries (Chocholowska Dolina and Koscieliska Dolina) rise in the crystalline basement and flow almost due north successively incised into the Mesozoic limestones and dolomites before they join to form the Czarny Dunajec at Roztoki by which time they have emerged from the Tatra massif into the basin of the Podhale flysch.

There are two striking consequences of this sequence of lithologic units which make it ideally suited for this investigation. Firstly, with the minor exception of the sandstone members of the flysch series, no lithologic unit is repeated within the sequence and thus an unambiguous final downstream input of a given lithology can be identified from the geological map. Secondly, the actual sequence of lithologic units is ideal for this study in that the initial inputs of bed material into the headwaters are the most resistant rocks types (quartzites and granites), followed by limestone and dolomite (less resistant) and the sandstones and shales (least resistant). Since the most resistant rock type (quartzite) is one of the first to cease having a direct input into the system, it can be used to provide a 'reference curve' for lithologies which enter later in the sequence.

A second crucial control favouring the selection of this particular reach of the Czarny Dunajec is the recent Quaternary history of the area. During the last glaciation (of Vistulian age), the Tatra massif underwent valley glaciation, but none of the individual icestreams extended beyond the central part of the Tatras. Thus in Chocholowska Dolina and Koscieliska Dolina the maximum extent of the Vistulian ice is recorded by terminal moraines which are located within 1km of the boundary between the crystallines and the Mesozoic sedimentaries (fig. 1). This limit provides a common datum from which the downstream fining of the granites and quartzites can be determined since any crystalline bed material found north of this limit can only have been transported to its current position by fluvial processes. A similar statement can be made for the limestone bed material found north of Kiry on Koscieliska Dolina and north of Siwa Polana on Chocholowska Dolina (fig. 1). Thus for all three lithologies (granite, quartzite and limestone) the last possible input or point source into the Czarny Dunajec system can be determined quite unambiguously. In the case of the massive sandstone units within the flysch sediments the situation is more complex since there are numerous sandstone members within the Podhale flysch series. Nevertheless, it is possible to identify two particularly massive sandstone members: one 500m south of Kojsowka and the other directly west of the village of Chocholowska (fig. 1). The sandstones from both these sources breakdown very rapidly and thus there is no danger of large material derived from

these two sources being confused with each other or with other sandstone members.

Another possible complication which could in part vitiate the underlying rationale behind this project would be the large-scale reworking of coarse older Pleistocene covers. Field checking along the whole 20km reach demonstrated that this problem does not occur. Until the 19th century the Czarny Dunajec was an active, braided gravel-bed river (Krzemien 1981). Since then it has undergone localised river training such that the modern channel is generally undivided. However, where it is not bedrock controlled it is reworking floodplain deposits of very recent origin and not the older Pleistocene covers which are now at some distance from the active channel. When the river ultimately emerges from the Podhale flysch into the Novy Targ basin it becomes less confined and its tendency to braid substantially increases. This has necessitated extensive river engineering in the form of large concrete drop structures at a spacing of every few hundred metres northwards from the settlement of Koniowka (fig. 1). These act as large scale sediment traps for the coarsest bedload fractions and for this reason the downstream termination of this investigation occurs just upstream of where these drop structures commence.

The long profile of the Czarny Dunajec (fig. 2) reveals an overall gradient of 0.0175 which increases to 0.0208 for the first 5km. Throughout the whole of this long profile the river bed is intermittently in contact with the bedrock and, particularly in the flysch zone, bed material is directly incorporated into the bedload by hydraulic action during floods. Discharge data are available for two sites within the study reach: Koniowka (drainage area 120km^2)at the end of the reach has recorded a maximum daily flow of 66.4m^{3}s^{-1} over the period 1976-86, whilst Kojsowka (drainage area 94km^2) has recorded a maximum of 50.2m^{3}s^{-1} during the same period.

DATA COLLECTION

In any study concerned with the size of bed material the sampling strategy and measurement technique require careful selection and justification (Church et al. 1987) In this instance, given the number of sites to be investigated plus the particular problem being addressed, it was decided to concentrate on the largest fraction of each lithology present at individual sites. This decision can be justified on the grounds that not only is a consistent measurement required for each site, but also a method needs to be selected which will permit direct comparison between the different lithologies present. As far as possible lateral or mid channel bars were selected, typically up to 80m in length. At each of these 33 sites a systematic search was undertaken to locate the largest sizes present for each of the available lithologies, generally these were to be found either at the head of the bar or close to the active channel. For the most upstream sites only quartzite and granite were available, but as the survey progressed downstream limestone and eventually sandstone became incorporated into the search procedure. For each individual lithology a minimum of the 10 largest clasts was identified, the size of each clast being determined by measuring the B-axis. In the majority of searches substantially more than the 10 largest clasts were measured (in order to be quite certain that the sample did indeed include all the largest clasts present). Later these samples were 'censored' by ranking the data and only including the 10 largest values in the final analysis.

In terms of distribution of sites both headwater tributaries (Chocholowsa Dolina and Koscieliska Dolina) were sampled separately and, given the rapid decrease in grain-size, initially the sites were very closely spaced (within a few hundred metres). Thereafter, downstream of the confluence of these tributaries at Roztoki, the sites became more widely spaced. This was partly a reflection of the absence of suitable bars to sample and partly because of the local removal of coarse bed material from the river bed for local construction purposes. Fortuitously, this only affected the granite (the most highly prized lithology for building decoration) and was restricted to be a specific size range (300mm to 440mm). Only in one instance has this resulted in the granite sample for a particular site being discarded.

3

LITHOLOGIC CONTROLS ON DOWNSTREAM FINING

The results are presented in a series of diagrams demonstrating different patterns of downstream fining in each of the headwater tributaries and then in the main channel of the Czarny Dunajec. Taking all the data together for Koscieliska Dolina and its downstream continuation as the Czarny Dunajec, a rather confusing pattern emerges although, as expected, in general the granites and the quartzites both fine much more slowly than the limestones and the sandstones (fig. 3). In addition to the pattern of downstream fining from the last possible input of a given lithology the magnitude of coarse inputs for limestone (A) and sandstone (B, C, and D) are also shown.

Each lithology will now be analysed individually to determine the optimum empirical fit in terms of either exponential or log-linear regression equations. Taking the quartzite curves for Chocholowska Dolina first, a rapid decrease in maximum clasts size from approximately 950mm just beyond the terminal moraine down to 450mm within 5km is recorded in fig. 4. A similar pattern emerges for Koscieliska Dolina although the rate of fining is less with the value after 5km of 593mm. For both tributaries the goodness of fit as measured by the coefficient of determination is high with that for Koscieliska being somewhat better than that for Chocholowska. Thereafter, working downstream along Koscieliska Dolina and into the main stem of Czarny Dunajec, further diminution in size is very small, such that even after 20km quartzite boulders >550mm can still be found in within the river bed. The log-linear model best describes the pattern of downstream fining with higher reduction coefficients (-0.162 and -0.116) for the first 8km of Chocholowska and Koscieliska respectively. These coefficients are both higher than the -0.0875 for the 20km reach of Koscieliska Dolina and the Czarny Dunajec. In no instance did the exponential regression model provide a better fit than log-linear model. For quartzite bed material the very rapid initial decline in maximum clast size favours the log-linear rather than the exponential model (Table 1)

The pattern for fining of the largest granite boulders is broadly similar to that for the quartzites (fig. 5). Again a very rapid reduction from 950mm to 467mm within the first 5km is recorded for Chocholowska Dolina, although the comparable value for Koscieliska Dolina after 5km is higher (581mm). Again log-linear functions best describe the pattern of downstream fining for Koscieliska Dolina with a reduction coefficients of -0.165 being recorded. However, for Chocholowska Dolina the exponential model provides a better fit (R squared of 0.908) and so the reduction coefficient is not directly comparable. Extending the results for Koscieliska Dolina downstream to include the Czarny Dunajec, the size of the largest granite bed material continues to decline only slowly towards 434mm at a distance of 20km from its origin at the terminal moraine. The resulting pattern conforms extremely well to a log-linear curve (coefficient of determination of 0.922 which is strikingly better than the alternative exponential model (Table 1)). The only major anomaly over this 20km reach is at a distance of 7.1km where a value of 625mm is recorded rather than the 526mm predicted by the regression equation. As for the quartzites the log-linear model generally provides a better fit to the granite data than the exponential model.

The pattern of downstream fining for the limestone and dolomite bed material is quite different from that of the crystallines. As can be seen from fig. 6, although the initial values are larger than those for the granites and quartzites (since they include direct input into the river by rockfall) they rapidly decline from values >1000mm to <400mm within 5km. Thereafter, at 8km downstream from the terminal moraines (and approximately at the transition from limestone/dolomite to flysch) the largest limestone bed material is 180mm (Koscieliska) and 260mm (Chocholowska). This pattern of downstream fining continues such that 20km from the origin the largest limestone clasts are only 105mm in size. Given this pattern of results the reduction coefficients of -0.326 to -1.1068 record much more rapid fining than was found for the granites and quartzites. With the exception of Koscieliska Dolina, the log-linear model is generally favoured over the exponential model (Table 1).

4

The flysch sandstones behave quite differently to the crystallines and the limestones, and provide two major inputs of coarse sandstone blocks within the Czarny Dunajec study reach. The first (just upstream of Kojsowka, site A in fig. 7) comprises a 5m high vertical cliff of massive, well-jointed sandstones which are currently being undercut by the river. Within 300m the largest blocks have been halved in size from 1213mm (at A) to 645mm (at B). Within a further 2.2km they are further reduced to 300mm. An even higher rate of fining occurs at a second site, just west of the village of Chocholowska (fig. 7), where inputs from a bedrock source within the present channel and derived from the local Chocholowska sandstone is reduced from 713mm (D) to 413mm (E) within 400m. It is clear that the rate of reduction for the sandstones is far greater than that for any of the other lithologies. Taking the two very rapid fining cycles starting with coarse inputs at A and D the estimated reduction coefficients are -5.7 and -17.5 respectively. As anticipated, these are an order of magnitude greater than those reported for the limestones and over two orders of magnitude greater than comparable values for the crystallines. As was apparent from field evidence large sandstone blocks are broken up within a few hundred metres of their source.

A final, but rather ambivalent, piece of field evidence supporting the claim that abrasion is a substantial part of the overall pattern of downstream fining is shown by the steady increase in Krumbein sphericity recorded for the quartzites within Koscieliska Dolina as the bed material is transported downstream (fig. 8). The increase of sphericity from 0.66 to 0.73 over 8km is strongly suggestive of abrasion taking place. However, similar investigations of quartzite sphericity downstream in Chocholowska Dolina and the main stem of the Czarny Dunajec fail to reveal a similar pattern. There is thus no strong evidence over the whole study reach linking abrasion with an increase in quartzite sphericity.

On returning to Britain samples of the quartzite, limestone and sandstone were submitted to Lewin and Brewer for abrasion tests in their laboratory at University College Aberystwyth. Full discussion of the findings will be published elsewhere, but it was noteworthy that cubes of quartzite underwent a weight loss over 1-2% during simulated transport of 20km, the limestone recorded a loss of 12% by weight, whereas the sandstone samples suffered losses of 25% by weight and became very well rounded. Only the sandstone appeared to undergo reduction which might conform to an exponential model. The limestone sample underwent major weight loss by breakage. These preliminary experimental findings are completely in accord with the overall pattern of field-based results itemised above.

DISCUSSION AND CONCLUSIONS

Returning to the initial question, it is clear that the lithologic controls on the downstream fining of the coarsest fractions found with the upper reaches of the Czarny Dunajec drainage basin are very powerful. For quartzites, granites and limestones extremely well-defined fining cycles have been identified downstream of the last possible input ot a given lithology. Comparison of the relative merits of exponential and log-linear models to characterise these fining cycles generally favours the latter over the former - in 7 out 9 analyses the log-linear model provides a better fit. This largely arises because of poor fits for the exponential models in the uppermost reaches, especially when combined with a rapid reduction to quasi asymptotic values by 8km.. Such failures to conform to Sternberg's "law" has already been noted in a number of other studies which have investigated channel reaches very close to sediment sources. Thus both Bradley (1970) and Adams (1979) report very rapid declines in particle size at sites immediately downstream of a major sediment source. Similar observations have been made in proglacial and braided outwash environments (Boothroyd and Nummedal 1978, and Brierley and Hickin 1985). The implication of these findings is either that the Sternberg reduction coefficient is not consistent (a view favoured by Shulits (1941) or hydraulic sorting plays a major role (see Brierley and Hickin 1985). However, our concern in this paper is rather different since we have an opportunity to measure the

role of abrasion independently of selective transport.

Inspection of fig. 4 demonstrates that quartzite is the most resistant lithology contributed within this drainage basin. Once the largest quartzite clasts have reached c. 600mm they undergo very little further reduction. Although further abrasion does not occur, they still undergo transport throughout the study reach, and since there are no other sources of quartzitic boulders, one is forced to conclude that occasionally the Czarny Dunajec is competent to transport bed material of this size. Thus the curve in fig. 4 for quartzites in Koscieliska Dolina and Czarny Dunajec can be regarded as a 'reference curve' against which other lithologies can be compared. This curve defines the competence of the river throughout its 20km reach, the fact that the granites fail to retain a largest fraction comparable to that of the quartzites (especially at distances > 8km) arises not because of selective transport but rather because granites of this size are not available beyond 8km. The reason why they are not available is that they have undergone more abrasion. Typically the granites have undergone between 1.13 and 2.13 times more abrasion than the quartzites despite the fact that they share a common source in the two terminal moraines (Table 2). Similarly, the limestone reduction coefficients can be directly compared with those for the quartzites and their much greater susceptibility to abrasion (up to 9.54 times can be measured, Table 2). The sandstone data are difficult to interpret in a quantitative manner since the rate of breakdown is so rapid. A reasonable estimate of their behaviour is to characterise their size reduction as occurring at a rate at least 100 times more rapidly than that for the quartzites.

At this point a cautionary note needs to be added in terms of terminology, since the term 'abrasion' implies the style of wear of sedimentary particles recorded in the tumbling mills and recirculating flumes of Krumbein (1941) and Keunen (1956). It is now realised that size reduction independent of selective transport involves a number of processes including abrasion in place" (see Schumm and Stevens' 1973 and referred to as "vibration abrasion" by Shaw and Kellerhals 1982). There is also the role of physical and chemical weathering on clasts exposed on bars surfaces for considerable periods in between competent floods, and for limestone the role of solution when the clasts are buried beneath the local groundwater table. All these processes collectively contribute to size reduction which is independent of hydraulic sorting.

This study has served to demonstrate that for the largest fraction transported by a gravel-bed river, the role of abrasion and other related physical and chemical processes which can reduce the size of large clasts *in situ* is highly significant. In this particular study it has been possible to express the resistance of granites, limestones and sandstones in a quantitative manner relative to that of quartzite. This has a number of implications that warrant further investigation, not least being that of the relative resistance of the modal parts of grain size curves rather than examining the behaviour of the coarsest fraction present. The sampling problems inherent in such an investigation would, however, be very severe. More promising in the short term would be similar analyses involving controlled comparisons for other lithologies than those reported in this study. Since the processes described in this paper determine the shape of the upper tails in grain-size curves for fluvially-derived sediments, further investigations to validate or falsify the above conclusions are clearly needed.

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Table 1   **Regression equations for downstream fining patterns**

	Reduction coefficient	Coefficient of determination (log-linear)	Coefficient of determination (exponential)
(a) Quartzite:			
Chocholowska	-0.162	0.656	0.240
Koscielika	-0.1159	0.785	0.470
Czarny Dunajec	-0.0875	0.661	0.356
(b) Granite:			
Chocholowska	-0.183	0.732	0.908 *
Koscielika	-0.1652	0.881	0.669
Czarny Dunajec	-0.1864	0.922	0.751
(c) Limestone:			
Chocholowska	-0.326	0.865	0.684
Koscielika	-1.1068	0.980	0.992 *
Czarny Dunajec	-0.7894	0.888	0.715

* exponential model better fit than log-linear model

Table 2   **Comparative reduction coefficients from log-linear models**

	Chocholowska	Koscieliska	Czarny Dunajec
Qu. zite	-0.162	-0.1159	-0.0875
Granite	-0.183	-0.1652	-0.1864
Limestone	-0.326	-1.1068	-0.7894
Granite/quartzite	1.13	1.43	2.13
Limestone/quartzite	2.01	9.54	9.02
Limestone/granite	1.78	6.70	4.23

10
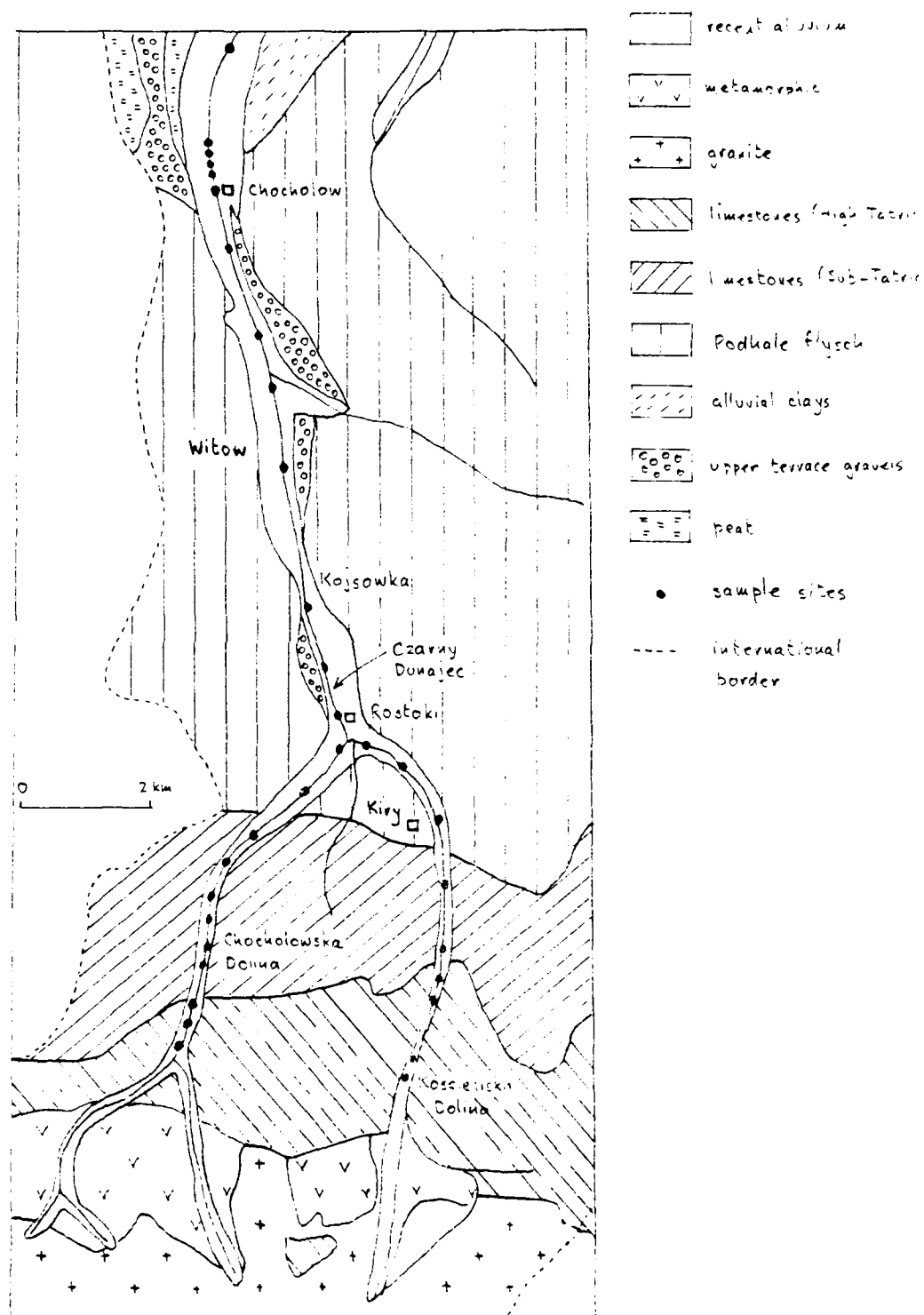
Figure 1   Sketch map of Czarny Dunajec drainage basin showing bedrock geology, stream
network and sampling locations.

Long profile of Koscieliska Dolina - Czarny Dunajec
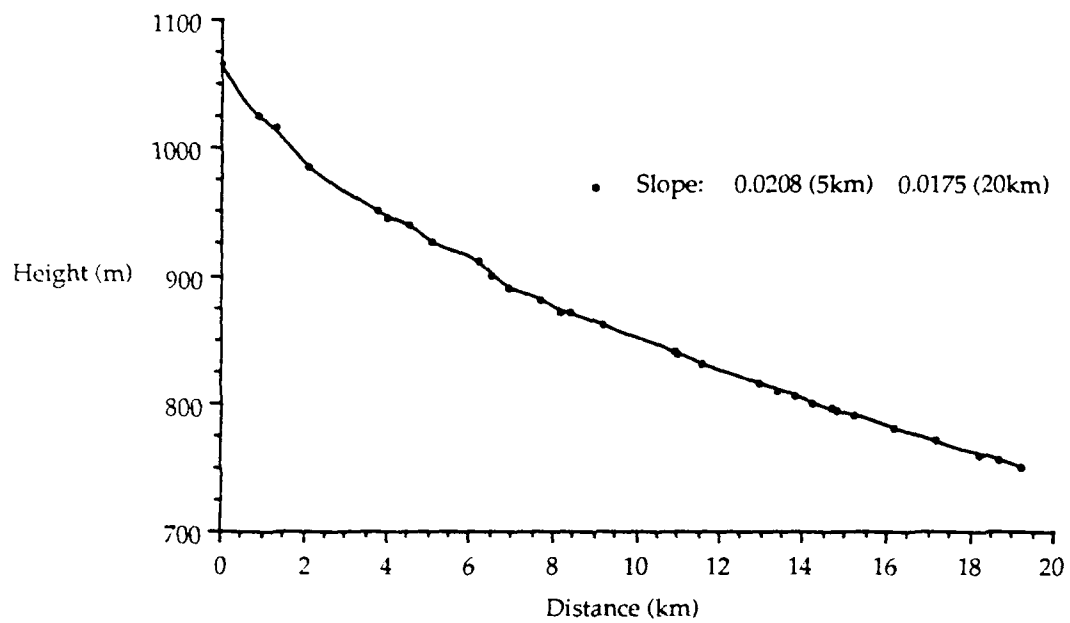
Figure 2   Long profile of Koscieliska Dolina - Czarny Dunajec

Downstream fining of quartzite, granite, limestone and sandstone:
Koscieliska Dolina and Czarny Dunajec
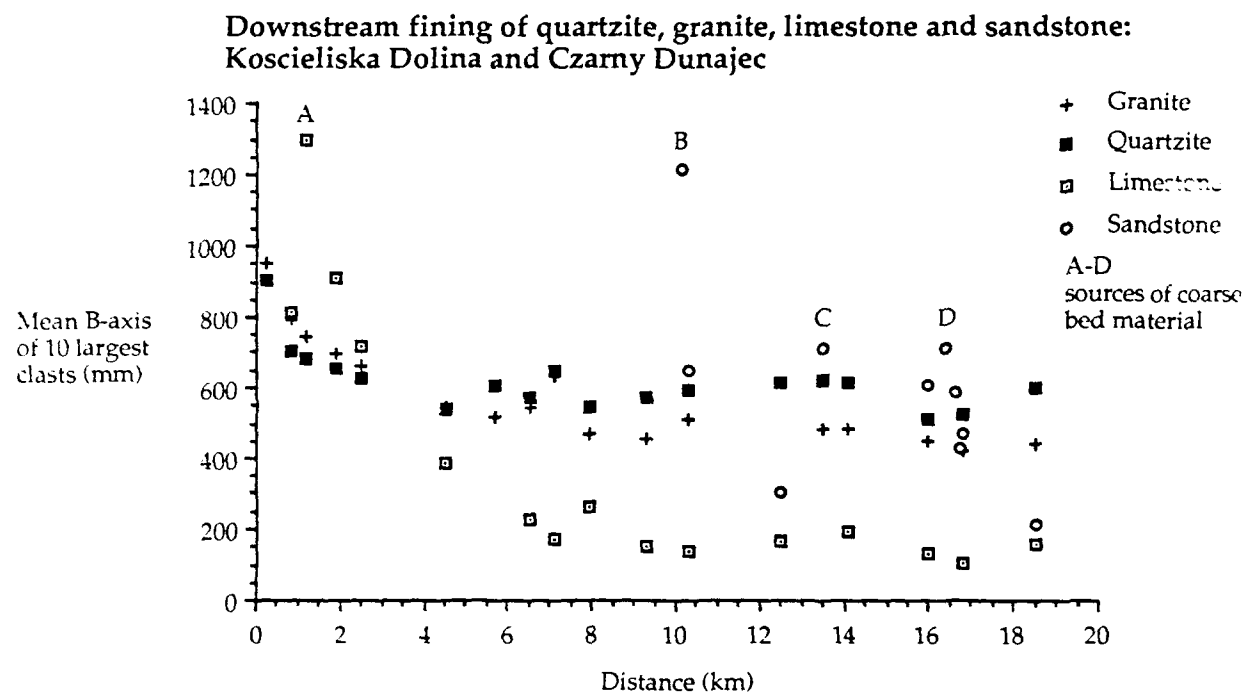
Figure 3   Downstream fining of all four lithologies for Koscieliska Dolina - Czarny Dunajec
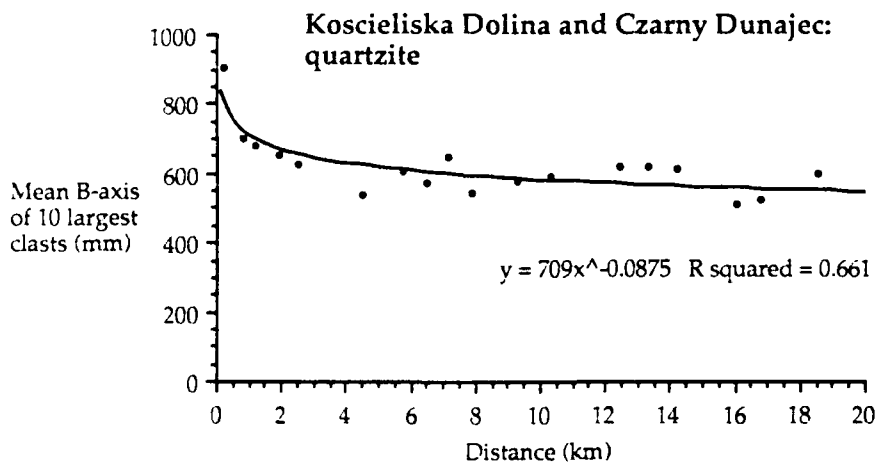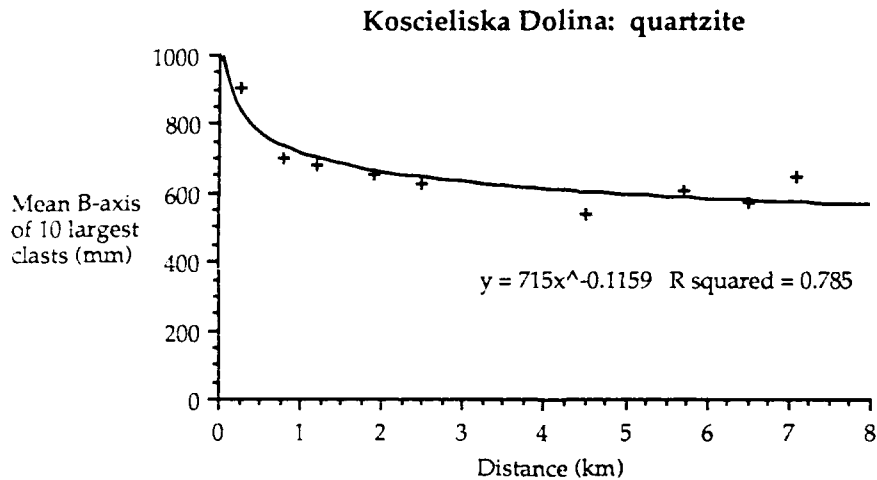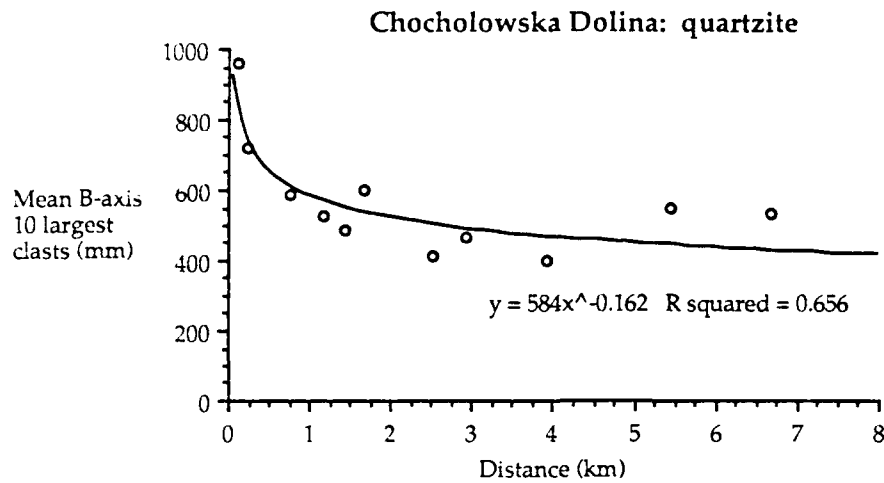
Figure 4   Downstream fining of quartzite

Chocholowska Dolina: granite

Mean B-axis of 10 largest clasts (mm)

$y = 625x^{-0.183}$ R squared= 0.732

Distance (km)

Koscieliska Dolina: granite

Mean B-axis of 10 largest clasts (mm)

$y = 758x^{-0.1652}$ R squared = 0.881

Distance (km)

Koscieliska Dolina and Czarny Dunajec: granite

Mean B-axis of 10 largest clasts (mm)

$y = 758x^{-0.1864}$ R squared = 0.922

Distance (km)
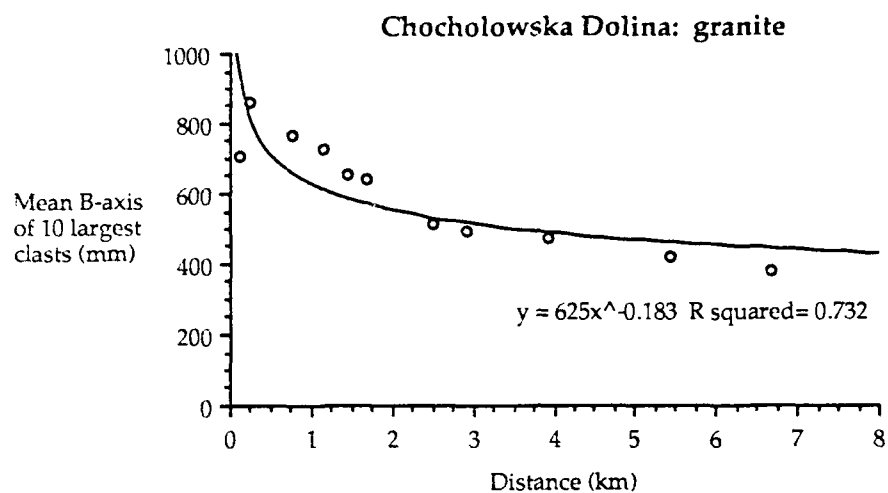
Figure 5   Downstream fining of granite

Chocholowska Dolina: limestone

$$y = 516x^{-0.326} \quad R\text{ squared} = 0.865$$

Mean B-axis of 10 largest clasts (mm) vs Distance (km)

Koscieliska Dolina: limestone

$$y = 1787x^{-1.1068} \quad R\text{ squared} = 0.980$$

Mean B-axis of 10 largest clasts (mm) vs Distance (km)

Koscieliska Dolina and Czarny Dunajec: limestone

$$y = 1124x^{-0.7894} \quad R\text{ squared} = 0.888$$

Mean B-axis of 10 largest clasts (mm) vs Distance (km)
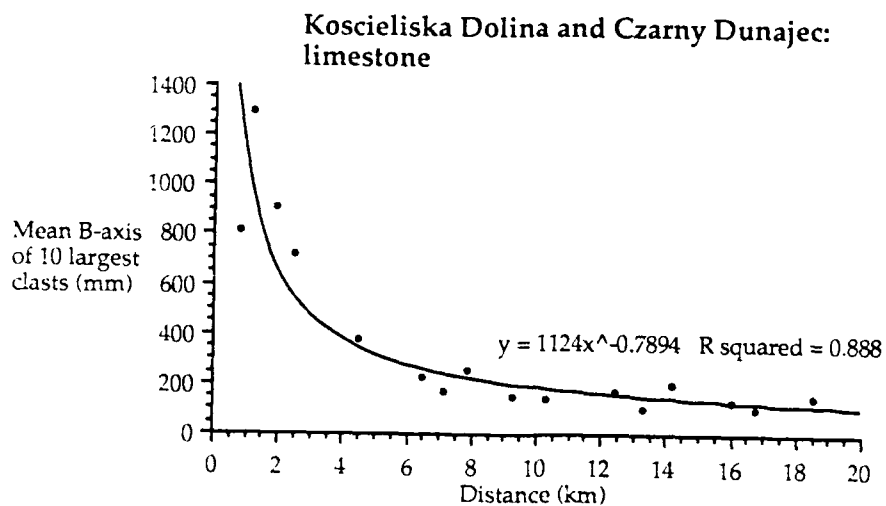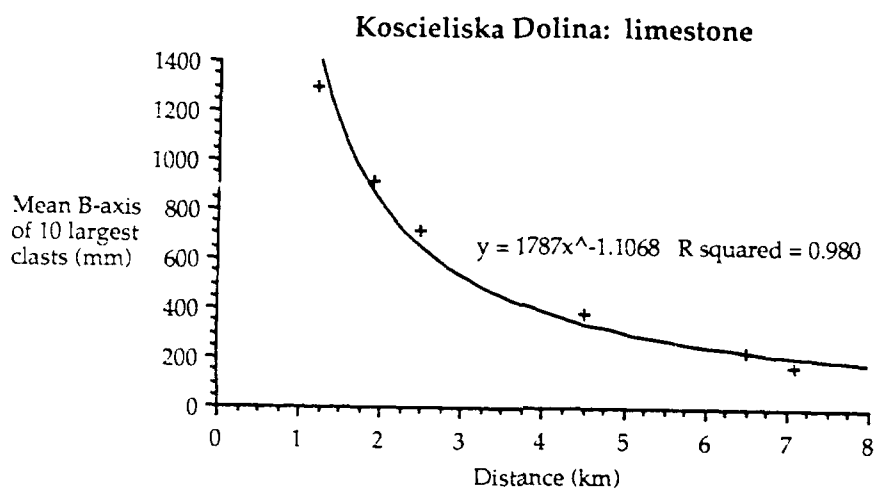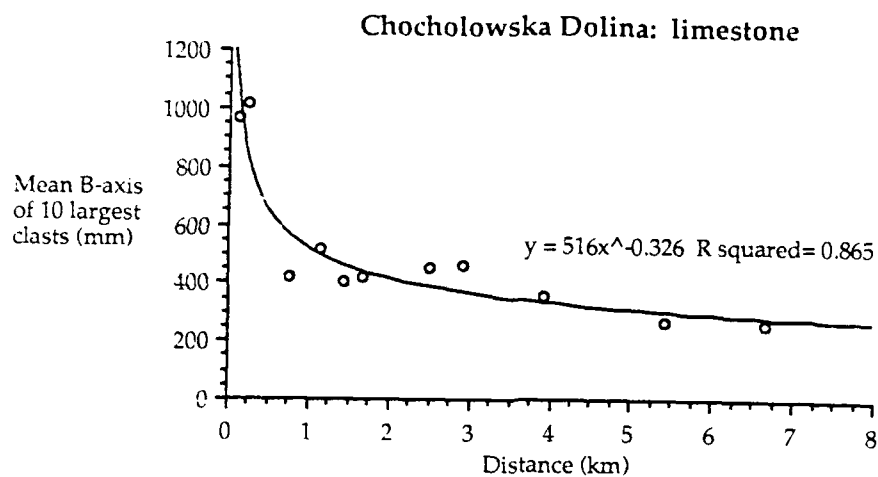
Figure 6   Downstream fining of limestone

Downstream fining of sandstone from a
series of coarse inputs: Czarny Dunajec

1400

A
1200

A, C, D
sources of coarse
1000 bed material

Mean B-axis   800   C   D
of 10 largest   B
sandstone clasts
(mm)   600   E

400

200

0

0   2   4   6   8   10   12   14   16   18   20
                    Distance (km)

Figure 7   Downstream fining of sandstones from a series of coarse inputs

Downstream change in sphericity:
Koscieliska Dolina

$y = 0.666 + 0.0084x$

R squared = 0.742
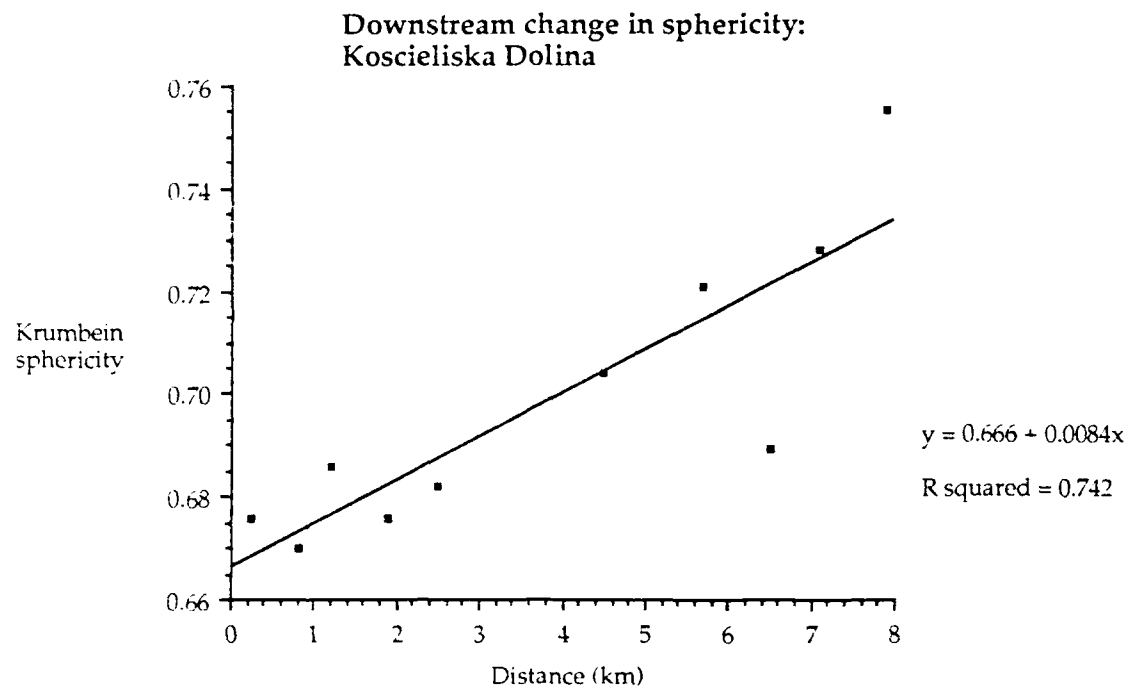
Figure 8   Downstream changes in sphericity: Koscieliska Dolina

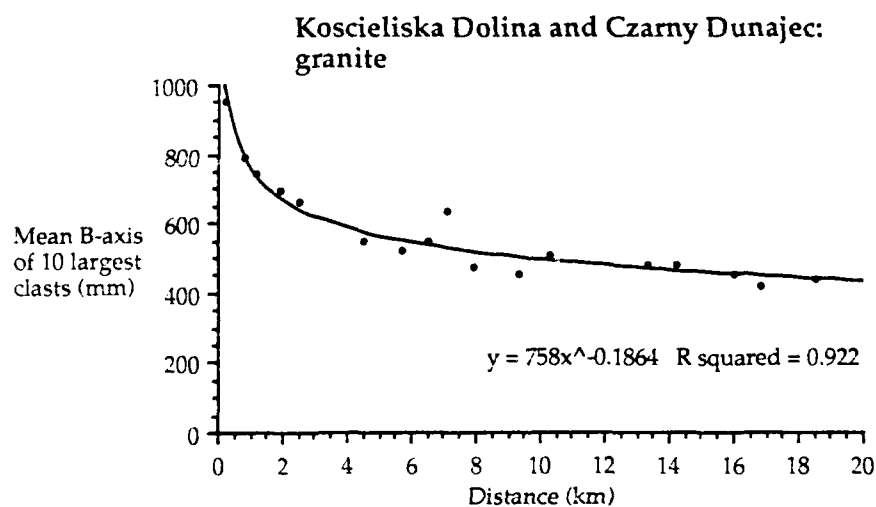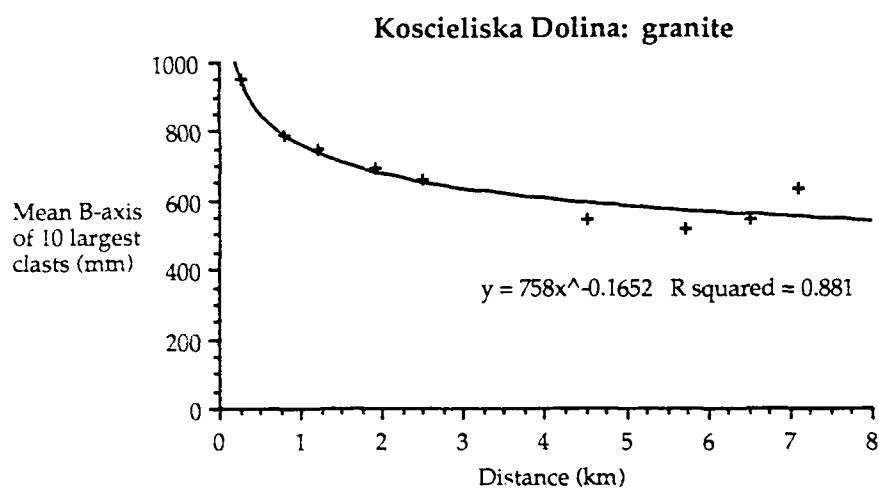# VARIATION OF BED AND TRANSPORT MEAN DIAMETERS
IN EROSION-DEPOSITION PROCESSES

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ABSTRACT

The paper presents a mathematical scheme able to model the evolution of deposition-erosion processes in a one-dimensional stream, also taking account of non uniform grain size mixtures and possibly non-equilibrium of sediment transport rate.

Instead of describing the sediment transport for fractions, that is, each fraction by means of its continuity equation. the model uses mean dia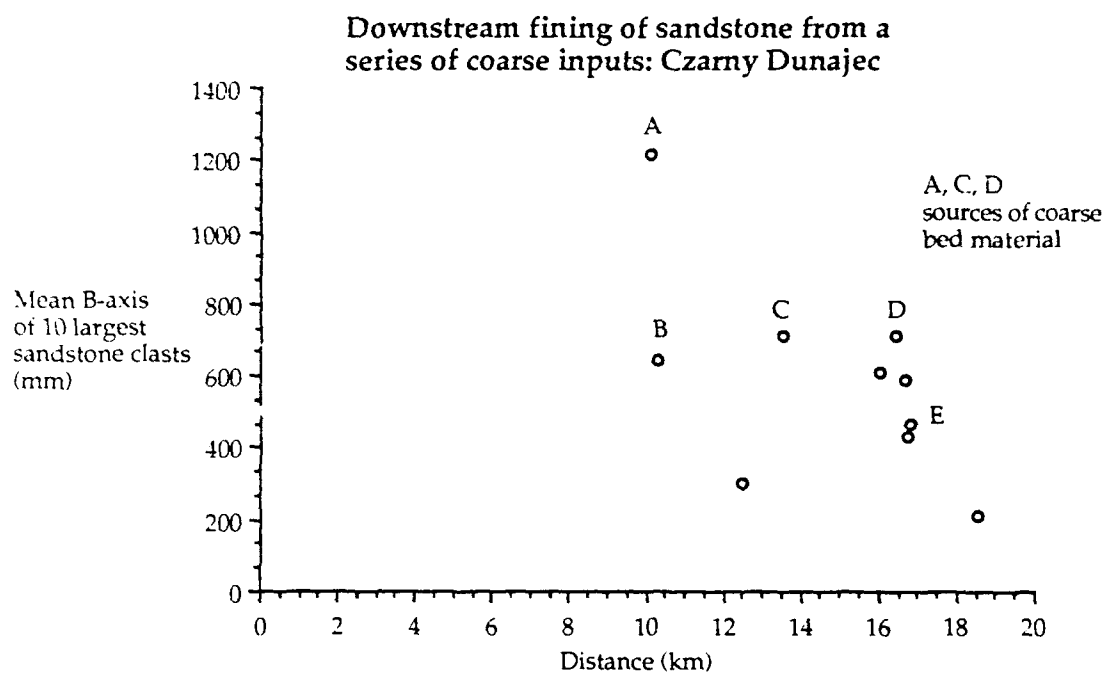meter and second-order (and possibly third-) moment of grain size distribution. As a consequence the number of equations and variables is reduced.

INTRODUCTION

The problem of aggrading and degrading of rivers has been strongly emphasized in the recent years. The interest in solving this problem is related not only to a certain unresolved problems in the field of the research. but also to a better insight of long and short time consequences of river training engineering.

A number of numerical models has been proposed in the literature in order to describe the sediment transport of non uniform grain size material, some of them even able to account for non-equilibrium transport rate (Ashida and Michiue, 1971) (Karim and Kennedy, 1982) (Armanini and Di Silvio 1982 and 1988) (Rahuel and Holly, 1989) (Andrews aand Parker 1987).

The general approach adopted in order to consider the different behavior of different grain-size materials is based on the idea of dividing the grain-size distribution curve into a discrete number of classes and describe the behavior of each class by means of separate mass- or volume- balances.

Following the procedure adopted by Einstein(1950). the actual transport rate of each class is related to the percentage present in the bed sediment transport.

Aim of this paper is to approach the problem in terms of first-, second- and possibly third-order moments of the grain-size distribution.

1

GENERAL FORMULATION OF THE PROBLEM

The continuity equation for sediment transport in a quasi one-dimensional stream is given by :

$$\frac{\partial T}{\partial x} + \frac{\partial Ch}{\partial t} = -\frac{\partial Z}{\partial t}$$

1)

where h and Z are water depth and bottom elevation; $T = \int ucdy$ is the actual transport rate and $C = \int cdy$ is the depth-averaged concentration of sediments.

In natural rivers, but also in many laboratory settings, the bed material is far different from uniform.

In this case the transport rate of each individual grain fraction depends on its size. Nevertheless the continuity within each grain size class must be established.

Following Einstein, the ratio between the actual transport rate T_j for each grain size fraction j, and its transport capacity T_{c_j} (that is the transport rate in case of steady flow and uniform grain size material) is equal to the percentage β_j of the same class present in the bed. That is :

$$T_j = \beta_j T_{c_j}$$

2)

This assumption, referred as *similarity hypothesis*, exhibits at least two limitations. The first one is that it refers to (local) equilibrium condition; that is, it does not account for different adaptation processes due to unsteadiness of sediment transport. The second limitation lies in the fact that it does not account for hiding effect, that is the sheltering exerted by the coarser materials over the finer fraction.

As far as the sediment transport is concerned, a linear adaptation model can generally describe the process (Armanini and Di Silvio 1988)as in the following eq.(3):

$$T_j = \beta_j T_{c_j} - L_j^* \frac{\partial}{\partial x} T_j - \tau_j^* \frac{\partial}{\partial t} T_j$$

3)

L_j^* and τ_j^* represent, respectively, the adaptation lenght and the adaptation time for the same class.

L_j^* accounts for the space lag between the actual transport rate T_j and the transport capacity $\beta_j T_{c_j}$: this effect is more pronunced the smaller the grain size of the class. For suspended load the following expression has been proposed by Armanini and Di Silvio (1988).

$$\frac{L^* w}{hU} = \frac{a}{h} + (1 - \frac{a}{h}) \exp\left[-1.5(\frac{a}{h})^{-1/6}\frac{w}{u_*}\right]$$

4)

where w is the particles falling velocity, a is the thickness of the bottom layer, h and U water depth and mean velocity.

In eq.3) τ_j^* accounts for time lag between the same quantities as above. For suspended load one can assume:

$$\tau^* = \frac{L^*}{U}$$

5)

For bed load, adaptation length and adaptation time are small (e.g. of order of magnitude of bed forms), so in general it is possible to assume immediate adaptation of bed load to equilibrium condition.

This assumpion is generally accepted in case of gravel bed rivers where the bed load prevails over suspended load.That is:

$$L^* \sim \tau^* \sim 0$$

6)

It should be remarked that β_j is the percentage of j class present on the bed surface. Experimental evidence in fact shows, under the layer where sediment transport occurs, the existence of a *surface layer*, where the longitudinal transport is negligeable, but where the vertical exchange of sediments with the

transport layer is determinant. This exchange is in fact responsible for the sorting process of the bed material.

The thickness of such a mixing layer (or surface layer) δ is order of magnitude of roughness elements. In duned beds it can be assumed as half the dune height Δ:

$$\delta = \frac{1}{2}\Delta \qquad 7)$$

As far as hiding effect concernes, it is possible to account for this effect by introducing a proper hiding factor. This factor reduces the transport capacity of finer fraction of the mixture and increases the transport capacity of the coarser fraction. Different expression for the hiding factor have been proposed in the literature, suitable to be applied to different kind of sediment transport formulas (Eghiazaroff 1965),(Einstein 1950),(Day 1980),(Ranga Raju 1985),(Andrews and Parker 1987). In general the hiding factor is a function of the ratio between the diameter d_j of j class and the mean diameter $\bar{D}$ of the mixture:

$$\xi_j = \xi_j (\frac{D_j}{\bar{D}})^m$$

The mixing process among the classes constituting the bed material may be described by means of the continuity equation inside the mixing bed layer δ (where just mixing but not longitudinal transport takes place), as described by the following eq.8).

$$\frac{\partial}{\partial t}\delta\beta_j = \beta_j^* \frac{\partial(\delta - Z)}{\partial t} - (\frac{\partial T_j}{\partial x} + \frac{\partial}{\partial t}C_j h) \qquad 8)$$

In the erosion processes, β_j^* represents the percentage in the subsurface layer, (the volume below the mixing layer) of the material that belongs to the j-th class ; while during the deposition phase $\beta_j^* = \beta_j$.

Finally the total transport T is the summation of the transport rates of each class j , that is :

$$T = \sum_{j=1}^{m} T_j \qquad 9)$$

If m is the number of classes, the set formed by eqs. 1), and 9), and n eqs.2) (or eqs.3) and 8), in the $2 \cdot n + 2)$ unknowns $Z, T, m \cdot T_j$, $m \cdot \beta_j$, (coupled with the appropriate hydrodynamics equations), are sufficient to describe the behavior of each class of sediments.

The relationship beetween C_j and T_j is given by the transport formula and concentration distribution; it is generally possible to put:

$$C_j = \frac{T_j}{U_j}$$

being U_j the celerity of the $j - th$ class of the grain size distribution.

The set of equation can be numericaily solved for a discrete number of classes. Many numerical models have been recently proposed in literature (Andrews and Parker 1987), (Armanini and Di Silvio 1988), (Rahuel and Holly 1989), (Bouchard et alii. 1989). The models differ for the sediment transport formula, for the expression of adaptation quantities L_j^* and τ_j^*, for the definition of the mixing layer, as well as for numerical scheme and number of classes. Since now the number of classes adopted in literature ranges from 2 to 5. From a numerical point of view the models are quite complex owing to the number of (non linear) equations: one possibility to simplify the problem is to write the equations in terms of mean values instead of single grain-size quantities.

EQUATIONS IN TERMS OF MEAN DIAMETERS

3

Adding up to n classes all the terms in eq.2, one obtains:

$$T = \sum_{j=1}^{m} \beta_j T_{c_j}$$

10)

Given that β_j is a distribution function, it vanishes as when the diameter tends to zero and infinity and the summation in eq.10) can be calculated expanding in Taylor series T_{c_j}.

If transport capacity for each class j is expressed by a formula linking it to the power of respective diameter D_j, one obtains:

$$T_{c_j} = A D_j^{-n}$$

11)

Expanding 11) in Taylor series, the second term of eq. 10) can be written as:

$$\sum_{j=1}^{m} \beta_{c_j} T_{c_j} = T_{cb}[1 + c_2\sigma_b + c_3\gamma_b + O(\kappa_b)]$$

12)

where T_b is the transport capacity related of the mean diameter of the bed material :

$$D_b = \sum_{j=1}^{m} \beta_j D_j$$

13.a)

σ_b, γ_b and κ_b are second-, third- and fourth-order dimensionless moments of the β_j distribution (variance, skewness and kurtosis)

$$\sigma_b = \sum_{j=1}^{m} \frac{\beta_j (D_j - D_b)^2}{D_b^2} \quad ; \quad \gamma_b = \sum_{j=1}^{m} \frac{\beta_j (D_j - D_b)^3}{D_b^3} \quad ; \quad \kappa_b = \sum_{j=1}^{m} \frac{\beta_j (D_j - D_b)^4}{D_b^4}$$

13.b)

The coefficients c_2 and c_3 are, respectively:

$$c_2 = \frac{n(n+1)}{2!}$$

14.a)

$$c_3 = \frac{-n(n+1)(n+2)}{3!}$$

14.b)

After substitution of eqs. 12) into eq.10), one obtains:

$$T = T_{bc}[1 + c_2\sigma_b + c_3\gamma_b + ...]$$

15)

The terms in eq 2) are now multiplied by the respective diameter D_j, and summed up to the n classes, so that

$$\sum_{j=1}^{m} T_j D_j = \sum_{j=1}^{m} \beta_j D_j T_j$$

16)

Introducing a mean transport diameter D_t, defined as follows:

$$D_t = \sum_{j=1}^{m} \alpha_j D_j$$

17)

where α_j is the percentage of the class j present in the transport layer; by definition it is:

$$T_j = \alpha_j T$$

18)

For the first term of eq.16), one has :

1

$$\sum_{j=1}^{m} T_j D_j = T \sum_{j=1}^{m} \alpha_j D_j = T D_t \qquad 19)$$

Following a procedure similar to that used in eq.12), one obtains for the rigth hand side term of eq. 16):

$$\sum_{j=1}^{m} \beta_j D_j T_j = D_b T_{cb}[1 + c_1 \sigma_b + c_2 \gamma_b + \ldots] \qquad 20)$$

where :

$$c_1 = \frac{n(n-1)}{2!} \qquad 21)$$

After substitution of eqs. 19) and 20) into eq. 16), one has:

$$T D_t = D_b T_{cb}[1 + c_1 \sigma_b + c_2 \gamma_b + \ldots] \qquad 22)$$

Finally multiplying eq.5) by the respective diameter D_j and summing up the n classes, gives:

$$\frac{\partial}{\partial t} \delta D_b = D_b^* \frac{\partial}{\partial t} \delta - Z - \frac{\partial}{\partial x} T D_t - \frac{\partial}{\partial t} \frac{T D_t}{U_t} \qquad 23)$$

Assuming that the variance and the skewness γ_b of β_j remains unchanged, the set formed by eqs 1), 15), 22) and 23), in the unknowns T, Z, D_t, D_b is complete and can be solved numerically or, in some simplified cases, analytically. If the latter hypothesis proves unsatisfactory, an equation for the second moment can be otained just multiplying each term of eq. 10) by D_j^2 and repeating the same procedure as above.

When the bed load prevails over suspended load, as in gravel bed rivers, it reasonable to assume:

$$\frac{\partial}{\partial x} T D_t \gg \frac{\partial}{\partial t} \frac{T D_t}{U_t} \qquad 24)$$

The complete set of equations it this case will be the following:

$$\frac{\partial T}{\partial x} = -\frac{\partial Z}{\partial t} \qquad 25.a)$$

$$T = T_{cb}[1 + c_2 \sigma_b + c_3 \gamma_b \ldots] \qquad 25.b)$$

$$T D_t = D_b T_{cb}[1 + c_1 \sigma_i + c_2 \gamma_b + \ldots] \qquad 25.c)$$

$$\frac{\partial}{\partial t} \delta D_b = D_b^* \frac{\partial}{\partial t}(\delta - Z) - \frac{\partial}{\partial x} T D_t \qquad 25.d)$$

5

It should be noted that a similar procedure can also account for the adaptation process (see eq.3). In this case an adaptation lenght and time relevant to mean transport diameter will appear (Armanini, 1989).

Combining eqs. 25.b and 25.c) one has:

$$\frac{D_t}{D_b} = \frac{1 + c_1\sigma_b + c_2\gamma_b + \cdots}{1 + c_2\sigma_b + c_3\gamma_b + \cdots}$$

26)

Eq. 26) states that the ratio between the mean diameter of the transport load and mean diameter of the surface bed material , is function of the moments of the bed load distribution and of the transport formula, (trough the coefficients c).

If transport formula can be expressed as a power law (see eq. 11), the above ratio depends just on the exponent n . In fact substituting eqs. 14) and 21) into eq 26) (neglecting $O(\gamma_b)$ terms) gives:

$$\frac{D_t}{D_b} = \frac{2 + c_1\sigma_b n(1-n)}{2 + \sigma_b n(1+n)}$$

27)

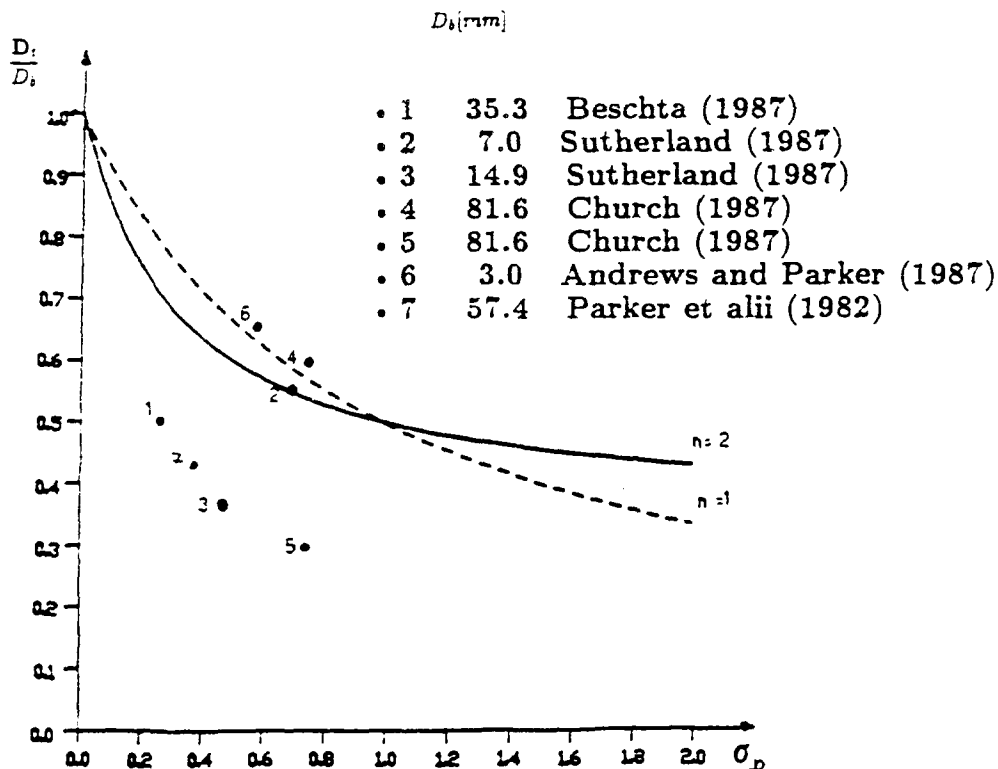|		
• 1	35.3	Beschta (1987)
• 2	7.0	Sutherland (1987)
• 3	14.9	Sutherland (1987)
• 4	81.6	Church (1987)
• 5	81.6	Church (1987)
• 6	3.0	Andrews and Parker (1987)
• 7	57.4	Parker et alii (1982)

Fig.1 The ratio between mean diameters of bed-load and surface bed material are compared to eq.27)

6

In fig. 1, eq. 27) is compared to some experimental data.

The value $n = 0$ corresponds to equal mobility condition; in this case, infact, $D_t/D_b \mapsto 1$; the same result is obtained when $n \mapsto \infty$ (absence of sediment transport).

The value $n = 1$ for the exponent in eq. 11) is a reliable value accaunting for hiding effect.

The two curves in Fig.1 correspond respectively to $n = 1$ and $n = 2$: for small values of σ_b , the difference between the two curves is not striking. In the same figure almost 3 points lie definitely above the curves. All these points correspond to a static armouring situation.

From eq. 27) one has that no real values for exponent n can match these experimental points.
On the other hand, it must be remarked that the power low (eq. 11) for sediment transport formula can be used if the gradation curve is not very wide, and , above all, if sediment transport is far from incipient motion condition.

Instead of eq.11, it is possible to use a binomial expression (e.g. Mayer-Peter Mueller type equation):

$$T_{c_j} = const_1 \sqrt{g} D_j^{1.5} \left(\frac{u_*^2}{g D_j} - \xi_j \cdot const_2 \right)^{1.5} \qquad 28.a)$$

Eq. 28 can be generalized in the following form:

$$T_{c_j} = (B_2 - A_2 D_j^{n_1})^s \qquad 28.b)$$

where B_2 and A_2 depend on hydrodynamic conditions, but not on D_j.
In this case the coefficients c_2 and c_2 will be:

$$c_1 = -\frac{ns A_2 (1-n) D_b^n}{2!(B_2 - A_2 D_b^n)^2}(-B_2(1+n) + A_2(1+ns)) \qquad 29)$$

$$c_2 = c_1 - \frac{ns A_2}{B_2 D_b^n - A_2} \qquad 30)$$

Eqs. 29) and 30), inserted in eq. 26), allow for any solution for $D_t/D_b < 1$. provided that the correct sediment transport formula is given.
Unfortunately eqs. 29) and 30) are more difficult to be used. because they require the knowledge of the sediment transport formula (this quantity does not appear in eq. 27). Comparison between theory and literature experiments is under way.

Eqs.29) and 30) must be used instead of eq. 14.a) where the gradation curve is wide.

When $n \mapsto 0$ in eqs. 29) and 30) (equal mobility), $D_t/D_b \mapsto 0$. The same results when transport rate tends to zero.
Similar conclusions can be drawn for the coefficient c_3. More in general c_1, c_2 and c_3 are function of D_b.

From eqs. 25.a) - 25.d), algebraic manipulation (and neglecting skewnes terms) yields:

$$\delta \frac{\partial}{\partial t} D_b = D_b^* \frac{\partial}{\partial x} T_{cb}[1 + c_2 \sigma_b] - \frac{\partial}{\partial x} D_b T_{cb}[1 + c_1 \sigma_b] + (D_b^* - D_b)\frac{\partial \delta}{\partial t} \qquad 31)$$

In aggrading processes. (deposition) D_b^* is equal to D_b, and eq.31) leads to:

$$\frac{\partial}{\partial t}\delta D_b + T_{cb}[1 + c_1\sigma_b]\frac{\partial D_b}{\partial x} = D_b\frac{\partial}{\partial x}T_{cb}\sigma_b[c_2 - c_1] \qquad 32)$$

and:

$$\frac{\partial}{\partial t}\delta D_b + [T_{cb}(1 + c_1\sigma_b) - nD_b\sigma_b\frac{\partial T_{cb_j}}{\partial D_b}]\frac{\partial D_b}{\partial x} = nD_b\sigma_b\frac{\partial U}{\partial x}\frac{\partial T_{cb}}{\partial U} \qquad 33)$$

Eq.33) gives the following celerity, for the propagation for the mean diameter:

$$\tilde{c}_{D_k} = \frac{[T_{cb}(1 + c_1\sigma_b) - nD_b\sigma_b\frac{\partial T_{cb}}{\partial D_b}]}{\delta} = \frac{T_{cb}}{\delta}[1 + \sigma b\,(c_1 + n^2)] \qquad 34)$$

In nonuniform flow, the celerity is obviously non constant, and increases with increasing velocity. The thickness δ of the mixing layer reduces then the celerity.

CONCLUSIONS

This mathematical model can describe the behavior, in one-dimensional channels, of a mixture of materials having non uniform grain size distribution.

Compared to complete schemes, where each fraction of the gradation curve is described by one (or two) partial differential equation(s), the model is based on mean diameters and variance of grain size distribution (and if necessary, of the other moments of the distribution).

From a numerical point of view the model, reducing the number of variables, represents a relevant simplification.

some questions are still open in particular concerning:

-the role of the hiding factor (this is a open point also for the other models);

-the possibility of describing the armouring processes (static armouring);

- the legitimacy of keeping variance (and other moments) constant in space an time;

- the legitimacy of neglecting higher order moments.

Comparison between the proposed scheme and literature experiments, as well as comparison with complete models, are under way.

Finally an improvement in the model can be obtained, writing the moments in terms of logaritm of the particles diameter instead of the diameter itself.

NOTATIONS

The following symbols have been used in this paper:.

A_2, B_2	parameters of sediment transport formula
C	concentration
c_1, c_2, c_3	coefficients
$\bar{c}_{D_b}$	celerity of transported mean diameter
D_j	mean diameter of class j-th
D_b	mean diameter in the mixing layer
D_t	mean diameter of the transported material
h	water depth
L^*	adaptation length
m	number of grain size classes
n	exponent
x	longitudinal coordinate
t	time
T_j	transport rate of j-th grain size class
T_c	transport capacity of the j-th grain size class
U	depth averaged velocity
$U_j = \frac{T_j}{C_j}$	
u_*	friction velocity
w	particle fall velocity
Z	bed elevation
α_j	percentage of the j-th class inside the transport layer
β_j	percentage of the j-th class into the exchange layer
δ	exchange layer thickness
Δ	heigt of the dunes
σ_b, γ_b	second and third moments of grain size distribution inside the mixing layer
σ_t, γ_t	second and third moments of grain size distribution of transported material
τ^*	adaptation time
ξ_j	hiding factor for j-th grain size class

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THE MYSTERY OF BIMODALITY: FALSIFICATION OF CURRENT CONCEPTS, PROPOSAL OF A NEW ONE

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Preface

This paper is part of an extensive study of provenance in Calabria, South Italy (Fig. 1). Bimodality is only one of many aspects treated in this context. However, the broad background of quantitative information in Calabria makes it possible to check elements of bimodality against most of the variables assumed to be relevant for this problem.
The study area: The Ionian slope of the Calabrian Massif is an active plate margin, 19 basins drain the 1850 km^2 research area (area 16-160 km^2, maximum length 20 km, maximum elevation 2000 m). Four of these 19 braided rivers reach the seventh order, the other 15 rivers are of the sixth order. Drainage density ranges from 5.1 1/km to 10.36 1/km, bifurcation ratios vary between 3.8 and 5.8. The relief (altitude standard deviation) ranges between 213 and 481 m, the slopes of the main trunks vary between 1:30 and 1:110. Mean annual precipitation is between <600 mm and >2000 mm, rainstorms with precipitations of more than 400 mm a day are the most striking climatic events of the area.
The hinterland comprises granite (24 % area), gneiss (7 %), schist (30 %), limestone (1 %), conglomerate (9 %), sandstone (9 %), siltstone (12 %) and argillite (8 %). Mean longterm (1 Ma) erosion rate is 206 mm/ka, mean shortterm rates (50 a) are 1700 mm/ka, calculated over landslide activity, geodetic survey of long profiles and the refill of a landslide-created lake.

Summary

Calabrian river-mouth sediments are both bimodal and unimodal. The gravel mode of the bimodal distributions, mostly Rosin-distributed, is interpreted as a first order element, a primary feature inherited from the jointed and weathered source rocks whereas the sand mode can be seen as a second order element acquired during fluvial attrition and transportation.
If the percentage of the fraction 1-20 mm exceeds about 35 % the distributions became unimodal. So this fraction is crucial for modality and it is called, consequently, a third order element.
All single factors assumed to control modality such as relief, basin shape, source rock-petrography, the petrographical composition of the river-sediment itself or postdepositional refill of interstices fail to exert a significant or mono-causal influence on modality. In addition, no hydrodynamic conditions are known which specifically select, as a rule and within the grain-size range of gravel and sand, this fraction of 1-20 mm . Thus, other possibilities having excluded, only availability
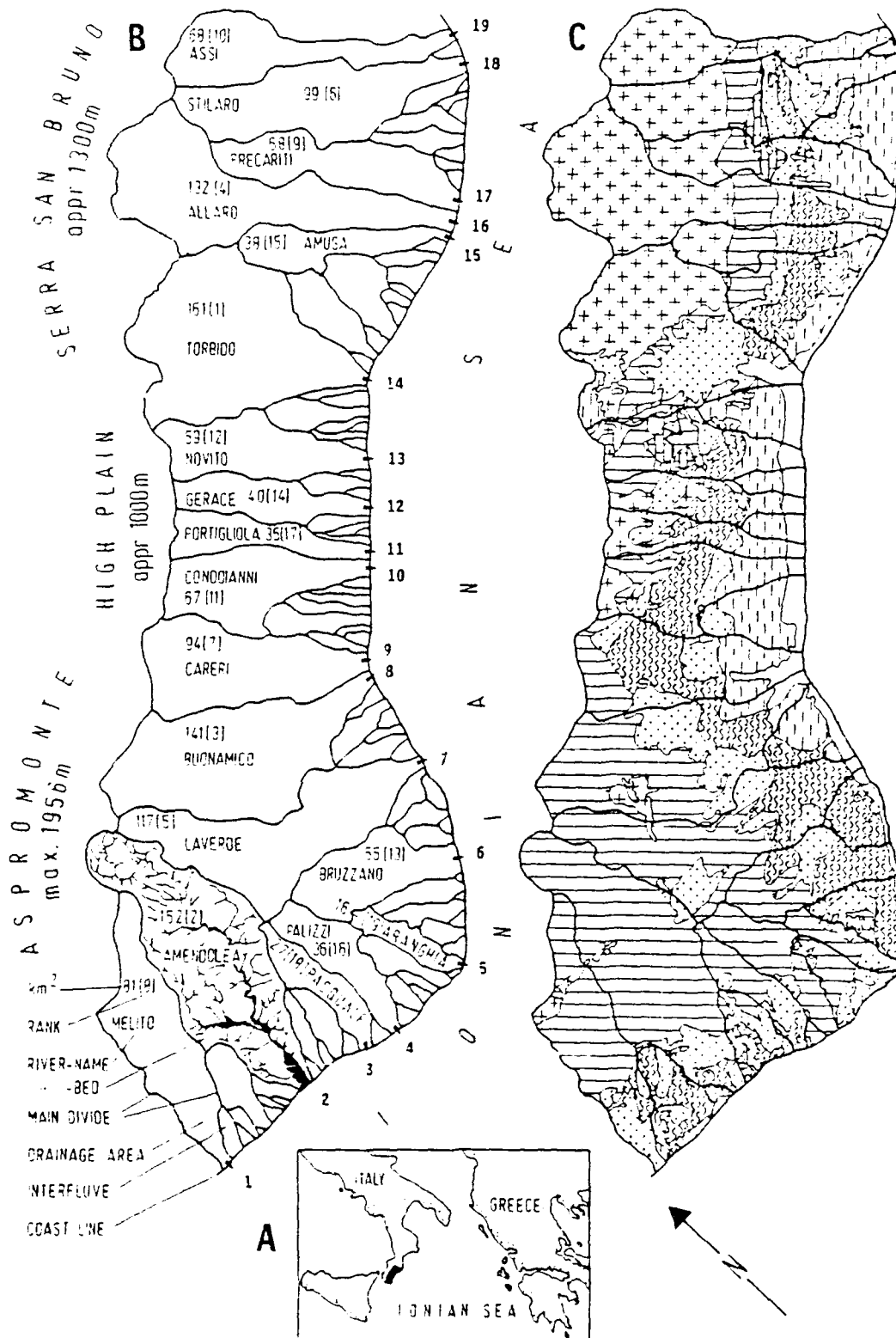
Figure 1. **A**: Position of the study area. **B**: The 103 basins and
interfluves and some data of the 19 rivers studied in more
detail. **C**: Regional distribution of the most important rock-
families: Granite (++++); metamorphic rocks (———);
limestone (⊥⊤⊥⊤); sandstone and conglomerate (::::);
siltstone (≈≈≈) and argillite (- - -)

remains, but not in the sense of source-area related sediment
production. It is assumed that different floods drive the "se-
dimentary mill" of these mountain rivers differently and that
the formation of the fraction 1-20 mm is sensitive to these
differences.

Data base: 133 samples, jointed and weathered source rocks
(in situ, 25 samples), transition (regolith, talus debris,
alluvial cone, 22), river-course (downstream of the fall
line, 17), river-mouth (35) and beach (shoreface, 34)

Method: The mean sample size is 270 kg, the maximum grain-
size 256 mm. The samples were taken as channel samples by
means of a crowbar (handspike) or motor hammer. The frac-
tion >16mm was sieved in the field by means of a trans-
portable sieving and splitting machine in 1/3 phi steps
(Ibbeken 1974), the fraction <16 mm was split and taken to
the lab. The results of the sieve analyses are stored in a
data bank, like all other data, i.e. digitized maps and
analyses of the river-mouth sediments.

Unimodal and bimodal distributions

The problem of bimodality or, more generally, of deficiencies
of certain grain-size classes within grain-size distributions
of terrigenous sediments, expressed by "kinks" of the cumula-
tive frequency distribution, finds much interest among sedimen-
tologists. We refer exclusively to the deficiencies of sandy
gravel deposits somewhere in the range of 1 to 20 mm. Shea
(1974) gives an overview of the problem. He states that most
contributions to the deficiency complex are not accompanied by
statistically rigorous techniques and that many theories are
simply contradictory. Shea presents a histogram of an overall
weighed average of 11,212 analyses, which shows no deficien-
cies. However, many of our modern Calabrian deposits display
both evidently bimodal and unimodal distributions and our know-
ledge of the source area conditions should lead us to a better
understanding of the problem. This is of special interest, be-
cause some authors believe bimodality to be a provenance fac-
tor. But this can be only partly true. As pointed out by Ibbe-
ken, 1983, only the shape of the gravel mode has to be seen as
an inherited element, whereas the sand mode represents an ac-
quired one, created mainly during fluvial transportation. Howe-
ver, bimodal or unimodal, in Calabria this depends mainly on
wether the amount of the fraction 1-20 mm exceeds 35 % or not.

The gravel mode (first order element)

Gravel modes of bimodal distributions in Calabria are mostly
Rosin- instead of lognormally distributed (Ibbeken 1983). The
"RG-ratio", the Rosin/Gauss ratio, stands for the quotient of
the number of Rosin- to Gauss- distributed samples of an envi-
ronment. In Calabria, the jointed and weathered source rocks
yield a "RG-ratio" of 4, transition (3.6), river-course (5),

river-mouth (8) and beach (0.65). This means that the Rosin-tendency increases towards the river-mouth and it is only the beach environment which creates a symmetrization.

The sand mode (second order element)

The sand budget of both bimodal and unimodal river-mouth sedi-ments is exactly balanced, the average sand content is 24.95 %
(bimodal sediments) and 25.19 % (unimodal sediments), 60 % of
the bimodal and 61.9 % of unimodal river-mouth sediments have
sand contents under 25 %. In single cases the sand content va-ries considerably, between about 10 % and 50 %. Thus, compared
to the never missing gravel mode, the Calabrian sand modes are
a relatively unstable feature.
The mean sand content increases continuously from the jointed
and weathered source rocks (4.6 %) over the transitional stage
(13.6 %) and the river-course (16.5 %) to the river-mouth (28
%).
Calabrian sand modes are almost exclusively lognormally distri-buted. Ibbeken (1983) interpreted this as a transport-specific,
acquired property. This implies that the sand comes from rock
disintegration and soil formation. Geochemical evidence makes
this argument refutable: Constant oxide- and gain and loss cal-culations of the geochemical composition of the source rocks,
source soils and river-mouth sediment, comparing the 63-500 μm
fraction, demonstrate that the river-mouth sediment resembles
the fresh source rock rather than the soils. It is probable,
from this, that attrition during bankfull stages of these po-werful, high gradient mountain rivers is the most important
process of sand formation in Calabria.

The 1-20 mm fraction (third order element)

Figure 2A shows the average of 51 bimodal distributions. The
gravel modes (A1) shift from 12.5-128 mm, the sand modes (A2)
vary between 0.135 and 1.6 mm, but the position of the "valley"
in between (A3), much more stable, ranges only between 1.25 and
4 mm. The "kink" of the 1.6-2 mm class stems from sample split-ting at 2 mm.
Figure 2B displays the average of 22 unimodal distributions,
the much less peaked modes vary between 8 and 80 mm.
As pointed out, the sand content of both bi- and unimodal dis-tributions is well balanced, ranging from about 10 to 50 % with
an average of 25 %, regardless of bi- or unimodal distributi-ons. Thus it is not the amount of sand which is responsible for
modality. This is demonstrated by Figure 2C. Here the diffe-rence between the two types of distribution is calculated sim-ply by substracting the mean unimodal from the mean bimodal
distribution. It turns out that the fraction 1-20 mm is the es-sential one which makes the difference between bi- and unimodal
distributions. It is called, consequently, a third order ele-ment (Ibbeken 1983, p.1230). Figure 5A shows that most unimodal
rivers contain more than 35 % of this fraction, the Bruzzano
[06] excepted. Ibbeken (op.cit., Fig. 8C) demonstrated that it
is possible to split the average Calabrian unimodal distribu-tion into a Rosin-distributed gravel mode (first order), a log-
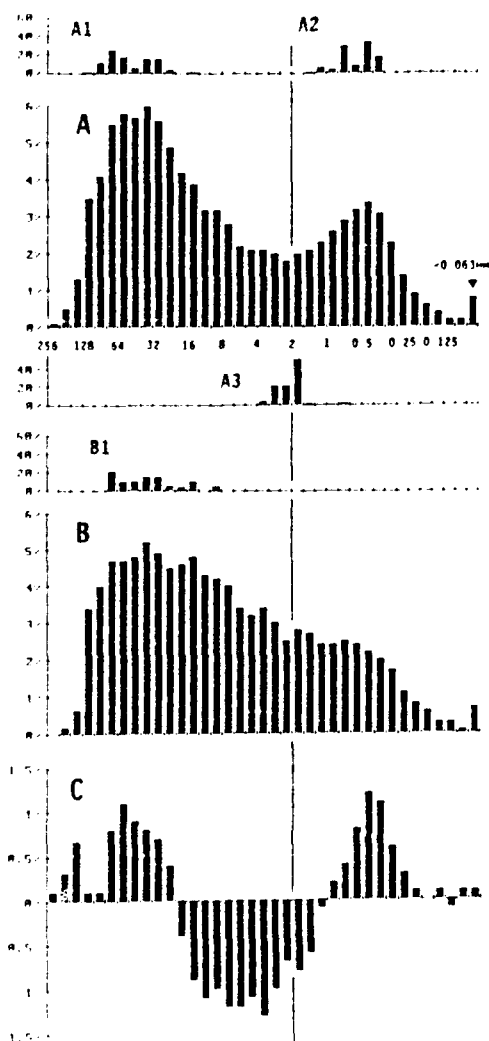
Figure 2. **A:** Average of 52 bi-
modal distributions;
A1: Distribution of the gravel
modes;
A2: Distribution of the sand
modes;
A3: Distribution of the "val-
leys" (lowest percentage) in
between;
B: Average of 22 unimodal dis-
tributions;
B1: Distribution of the gravel
modes;
C: Average bimodal distribution
(A) minus average unimodal di-
stribution (B). The presence or
absence of the fraction 1-20 mm
decides on uni- or bimodality.
There is no correlation either
between the position of the mo-
des and the "valleys" or bet-
ween the respective percenta-
ges, all these features vary
independently from each other.

normal distributed sand mode (second order), and a remainder
which comprises the fraction 1.6-12.5 mm, corresponding almost
perfectly to the 1-20 mm interval which makes the difference
between the average bimodal and unimodal distribution. This
third order element, crucial for modality, is believed to be
more dependent on availability than on hydrodynamics and this
problem will be resumed in the conclusion section after more
data have been presented.

Test sampling

The objective of the test-sampling campaigns was to find out
wether modality depends on the following circumstances:

* sampling technique (operator bias)
* position within the river bed (close to banks or midstream)
* downstream development (distance from the river-mouth)
* repeated sampling over several years

We took randomly chosen samples at certain distances from bank to bank and midstream at the walls of gravel bar cuts. The mean size of the channel samples is 271 kg (maximum grain-size is about 120 mm), thus guaranteeing that we are dealing with real bulk samples cutting several sedimentation units. To prove this, nearly all samples were taken as twin or double samples, that is, after having taken and treated one channel sample we continued in the same sample channel to process the second or twin sample (Fig. 3).
Taking channel samples of the walls of river cuts it is not possible to predict weather a unimodal or bimodal distribution

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Figure 3. Sampling test campaign: →→→→→→→→→→→

A: Amendolea river-mouth area: Position (●) of the randomly taken twin samples of the Amendolea [02] river-mouth area and their grain-size frequency distributions. Relatively stable sand modes contrast with relatively variable gravel modes.

B: Buonamico upstream: Position (●) of the randomly taken twin samples of the Buonamico [08] upstream area 10.6 km from the coast and their grain-size frequency distributions. Sand modes are only weakly developed or lacking, gravel modes are relatively stable.

C: Buonamico midstream: Position (●) of the randomly taken twin samples of the Buonamico [08] midstream area 3.6 km from the coast and their grain-size frequency distributions. As in the upstream section (B), sand modes are only weakly developed or missing, gravel modes are relatively stable.

D: Buonamico river-mouth area: Position (●) of the randomly taken twin samples of the Buonamico [08] river-mouth area and their grain-size frequency distributions. As in the up- and midstream section (B and C) sand modes are only weakly developed or lacking; gravel modes, however, are somewhat less stable.

E: Torbido river-mouth area: Position (●) of the randomly taken twin samples of the Torbido [14] river-mouth area and their grain-size frequency distributions. South bank samples are nearly unimodal, midstram and north bank samples are not.

F: Amusa river-mouth area: Position (●) of the randomly taken twin samples of the Amusa [15] river-mouth area and their grain-size frequency distributions. This small river displays the most homogeneous grain-size frequency distributions during this campaign. The basic granulometric structure of unimodality is accompanied by many small variations.

A

B

C

D

E

F

occurs. Consequently, an operator error can be excluded. The Amendolea [02], 152 km^2, is one of the largest rivers of the area. The source area is dominated by metamorphic rocks, and conglomerates feed recycled granitic gravels into the river-mouth sediments. The twin samples, at distances of several 100 m, taken within some days under constant conditions, show sur-prisingly uniform sand modes, less stable gravel modes and an obviously stable bimodality (Fig. 3A). From this one is tempted to speak of evidently bimodal river-mouth sediments of this ri-ver. Two other samples taken in other years (Fig. 4) disturb this picture, because one of these samples display only a very weak sand mode close to unimodality.

The Buonamico [08], 139 km^2, one of the largest and most cha-racteristic Calabrian Fiumaras with a mostly metamorphic source area was chosen to investigate a possible downstream develop-ment of modality. The three twin samples from upstream, 10 km from the coast (Fig. 3B), display nearly unimodal distributions with relatively stable gravel modes. The midstream section, 3.6 km from the coast, shows a bimodal and a nearly unimodal dis-tribution close to the banks as well as the same ambiguous si-tuation in a triple sample sequence in between (Fig. 3C). The same picture is yielded by the river-mouth sediments (Fig. 3D). Thus, in this campaign, the Buonamico appears as an producer of very slightly bimodal sediments. But only during this campaign. Additional sampling in other years (Fig. 4) in the river-mouth area presents both a strictly unimodal and, one year later, a strictly bimodal Buonamico.

The Torbido [14], 160 km^2, is the largest river under investi-gation, the source area is granite- and sediment-dominated. A river of this type is commonly expected to be a powerful produ-cer of sand. But again, the sand modes are very modest or lac-king (Fig. 3E). The modality of these river-mouth sediments de-pends on where these bulk samples are taken. Additional samp-ling in different years makes this even more evident (Fig. 4). Here, we obtained a weakly bimodal sample as during the test-sampling and a year later a strongly bimodal one.

The Amusa [15], 38 km^2, is a very small river with a predomi-nantly granitic source area, followed by metamorphic rocks. This river displays a strong unimodal tendency over all samples (Fig. 3F), one of the SW-bank excepted. The same is true for the additional samples (Fig. 4).

The results of the sampling-test campaign demonstrate that none of the criteria listed displays a significant and mono-causal influence on bi- or unimodality. Both types of distributions may occur more or less irregularly under all circumstances. This becomes even more striking if we compare the overall grain-size distributions of the river-mouth sediments of the 1972 and 1973 campaigns, before and after the disastrous Janu-ary flooding (Fig. 4). Strong differences in modality from year to year signalize that nearly all rivers are able to produce both types of distribution, uni- and bimodal. From this it be-comes evident that modality is not a given characteristic of a given river in space and time and it may appear useless to look at possible links between modality and source-area conditions

Figure 4. Grain-size
frequency distributions
of the 19 river-mouth
sediments, 1-19: 1972
and 1973 campaign, the
Palizzi [04] was only
sampled once. 20: aver-
age 72 and 73. The Ca-
reri [09] is the most
striking example for
changing modality con-
ditions within one and
the same river.
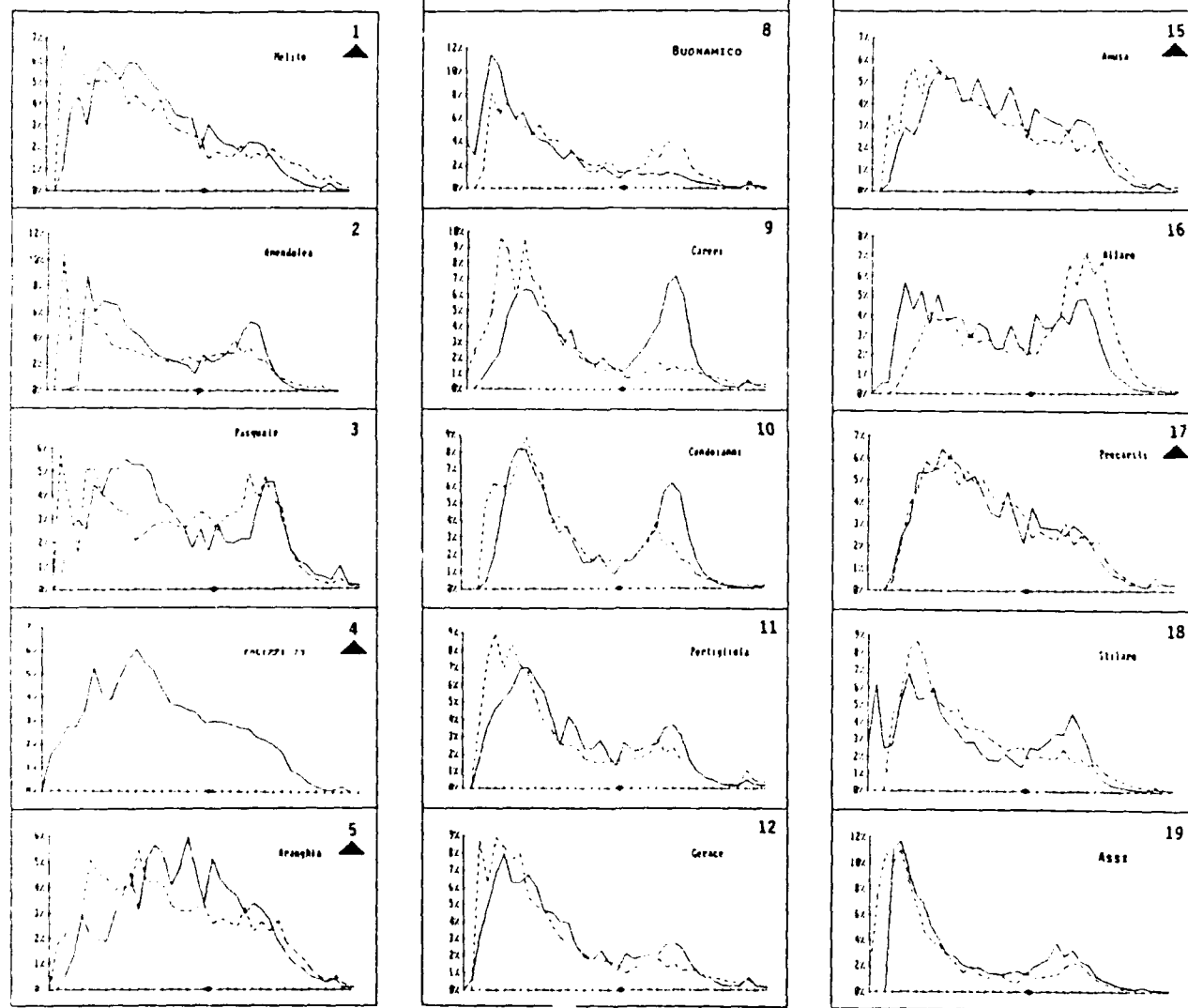
Abscissa in 1/3 phi steps, dots indicate the
1.6-2 mm interval. Triangles (▲) refer to the
predominantly unimodal distributions which are
used for possible source area correlations
(Fig. 5).
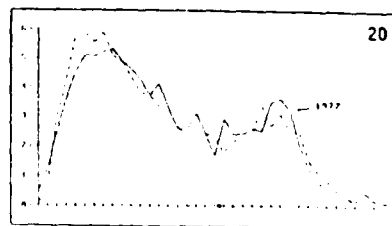
or other properties which have to be assumed as constant in space and time. However, the phenomenon of modality exists as such and our understanding of it may be improved by narrowing down possible solutions at the expense of falsificated ones.

Modality and the source-area variables

Looking for explanations for bi- or unimodal river sediments we will ignore the possible longterm changes in modality of a given river and trust the results of the two-year campaigns plotted in Figure 4. Here the river-mouth sediments of 7 rivers display at least a certain unimodal tendency and an acceptable similarity between the two campaigns:

Melito	[01]	Novito	[13]
Palizzi	[04]	Amusa	[15]
Aranghia	[05]	Precariti	[17]
Bruzzano	[06]		

The Palizzi, however, was only sampled once. Suppose these 7 rivers produce predominantly unimodal sediments and the other 12 rivers favour bimodal ones, we are now able, in the case of Calabria, to check most arguments for or against bimodality cited in the literature. These are geomorphology, petrography and primary grain-size distributions of the source areas, the petrography of the river-mouth sediments itself, packing and the refill of interstices and, only indirectly interpretable, runoff- and hydrodynamic conditions.

Source area morphology and modality

If runoff- and hydrodynamic conditions characteristic of high-gradient mountain rivers were be responsible for the uni- or bimodality of fluvial sediments, and if these runoff- and hydrodynamic conditions were be relatable to morphological features of the source area such as relief, dissection, basin shape or the gradient of the trunk slope - then these features, in turn, should correlate with the modality of the river-mouth sediments. Strahler (1964, p.4-45), for example, expects low but extended peak flows from elongated basins and sharp peaks from rotund basins. High relief rivers should create river-mouth sediments different from low relief rivers. Furthermore, current velocities of a river with a steep trunk slope should be higher than those of rivers with lower gradients. However, the Calabrian data do not reveal the slightest relationship between the morphology of the source area and the modality of river-mouth sediments. Figure 5B shows the relief of the 19 basins in terms of the altitude standard deviation, Figure 5C presents the hypsometric integral or dissection in terms of the altitude skewness and Figure 5D displays the basin shapes according to Horton's formula (Horton 1932): On all plots the rivers with predominantly unimodal river-mouth sediments are marked. There are certain tendencies, but they are never exclusive of one modality alone. For example, all unimodal river-mouth sediments belong to elongated basins characterized by low Horton ratios. But Laverde [07], Portigliola [11] and Assi [19],
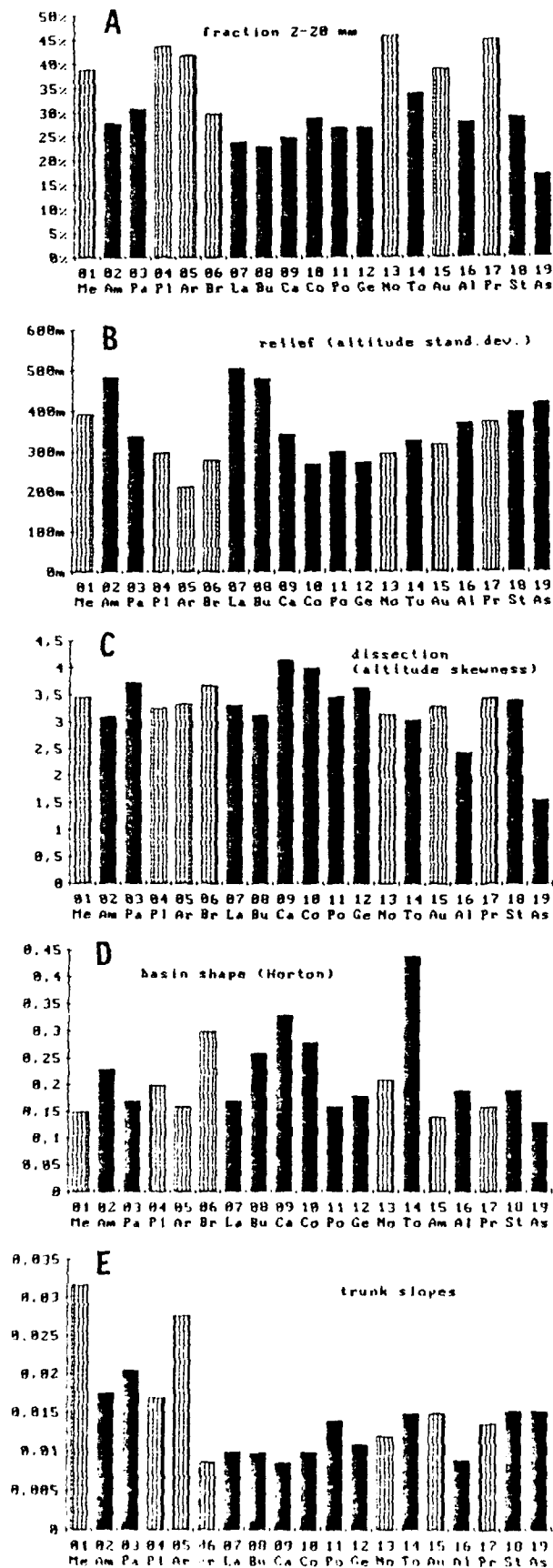
Figure 5. Uni- and bi-modality of the river-mouth sediments of the 19 basins versus sediment- and source-area variables. River-mouth sediments with predominantly unimodal distributions are hatched, bimodal ones are marked black.

A: Percentage of the fraction 2-20 mm. Rivers with more than about 35 % of this fraction are unimodal, the Bruzzano [06] excepted (Fig. 4/6). So this fraction is crucial for modality.

B: Relief (altitude standard deviation) does not correlate with modality.

C: Dissection (altitude skewness) does not correlate with modality.

D: Basin shape R_f (Horton 1932) does not correlate with modality.

E: Slope of the trunk (river segment of the highest order) does not correlate with modality.

with partly even lower Horton ratios, produce predominantly bi-modal river-mouth sediments. Figure 5E shows the trunk slopes of the 19 rivers, the gradients of the main channel of highest order: These gradients vary noticeably between about 0.03 and 0.01, this is, by factor 3, but unimodality is combined with both lowest and highest gradients.

Source area petrography and modality

Source rock petrography is probably the most popular argument to account for bimodality. The missing grain sizes - mostly granules - are said to be less frequently produced by certain rock types. There is no support for this in Calabria.

The area percentage and the erosional budget of the 19 basins in terms of granitic, sedimentary- and metamorphic rocks demon-strate some combinations of unimodality and granite-poor source areas, but not exclusively. Five "bimodal rivers" fall into this category, too. The test was repeated using various combi-nations of metamorphic- and sedimentary rocks, but no preferred correlation occurred. Finally a combination of granite, sand-stone and conglomerate of the source areas versus the modality was plotted in order to find out whether these predominantly sand producing units correlate with the modality of the river-mouth sediments - again without success. In Calabria the moda-lity of river-mouth sediments is definitely not controlled by source rock petrography, as supposed and proposed in other re-gions by Hough (1942), Yatsu (1955) or Rogers et al. (1963).

Source area grain-size distributions and modality

Wolcott (1988) proposed the size distribution of the input ma-terial as being largely responsible for the size distribution of channel deposits. Apart from the fact that it is hard to ex-plain how such a hydrodynamically delicate structure like bimo-dality is maintained during fluvial transport over several ki-lometers - in the case of Calabria it hase been known for years that most jointed and weathered source rocks are extremely uni-modal and most river-mouth sediments are strikingly bimodal. So the bimodality of river-mouth sediments in Calabria is defini-tely not controlled directly by the granulometric properties of the source rocks.

Petrography of river-mouth sediments and modality

Very little attention has been paid to the petrography of the sediments itself as a possible reason for modality (Davies et al., 1978). However, in Calabria this question leads into a blind alley, too. The petrographic composition of the gravel fractions 16-80 mm and 1-20 mm of the 19 river-mouths was de-termined. There is definitely no correlation between the cha-racter of modality and any petrographic composition of the two gravel fractions. This is especially important in the case of the range of 1-20 mm, the third order element, because this fraction constitutes the main part of the mysterious "gap"

which defines modality. The same is true for the petrographic
composition of the sand mode. The light mineral indices of all
19 river-mouth sediments are analysed. The content of rock
fragments or the quartz/feldspar ratio may account for the
breakdown of certain fractions and may be responsible, in this
way, for modality - but there is not the slightest correlation
between these or other compositional parameters and modality.
So even the petrographic composition of the river-mouth sedi-
ments itself can definitely be excluded as a possible factor
controlling modality - in the case of Calabria.

Packing and refill of interstices and modality

In 1935 an important statement concerning our topic was made by
Fraser, barely cited thereafter (Plumley 1948, p.542, Pettijohn
1957, p.45). Fraser (1935, p.987) says: "... it is difficult to
understand how a river capable of carrying all sizes up to
large boulders can deposit them at the same time. At any in-
stant, the largest particle that may be moved varies at the
sixth power of the current velocity". Fraser calculates that
the velocity of a stream capable of carrying 250 mm pebbles has
to decrease 60 % before 1 mm sand could be deposited. "Hence it
seems improbable that large pebbles and fine sand are necessa-
rily deposited simultaneously", and "The periods during which
river gravels may be moved, sized and sorted are usually brief
and are followed by much longer quiescent periods during which
infiltration and arrangement of the finer sizes may be accom-
plished" (op.cit., p.984). Fraser calculates the effects of va-
rious gravel grain-sizes enabling finer fractions to enter the
framework after deposition and defines a "critical ratio of
entrance" (op.cit., p.919). Plumley (1948, p. 544) agrees with
Fraser and finds, by theoretical considerations, ratios between
22 % and 32 % for the sand content of gravels. Plumley finds
this confirmed by his samples and states that the weight ratios
of the infilling fractions increase with decreasing mean size
of the sediment.
As pointed out, this and all other features of the grain-size
distributions were checked, but no correlation was found: the
grain-size classes of both the gravel and the sand modes, the
"valley" in between, the sand content and finally the percen-
tage of the third order element, the fraction 1-20 mm - all
vary independently among themselves.
Only the mean sand content of 25 % corresponds nicely to the
sand contents claimed for filling interstices, but this, unfor-
tunately, is irrelevant for uni- or bimodality.

Hydrodynamic conditions and modality

Many authors believe that selective transportation due to dif-
ferent hydrodynamic behavior of different grain sizes is a main
factor controlling modality. These arguments are largely refu-
ted by Shea (1974, p.1000). The most cited authors in this con-
text are Sundborg (1956) and especially Russel (1968), who ar-
gues "that the fractions 1-6 mm are more readily entrained and
more rapidly transported than grains of larger or smaller size.
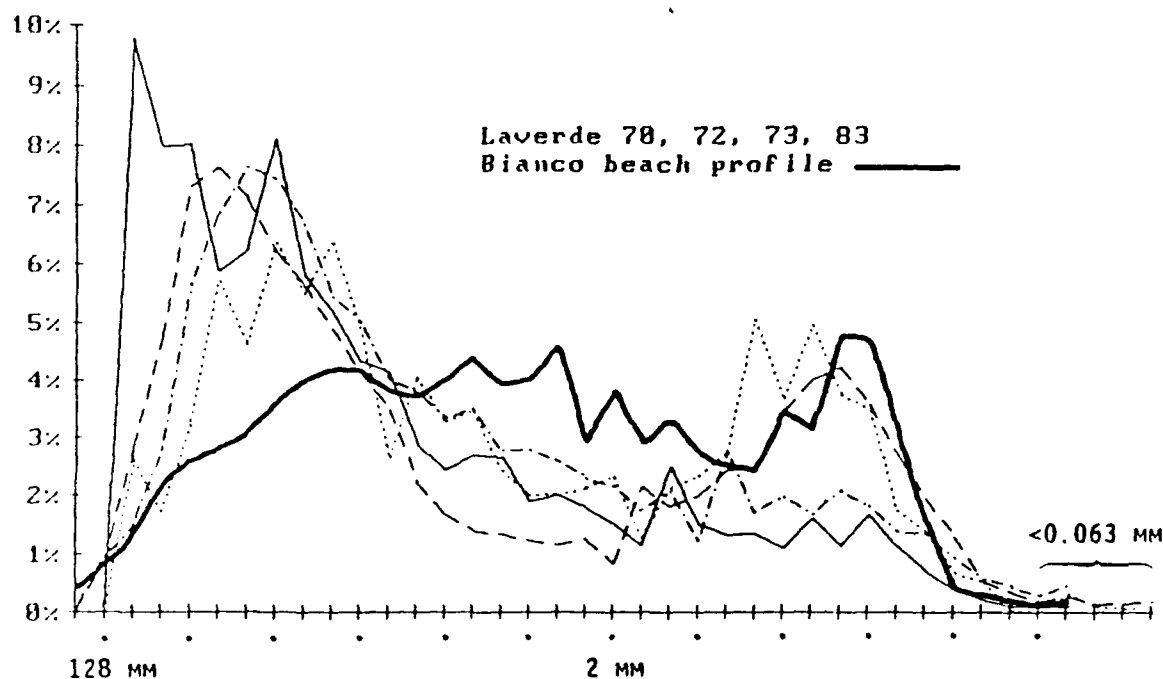Currents, mainly of wave origin, concentrate these grains in

Figure 6. Comparison of fluvial and coastal grain-size distri-
butions, Laverde [07] and Bianco Beach. The Laverde is the main
feeder of the Bianco Beach. The four Laverde samples stem from
the 1970, 1972, 1973 and 1983 campaigns. The heavy line refers
to the "Strand Profil", a continuously sampled beach-normal
profile 70 m long, from 2 m above to 4 m below sea level, com-
prising a total of 7 t. The profile tries to bridge the entire
beach in motion.

beaches, where they are present in more-than-ordinary abun-
dance" (op.cit.p.31). The first problem which arises with re-
gard to Russel's idea is that the Calabrian grain-size gap is
much larger than from 1-6 mm. As pointed out by Shea (1974,
p.1000) no hydrodynamic mechanism is known which favours conti-
nuously the entrainment and transportation of grain-sizes bet-
ween 1 and 20 mm. It was not possible to check the hydrodynamic
qualities of this argument, but the bulk grain-size distribu-
tion of the beach was calculated to check the consequences
claimed by Russel. This sample (Fig. 6) represents the average
of 32 excavator samples, 7 tons as a whole, of a 70 m profile
of the entire beach in motion, normal to the coast from the
backshore 2 m above sea level to the offshore strip 4 m below
sea level. From the Laverde river-mouth [07], which feeds this
beach there are 4 samples, from 1970, 72, 73 and 83. The single
fluvial distributions vary between nearly unimodal and strongly
bimodal. The coastal distribution admittedly fills the fluvial
gap (Fig. 6) if referred to the bimodal distributions. However,
the coastal sand mode slightly exceeds the gravel mode, and
there is a deficiency, too, somewhat shifted towards the fines.
Thus from the merely descriptive point of view Russel is not

wrong, but the question remains whether the fill of the gap in the coastal realm stems from increased fluvial input or simply from attrition of the coarse mode during longshore transportation.

As a result, none of the reasons assumed to be responsible for uni- or bimodality can be accepted in Calabria. The interpretation given in this paper focuses attention on attrition and selective transport or deposition as already supposed by Shea (1974, p.1001).

Conclusion

To return to the synoptical comparison of the river-mouth sediment's grain-size distributions of the 1972 and 1973 campaign (Fig. 3, 01-20). The basic or first order feature present in all samples is the positively skewed gravel mode. This Rosin distributed gravel mode is interpreted as an inherited, primary source rock element (Ibbeken 1983, p.1227), which is maintained during fluvial attrition and selection until the river-mouth areas are reached. It goes without saying that during this development the gravel mode is modified shifting towards the fines and flattening. Thus the gravel mode carries both a compositional and a textural provenance signal.

The other important, second order feature is the sand mode of the bimodal distributions, usually lognormally or Gauss-distributed. This mode may be very pronounced (Careri [09] and Condoianni [10]) or nearly absent (Melito [01], Palizzi [04] and Precariti [17]). The geochemical evidence makes it a reasonable assumption that the overwhelming portion of fluvial sand does not come from the jointed and weathered source rocks or soils, nor from disintegration or decomposition (Pettijohn et al. 1987, p.252) but from attrition mainly during the floods of the high-gradient mountain rivers. This is the only way to create sand whose geochemical composition is fresher and more original than that of the debris of slopes and soils. It is not clear why the sand modes of the bimodal distributions almost exclusively display lognormal instead of Rosin distributions, as expected from the process of sand formation by attrition. Lognormality is usually seen as a transport-specific feature. Obviously the movement of sand, much more frequent than that of gravels, is sufficient to cause a Rosin/Gauss transformation. Thus the sand mode carries only a compositional or petrographic provenance signal, but not a textural one.
Bimodality, in Calabria, may be simply explained as the result of mixing between these different modes, gravel and sand. Ibbeken (1983, Fig. 10) mixed an ideal, coarse grained Rosin-distribution and an ideal fine grained lognormal distribution in accordance with the grain-sizes and parameters of the average Calabrian river-mouth sediment, 70 % and 30 %, and obtained a bimodal distribution which corresponds nearly perfectly to that of the mean bimodal river-mouth sediment in Calabria. It may be reasonable to infer that the bimodal Calabrian river deposits originate from gravel modes with a long fine tail, superimposed by a sand mode.

This picture is seriously disturbed by the presence of unimodal distributions. These are characterized by amounts of more than about 35 % of the fraction 1-20 mm, the third order element. The crucial question is what controls the appearence or non-appearence of this grain-size interval. All source area variables and all variables of the sediment itself, as petrographic composition or grain-size parameters were falsificated and excluded. So the only remaining variables not directly tested are nydrodynamics and availability. Obviously hydrodynamics can be excluded, too. As pointed out, there is no reason why the fraction 1-20 mm should somehow be selected by hydrodynamics. On the other hand, the complex process of attrition during bankfull stages of mountainous rivers is not accessible as yet. Consequently the following can only be a matter of belief: The process of crushing during floods differs, in space and time, between bed load, transitional stages and suspended load. The suspended load, at any given time, is largely protected against further crushing, the bed load itself is not. Thus floods of varying intensity may produce varying grain-sizes, and the third order element, the fraction 1-20 mm, may be especially susceptible for these differences, being destroyed or not, so that bi- or unimodal lag sediments result. It is necessary to speculate on this topic alone, availability via attrition, in order to explain the confusing variability of modality in space and time: all other variables cited can definitely be ruled out, in the case of Calabria.

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COUNTING INSTEAD OF WEIGHING. GRAIN-SIZE COMPOSITION OF COARSE MATERIAL BEDLOAD IN A MOUNTAIN STREAM EXPRESSED IN PARTICLE NUMBER TRANSPORT RATES

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INTRODUCTION

Coarse material bedload transport processes are not fully understood. In particular the interpretation of fluctuations in bedload transport needs clarification. The processes on the river bottom require closer attention. However, direct observations and measurements of processes on the river bottom during transport events under natural conditions are difficult.

Statistical methods should be developed that yield more information about processes during bedload transport events based on novel analyses of coarse material bedload. At Squaw Creek improved methods of bedload sampling and a special data analysis were developed

HYDROLOGICAL AND SEDIMENTOLOGICAL CONDITIONS AT SQUAW CREEK

The Squaw Creek is a mountain river draining parts of the Gallatin Range in Montana. USA. Coarse bedload is exclusively transported during spring snow-melt. The average discharge during spring high-flow is 4 - 6 m^3/s with diurnal fluctuations of flow. Conditions at Squaw Creek are described in several papers (ERGENZINGER & CUSTER. 1983: BUNTE et al.. 1987; CUSTER et al.. 1987: BUGOSH & CUSTER. 1989). The coarse material on the river bottom (> 11.2 mm) has a D_{50} of 150 mm and is coarser than the average bedload.

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METHODS OF BEDLOAD TRANSPORT MEASUREMENTS. BEDLOAD SAMPLING AND DATA ANALYSES

Magnetic tracer technique

Bedload transport was measured continuously at Squaw Creek using the magnetic tracer technique. (This methods makes use of the naturally magnetic pebbles and cobbles of the river bed. When a pebble crosses a detector log expanding over the whole river width a voltage peak is induced according to the Faraday inductive principle. The voltage peaks are amplified and recorded.) This real-time record of bedload transport shows various patterns of a fluctuating wave-like transport. During the daily rising flow bedload transport increases fluctuating only moderately, whereas immediately after peak flow the transport decreases to a minimum and continues with several large waves of transport superimposed by smaller bedload waves during the falling limb of daily flow. In between bedload waves transport often occurs randomly. The relation of voltage peak rates (BUNTE et al., 1987) or bedload transport rates with discharge shows a variation of more than one order of magnitude (BUNTE. 1990) A further step of analysis was to determine the grain-size distribution of these bedload transport waves in order to investigate the processes on the river bottom which might be responsible for the bedload fluctuations.

Large frame-net bedload sampler

A large sampler is desirable because small devices exclude large heavy clasts which are detected by the real-time bedload transport magnetic tracer technique. During the snow-melt high-flow in spring of 1988 a large frame-net bedload sampler was used for bedload sampling. The frame has an opening of 1.6 m x 0.3 m and did not truncate the coarse tail of the bedload grain-size distribution. but the mesh width of the net allowed all particles smaller than 10 mm to pass through. Compared to standard Helley - Smith bedload samplers the large frame-net bedload sampler collects all the large particles representatively because the opening is much wider. Due to its construction the large frame bedload sampler can be operated in fast flows with flow velocities of 2 m/s in which a 6 x 6 inch hand - held Helley - Smith sampler is unmanageable. However. it takes three persons to maneuver the large frame bedload sampler and requires some

knowledge about the river's transport rates to determine a satisfactory sampling duration. Samples should be as large as possible without overloading the sampler. The sampling device is described in BUNTE (1990).

Grain-size distributions of coarse river sediments and their standard grain-size analysis

A standard grain-size analysis is based on the weight of the particles in the various grain-size classes. Plotting the weight or the weight percentages of each grain-size fraction in consecutive order yields in its simplest form a unimodal distribution curve resembling a Gaussian distribution. The statistical methods available to analyze these distributions (descriptive measures such as the mode, mean, variance and standard deviation) are only suitable if the distribution is a Gaussian distribution or at least approximates such a distribution. However, many grain-size distributions of non-cohesive river sediments are not Gaussian. Focussing on coarse material bedload samples in this paper, their grain-size distributions deviate from Gaussian distributions by being truncated on their tails and by showing internal irregular weight percentages in consecutive grain - size classes.

Truncation

The truncation on the lower end of a coarse material bedload sample is a sampling artifact owing to the smallest sieve used. The truncation of the upper end of a coarse material grain-size distribution can be a sampling artifact due to small samplers, the great mass needed to collect a statistically representative sample or by natural processes related to an insufficient sediment availability or an incompetent stream power. Facing these natural and practical restrictions, the coarse part of bedload is rarely sampled in a statistical representative way unless the sampling time is extended over a long time. Long sampling periods cause unmanageably large samples and prevent an investigation of short-term temporal variation of bedload transport rates.

Irregular grain-size distribution

Short-term coarse material bedload samples are often irregular distributed. Individual grain-size classes are over- or underrepresented and their weight or weight percentage deviates considerably from a smooth Gaussian distribution. This abundance or dearth of particles in individual grain-size classes. however. should be

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taken into account in the data analysis because these irregularities might yield some valuable information on the processes on the river bed.

Deficiencies of standard grain-size analyses

The results of a standard grain-size distribution analysis following FOLK & WARD (1957) and the "moment" method used by SCHLEYER (1967) is described in BUNTE (1990). Applying a FOLK & WARD analysis to truncated grain-size distributions is problematic because any extrapolation to obtain a value for the distal percentiles is an interpretation of the "missing" part of the grain-size distribution. The "moment" method is slightly better suited for the analysis of truncated coarse material bedload samples as it does not depend on distal, extrapolated percentiles. Nevertheless, neither the FOLK & WARD analysis nor the descriptive grain-size parameters obtained by the "moment" method (the "moments" similar to the parameters such as mean, sorting, skewness and kurtosis) are suitable to analyze any irregularities in the sample distribution. By solely describing a sample distribution as a whole, these parameters do not allow an analysis of the number of particles in the individual grain-size classes. Nevertheless, the abundance or dearth of particles in an individual grain-size class in a bedload sample might give insights into processes on the river bottom. Trying to study river processes this valuable information should not be omitted by using an unsuitable method of grain-size distribution analysis.

Grain-size distribution analysis: particle number transport rates

Although particles interact in all size levels, in sand individual particle interaction is difficult to study. In previous studies on (river) sediments, especially on sand material, the sediment size distribution was analyzed by the particles' weight percentage because sand grains are very small and the number of grains is uncountable. It had to be assumed that the weight percentage was related to the percentage of the number of particles in a grain-size class. As particle size increases, individual particles can more easily be seen. In future studies it might prove to be a better approach to analyze a grain - size distribution by the particle number. Since individual particles are interacting (hiding, exposure, ...) on the river bed during a bedload transport event individual particle interaction is important to be understood. Using the number frequency of a grain-size class leads itself to a statistical analysis.

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For coarse grained sediments with countable specimens present it seems to be a better way to quantify the particles by their number per sieve class than using the particles weight in a sieve class. The bedload samples collected at Squaw Creek did not reach the sample weights which IBBEKEN (1974) and CHURCH et al. (1987) recommended for a statistical grain-size analysis. However, the counting method for grain-size analysis can be effectively applied to truncated or irregular grain-size distribution which are common for short-term coarse material bedload samples in order to analyze particle dynamics.

The grain-size analysis depicted here examines the particle transport rate for each grain-size class present. Sieving was done in 0.5 phi units for particles in the range of 11.2 mm to 180 mm. The particle transport rate for a given grain-size class was calculated by counting the number of particles in each grain-size class per meter width and minute. A time unit of minutes was chosen to reduce the number of decimals to make the data more legible. Counting the particles in consecutive sieve classes gives the advantage of obtaining a grain-size distribution that can be described by a simple monotonic function (linear, exponential or power function) if the number of particles transported in consecutively increasing grain-size classes decreases regularly. The particle transport rate is plotted for all samples (s.fig.1)

Further calculations with these simple monotonic functions are more sensible and graphic than calculations with Gaussian style functions that are equivocal. As the number of particles transported in consecutive, increasing grain-size classes decreases as a monotonic function, the regression lines of particle transport rates of given grain-size classes versus discharge (s.fig.2) or versus total bedload transport rates (s.fig.4) plot in consecutive order. Plotting weight percentages of given grain-size classes versus discharge or total bedload transport rates would yield regression lines that one by one moved upward on a plot until the grain-size class with the largest weight percentage is reached. For the following grain-size classes with decreasing weight percentages the regression lines would move down on that plot, leaving a bundle of lines that is difficult to interpret.

DATA ANALYSIS

Comparison of the three sedimentary units grave bar, channel surface and bedload

The first step of analysis compares the gravel bar, channel surface and bedload distribution to assess of how these sedimentary units relate to each other. In a bulk sample from the gravel bar the number of particles per grain-size class decreases very regularly (exponentially) with an increase in grain-size. The channel surface sample shows the same regular decrease in the amount of particles for the gravel and pebble sizes but an abundance of cobbles (SUTHERLAND, 1987) which stabilize the river bed as a pavement. The number of cobbles is roughly the same for each size class in the cobble section and decreases only towards the very large cobbles. The bedload samples show a less regular decrease in the number of particles for consecutive grain-size classes. Compared to the grain-size distribution of the gravel bar the bedload samples consist of less gravels and more pebbles suggesting that pebbles are transported preferentially. This patterns can not be found in samples taken at very small rates of bedload transport which have a random grain-size distribution (samples #20 #21 (not shown in fig.1 but similar to sample # 22), #22, #24, #25.).

Comparison of consecutively taken samples

Two series of samples were taken consecutively in 1 - 2 hour intervals at normal high-flow discharges between 5 and 6 m³/s. Focussing on these serial samples of bedload provides an opportunity to investigate the particle size class responsible for rapidly fluctuating transport rates (bedload waves).

Particle number transport rates versus discharge

The particle transport rates (particles / m - min) of all bedload samples were plotted versus discharge (fig.2). The data show that for a given discharge the number of particles transported per minute in each grain-size class can vary by two orders of magnitude. To better analyze these data, exponential functions were fitted through the particle transport rates of each particle size class with discharge (s. last plot of fig.2). The correlation coefficient r^2 is generally low and still best for the smallest grain-size

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class (11.2 - 16 mm) (fig.3).

The regression lines group into two sets of lines: lines 1 - 4 representing the grain-size classes 1 - 4 (11.2 - 45 mm) and lines 5 - 7 representing the grain-size classes 5 - 7 (64 - 125 mm). With the exception of line1, both sets of regressions are each parallel to each other. Particles smaller 45 mm (line 1 - 4) have a common threshold of motion at around 3.5 m³/s and their transport rates increase with discharge exponentially. (As in log - linear plot there is no zero line the threshold for particle motion is set at a transport rate of 0.01 particles per meter per minute which equals a transport rate of 1 particle per meter width per 100 minutes). Particles larger than 45 mm do not start to move unless the discharge exceeds 4.5 m³/s and their contribution to the sample increases at a faster rate as discharges increase than the contribion of smaller particles. At higher discharge rates of over 5.5 m³/s the differences in the particle transport rates of individual size classes become less distinct which could suggest the possibility of an equal mobility transport at high discharge rates. More data might also bring the regression lines into parallism, but this possibility can not be definitely confirmed or denyed.

In order to examine the causes of the scatter in the data the particle transport rates of consecutive samples were marked. The first serial sample (sample # 5 - 9 marked by large dots in fig. 2) was taken at relatively low discharges on the rising limb of one night's high flow with discharges ranging from 5.0 to 5.26 m³/s and shows a basically regular increase of particle transport rates for the various grain-size classes up to the pebble range (64 mm). The second set of serial bedload samples (sample # 10 - 16) was taken at higher discharges (5.58 to 5.73 m³/s) during the next night's high-flow from waxing to waning flow. On the falling limb of the hydrograph the total bedload transport rates fluctuate considerably (BUNTE, 1990). Marked by squares and triangles to distinguish between large and small total bedload transport rates this second set of serial samples shows much variation in the transport rates of the small and medium sized particles (gravels and pebbles). To further investigate this variation the particle transport rates for all grain-size classes were plotted versus the total bedload transport rate of each sample (fig. 4).

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<u>Particle number transport rates versus total bedload transport rates</u>

In each grain-size class the number of particles transported per meter and minute increases very regularly with the total bedload transport rate in grams per meter and second. The power functions that were fitted through the data (s. last plot of fig. 4) yield high correlation coefficients (fig. 3) which are best for the particle number transport rates in the range of pebbles (22.4 - 64 mm). This shows that the number of pebbles increases rather regularly with the total amount of bedload while smaller as well as bigger particles are variable in their contribution to a sample's total bedload rate.

The parallelity of the regression functions suggest that with an increasing total bedload transport rate the number of particles transported in each size class increases regularly. The regression lines group into two sets again. Lines 1 - 3 (or 4) representing the namely grain-size classes are closer to each other than the regression lines 5 - 7. This indicates that the transport rates for smaller particles are similar to each other while the transport rates of bigger particles are distinctly different in each size class. Considering a particle number transport rate of 0.01 particles per meter per minute as a threshold for particle motion a total bedload transport rate of 2 grams per meter per second has to be exceeded before the a cobble of 90 - 125 mm can be moved.

For total bedload transport rates larger than 2 grams per meter per second the number of particles per size class increases regularly. This result could confirm the ANDREW'S & PARKER (1987) theory on equal mobility of all particles present for larger transport rates.

Closer examination of the two serial samples described above shows a regular tendency for all of the first set of serial samples (large dots) that were taken during rising flow: the number of particles in consecutive grain-size classes increase regularly with the total bedload transport rate. Conversely, the second set of serial samples taken from rising to waning flow can be clearly divided into samples with large total bedload transport rates ranging from 42 - 72 g/m·s (sample # 11,12,14 and 16 (see the squares)) and small total bedload transport rates of 3 -12 g/m·s (sample # 10. 13 and 15 (note the triangles)). While at large bedload transport events (squares) the particle number transport rate in each size class follows the general trend in the plots the particle number transport rates cause some scatter in the plot during smaller

transport rates (triangles). The deviation from the regression lines is strongest for small gravel particles which are either over- or underrepresented and for cobbles which are mostly absent. However, the medium particle range (pebbles) adjusts best to all bedload transport rates.

DISCUSSION

Coarse material bedload transport rates are poorly related to discharge (BUNTE et al.1987 and BUNTE.1990). Bedload transport rates often fluctuate and their fluctuation has been attributed to a number of processes by various authors (GOMEZ. NAFF & HUBBLE 1989).

Investigating the grain-size composition of bedload samples provides insight into the processes on the river bottom. The method suggested in this paper offers a relatively easy tool to analyze the grain-size distribution of coarse material bedload samples.

Various patterns of bedload transp t have been detected at Squaw Creek during the daily high-flows (BUNTE et al., 1987). Statistical representative bedload samples were taken during different transport patterns and their grain-size distribution was analyzed. Different transport patterns revealed different grain-size distributions.

A series of bedload samples collected during rising flow with only moderately fluctuating transport rates show grain-size distribution in which the number of particles of consecutive grain-size classes decreases rather regularly with discharge and with total bedload transport rate, suggesting that bedload transport in this transport phase is mainly hydraulically controlled.

During the passage of a crest of a bedload pulse the grain-size distribution also follows regular pattern. The number of particles per each grain-size class decreases regularly with increasing grain-size class. All or almost all the grain-sizes present are transported. In this phase of transport the ANDREWS & PARKER (1987) theory on equal mobility of all grain-size classes present on the river bottom seems to be applicable.

During the passage of a trough of a bedload transport wave the grain-size distribution

of the bedload deviates from the regular patterns described above. These small bedload transport rates do still consist of an appropriate number of pebbles compared to the small amount of bedload, but the number of gravel and cobble particles is less predictable. The gravels and cobbles seem to be subject to selective transport during phases of the passage of troughs in wavy transport. A possible reason for this selective transport is that gravels and cobbles are most strongly influenced by the effects of hiding and exposure. While cobbles are only transported after being fully exposed to flow, gravels are abundantly transported when being winnowed out and are rarely transported when trapped in between cobbles. The pebbles are evenly transported as they are too big for excessive hiding and are not easily trapped. Pebbles can be transported when the shear stress is large enough for motion during high-flow's bankfull stage. Gravels and cobbles strongly interact on the river bottom, hindering or favoring each other's transport while pebbles are effected by the roughness of the river bottom to a much lesser extent and smoothly move through.

The temporal variation of coarse material bedload transport rates at Squaw Creek can be explained by a switching of transport modes which alternate between equal mobility and selective transport owing to the changes in roughness of the river bottom and the resulting effects of hiding and exposure of particles. The equal mobility transport mode causes high bedload transport rates, the selective transport reduces bedload transport rates. An alternation between the two transport modes leads to a fluctuation in bedload transport rates.

Though the fluctuation in bedload transport rates occur at a much faster rate (hourly intervals) than variations in discharge (daily intervals) it remains to be clarified whether the temporal and spacial variation in the hydraulic conditions of a diurnally changing flow during a snow-melt run-off contribute to the rapid temporal variation of bedload transport rates.

ACKNOWLEDGEMENTS

computer and Ulf Achter who helped with some of the graphs.

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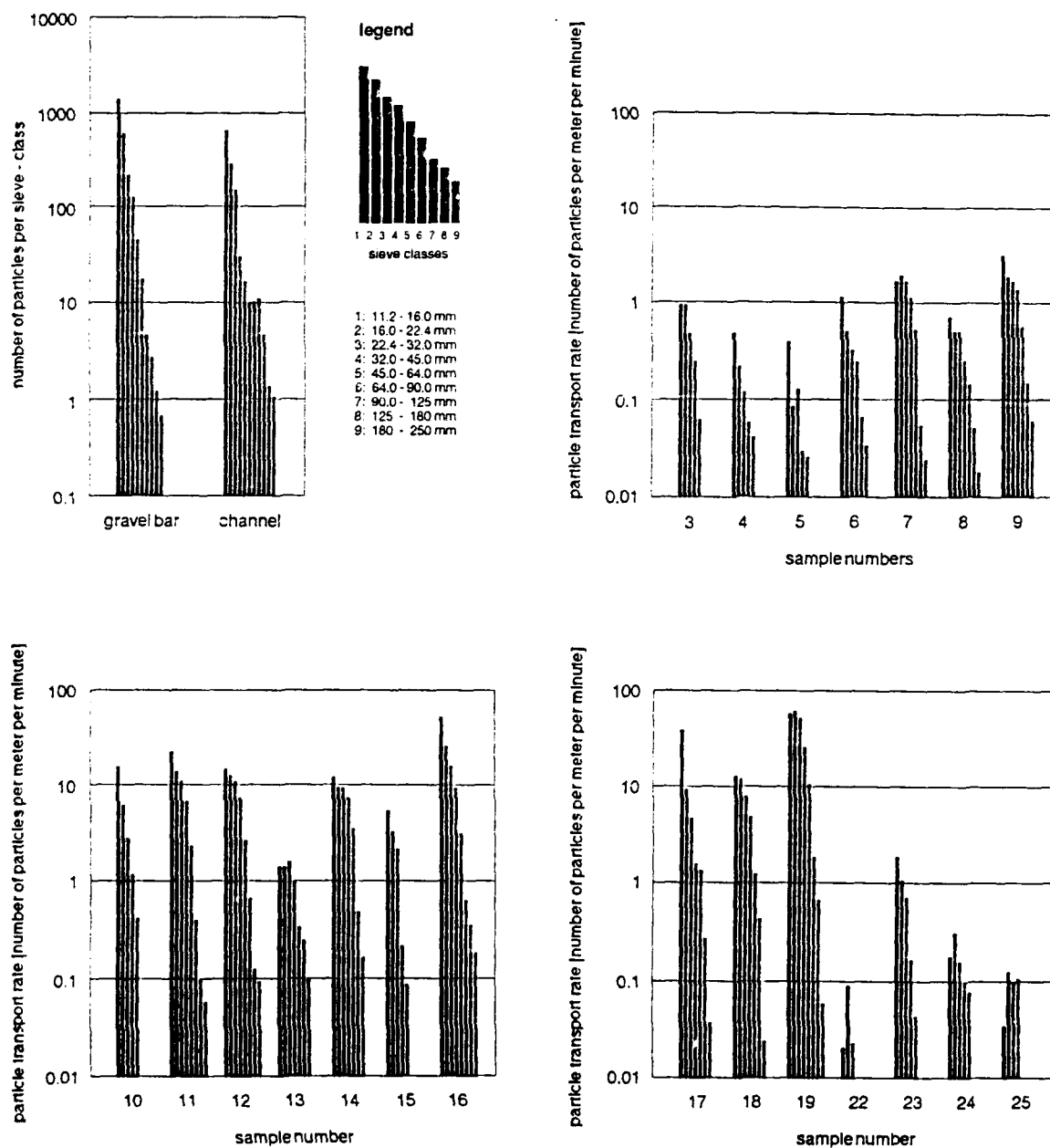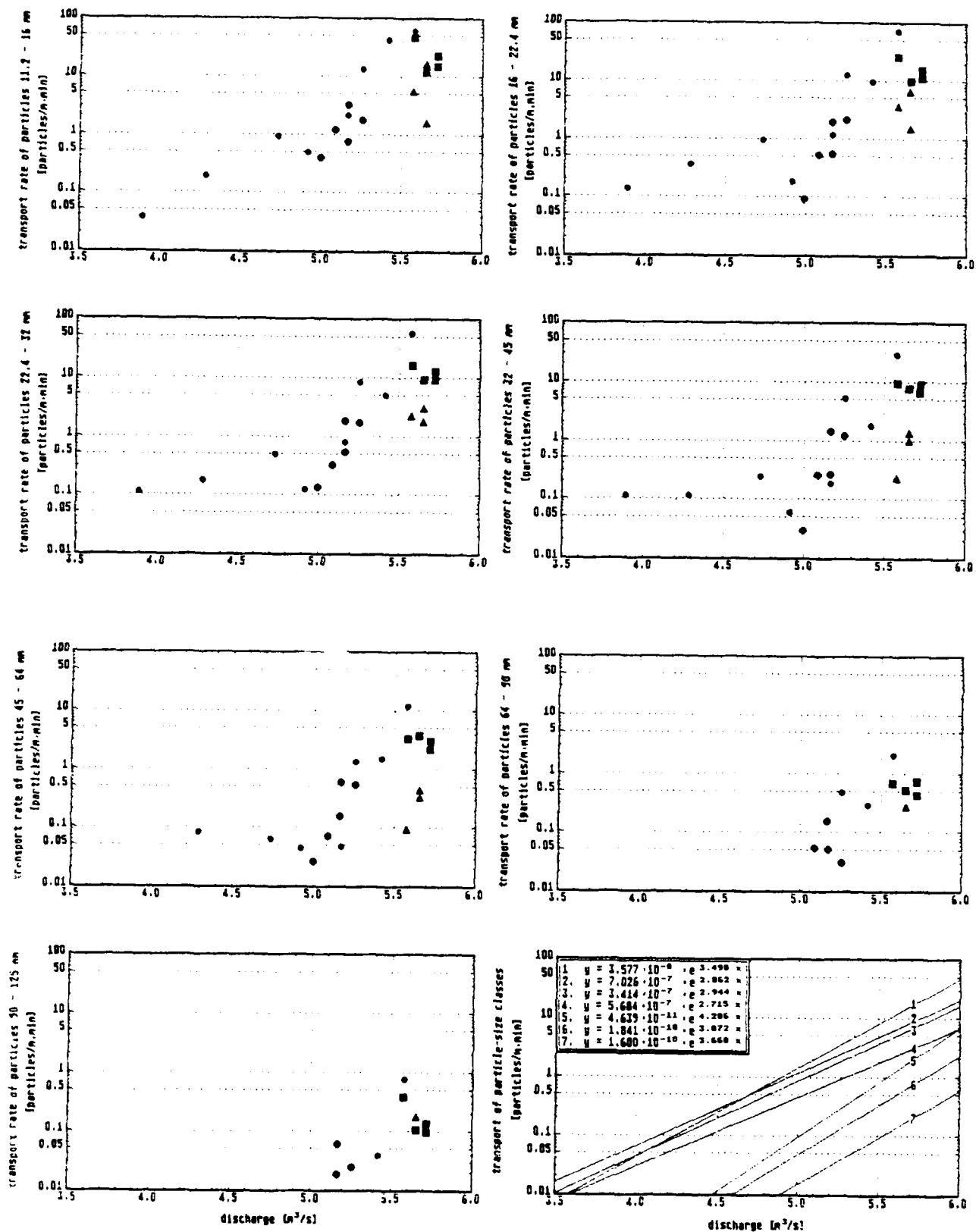
number of particles per sieve - class

10000
1000
100
10
1
0.1

gravel bar   channel

legend

1 2 3 4 5 6 7 8 9
sieve classes

1: 11.2 - 16.0 mm
2: 16.0 - 22.4 mm
3: 22.4 - 32.0 mm
4: 32.0 - 45.0 mm
5: 45.0 - 64.0 mm
6: 64.0 - 90.0 mm
7: 90.0 - 125 mm
8: 125 - 180 mm
9: 180 - 250 mm

particle transport rate [number of particles per meter per minute]

100
10
1
0.1
0.01

3   4   5   6   7   8   9

sample numbers

particle transport rate [number of particles per meter per minute]

100
10
1
0.1
0.01

10   11   12   13   14   15   16

sample number

particle transport rate [number of particles per meter per minute]

100
10
1
0.1
0.01

17   18   19   22   23   24   25

sample number

Fig. 1: Particles and particle transport rates: number of particles in consecutive grain-size class per meter per minute.
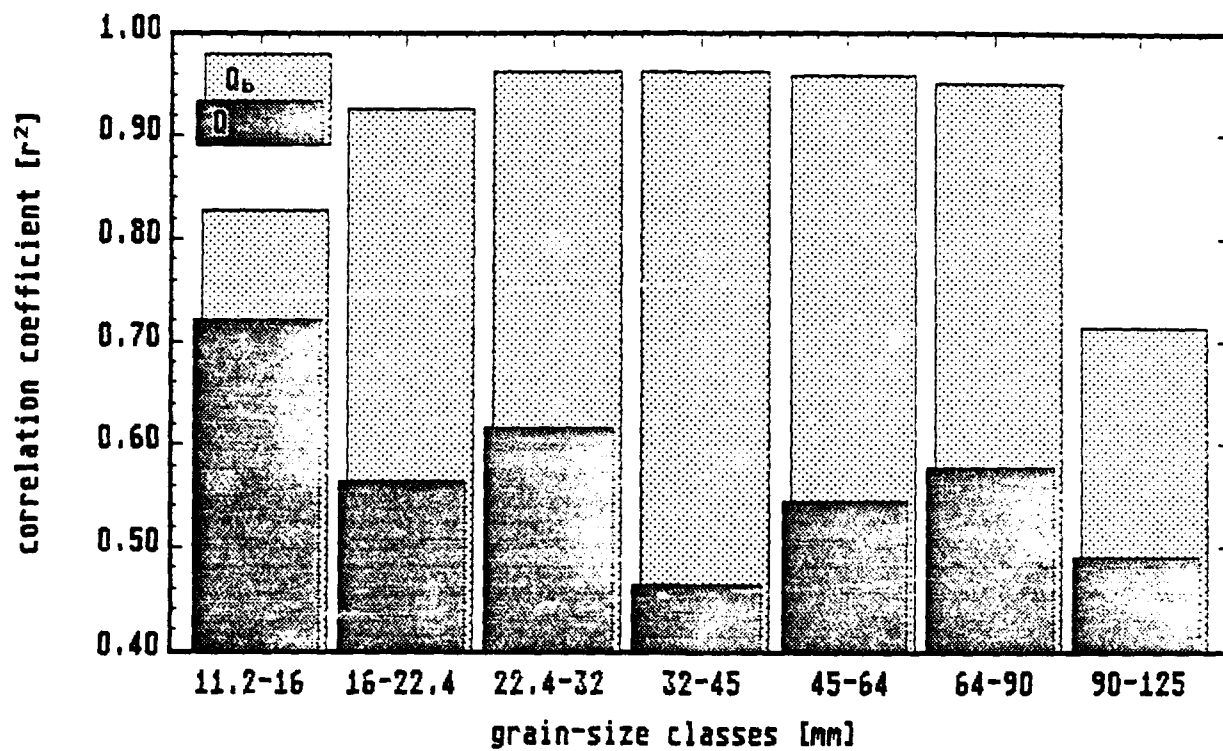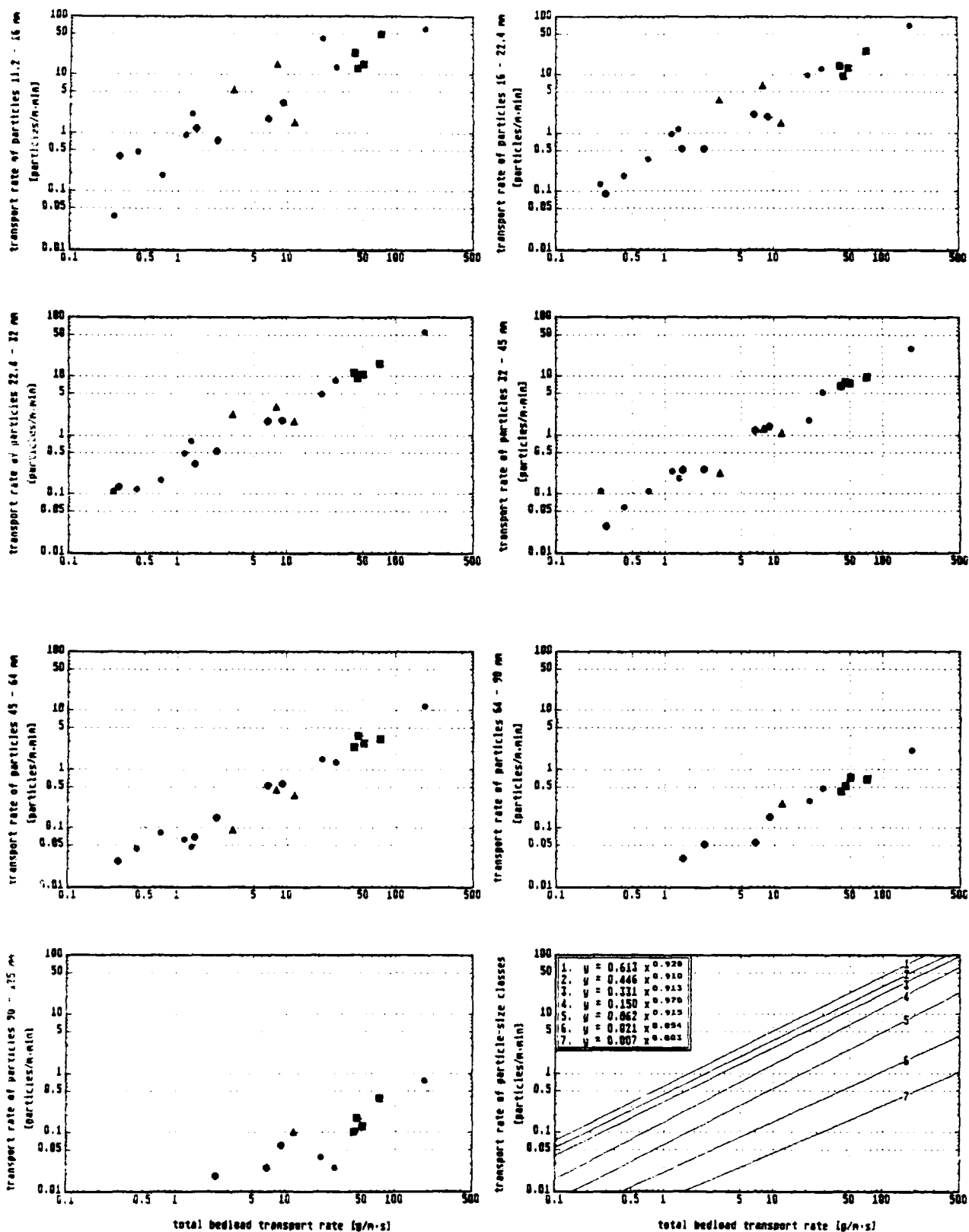
Fig 2: Particle transport rates versus discharge.

Fig 3: Correlation coefficients of regressions of particle transport rates versus discharge and versus total bedload transport rates.

» May 23 - 24, 1988,   0 (5.8 - 5.3 m³/s),   q (0.3 - 5.3 g/m·s)
▲ May 24 - 25, 1988,   0 (5.2 - 5.8 m³/s),   q (3 - 12 g/m·s)
■ May 24 - 25, 1988,   0 (5.3 - 5.8 m³/s),   q (21 - 73 g/m·s)
• May 28 - June 6, 1988, 0 (3.9 - 5.8 m³/s), q (0.1 - 187 g/m·s)

Fig. 4: Particle transport rates versus total bedload transport rates.

Deposition and Removal of Fines in Gravel-Bed Streams

by

Panayiotis Diplas[1] and Gary Parker[2]

ABSTRACT

Laboratory experiments, which model gravel-bed streams with poorly sorted bed material, were conducted to examine the deposition of fines into and their release from a channel bed. Two flumes were used. The first one was long and narrow and operated as a recirculating system. The second flume was shorter and wider and operated as a feed system. Several features of gravel-bed streams, such as the existence of a pavement and a pool-and-riffle structure were reproduced in the laboratory. As it has been indicated from several field studies, these features play an important role in the infiltration and removal of fines in the bed of gravel-bed streams.

The fines initally introduced into the flume tend to collect in the immediate substrate, where they create a seal that restricts any deeper penetration within the bed material. As more fines are added, they saturate the subpavement and start appearing in the surface layer. The amount of fines in the pavement depends on the mean flow concentration of fines and other flow parameters. Higher amounts tend to deposit at the bar tail and the pool areas.

In the absence of bed load motion fines are released only from the surface bed layer. Particle entrainment allows for the removal of fines from the subpavement. Infrequent floods, that mobilize the whole pavement, tend to obliterate the pool-and-riffle structure and clean the channel bed from fines up to the bottom layer.

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INTRODUCTION

As an ecological phenomenon, the process of fines input to and accumulation in gravel bed streams falls in the category of non-point source pollution, and is considered to be one of the most significant pollution problems in this category. As indicated by a voluminous body of research, the presence of excessive amount of fines in the stream bed is one of the most important factors that could adversely affect the spawning success of salmon and other species of fish (Shea and Mathers, 1978; Koski, 1966).

The pool and riffle structure, which is characteristic of gravel bed streams, provides considerable variation in flow depth and velocity as a function of location that can in turn result in areas suitable for spawning activities for a variety of fish. The intergravel stage of life of the fish, as incubating egg first and as alevin later, represents a period they may be most susceptible to damages from high levels of fines. Increased amounts of fines may injure the ova, (Gibbons and Salo, 1973), reduce the intergravel flow which is necessary for the removal of the toxic metabolic wastes produced by the eggs, (Iwamoto et al., 1978), and lower the levels of dissolved oxygen supplied to the incubating eggs and alevins which is necessary for their adequate survival and growth (Phillips and Campbell, 1961). In addition, increased amounts of fines reduce the population and diversity of the benthic community which is the main food resource for the fish emerging from the intergravel environment (Gibbons and Salo, 1973). The reason for the paramount importance of fines is that, of all factors deleterious to spawning likely to be exacerbated by the activities of man, it is one of the most prevalent. This can be caused by human activities such as clear-cutting, road building and agriculture, which result to inputs of fine material into streams that are well above the inputs due to natural erosion processes. Furthermore, stream regulation tends to eliminate the annual peak discharges which are capable of activating the bed and thus releasing the fines that accumulate during low flows.

It is therefore evident that stream sedimentation is a serious source of stream pollution. In order to keep the pollution of a stream below dangerous levels, management criteria for the land use in the corresponding watershed should be developed. This requires the knowledge of the following three general relations:

1. A relation predicting the quantity and type of fines flowing into streams as a function of specific watershed activity, basin geological and vegetal structure, and hydrological regime.

2. A relation predicting the characteristics of deposition of fines in and their release from the stream bed as a function of fines input, flow discharge and other factors.

3. A relation indicating the effect of fines in the substrate on spawning success.

The purpose of the present study is to examine the second relation, the deposition and removal of fines in gravel-bed streams. The available literature on the subject is reviewed first. Subsequently, the results of laboratory experiments simulating the process of fines accumulation and retention in gravel beds, and their effect in the stream behavior are examined.

PREVIOUS RESEARCH

While the detrimental effects of fines on the survival and emergency of salmonid fry were recognized quite early (Harrison, 1923), most of the research on the mechanisms of deposition and erosion of fines in gravel matrices has been conducted during the last two decades. Both field studies and laboratory experiments have been performed.

Various aspects of the mechanics of fines retention in gravel-bed streams have been examined in field studies carried out with adequate instrumentation by Milhous (1973) on Oak Creek, Adams and Beschta (1980) on five small streams located in the Oregon Coast Range, Frostick, Lukas and Reid (1984) on Turkey Brook tributary, Lisle (1989) on three streams located in the Coast Range of northern California, and by Alonso et al. (1988) on Tucannon River.

Milhous (1973) verified the existence of a surface bed layer, called pavement, that is coarser than the subsurface material. He also observed that the role of the pavement on the deposition and retention of fines is quite important. At low flows the immobile pavement layer acts as a sink of the suspended sediment, which tends to deposit in the interface between the bottom of the pavement and the top of the subsurface, called the "silt reservoir". At higher stages, when the pavement is set in motion, the bed acts as a source by supplying fines for suspension, at rates determined by the pavement. The sand (<2 mm) fraction in the Oak Creek substrate takes a value near 12 percent. This value has been maintained over more than ten years, suggesting that it is in some sense in balance with the stream and watershed on a long-term basis.

In their study of five streams, Adams and Beschta (1980) found that the percentage of fines (<1 mm) within the stream bed varied between streams, between riffle areas in the same stream, and within the same riffle. It is also mentioned that the variation across the channel is typically more pronounced than it is along the channel. In addition to spatial variability, temporal variability of the fines in the substrate was observed. They attributed the spatial variation to differences in the size of the gravel substrate and fines, the supply of fines, and the hydraulic condition of the flow. For a single stream the amount of fine sediment in the bed was correlated to the stream sinuosity and bankfull stage; an equivalent set of flow parameters will be depth and velocity of flow. The temporal variability of the fines is attributed to the flushing of fines, which takes place mainly during high flows when the bed is set in motion. Furthermore, an extensive sampling program undertaken by Adams and Beschta (1980) on twenty one Coast Range streams during summer low flows indicated an average amount of fines content within the channel bed of 19.4 percent; for the streams on undisturbed watersheds the level of fines ranged from 10.6 to 29.4 percent, while for the streams on disturbed watersheds it reached a high value of 49.3 percent. The fines content was stratified with depth; an average value of 17.4 percent was measured for a 10-cm surface bed layer, while the corresponding value for a 30-cm subsurface bed layer was 22.3%.

Frostick et al. (1984) monitored the amount and type of fines infiltrating the pavement of Turkey Brook by installing six traps at its bed. Each trap had an opening of $0.5m^2$ and was 0.33 m deep, and was divided into sixteen removable compartments, that were filled with different size of substrate material. A lid of pebbles, having the size distribution of the pavement layer at Turkey Brook, covered each trap rendering its surface flush with the surrounding bed. Their results indicate that the presence of a coarser surface bed layer and bed load motion are the dominant parameters influencing the amount and size of the infiltrating fines. The coarse surface layer superimposed on a finer substrate, which is typical of gravel bed rivers, was found to encourage the clogging of the near-surface pores

with matrix fines. Flow events that caused bed activation resulted in higher amounts and larger size of ingressed fines. The intruding fines exhibited spatial variation, with higher amounts in pools compared to bars, while for the same cross-section the area of highest fines content coincided with the area of highest velocity. During high flows, that were capable of mobilizing the channel bed, fines were flashed from the bed up to a depth of 8 cm at the thalweg. This depth is approximately equal to 2 D_{p90}, where D_{p90} is the sieve diameter of the grain that is coarser than ninety percent of the pavement material.

Lisle (1989) studied the sedimentation of spawning gravel beds in three streams of north coastal California. He buried a number of cans, 17 cm in diameter and 22 cm deep, filled with well sorted subrounded gravel, below the pavement, in areas likely to be used by anadromous fish for spawning. He measured th. amounts of fines deposited into the cans as well as the depth of their infiltration. Lisle's observations are similar to those of Frostick et al. (1984). He found that silts and very fine sands were concentrated at the bottom of the cans, while sand and granules were caught in interstices near the top forming a seal. This seal penetrated deeper in areas of higher energy. The average depth of the infiltration of fines was 2.6 D_{90}, where D_{90} is the 90th grain size percentile of the subsurface material. The suspended sediment comprised the bulk of the total sediment transport, but it constituted only one-fifth of the infiltrated material found in the cans. This is attributed to the better access that the finer bed load particles have to the pores of the pavement, being continuously in contact with the bed. The fine sediment ($0.25 < D < 2$ mm) that accumulated in the subsurface, averaged overall the cans of a stream, correlated well with the total amount of bedload in the same size range, transported during storms. Furthermore, Lisle observed that scour of stream beds during floods and subsequent fill may deposit at least as much and as deep fine sediment in clear gravel beds as infiltration. With parts of the stream bed being scoured during individual storms to depths of 0.1 m, incubating eggs could be crushed. Therefore, scour and fill could be equally crucial to the survival of eggs and embryo, as is the infiltration of fines.

The purpose of the Alonso et al. (1988) field study was to calibrate a computer program that simulates the deposition of fines in coarse gravel matrices and its effect on the quality of aquatic habitat. They also reported that scour at some locations of the Tucannon River was severe enough to affect salmonid eggs. Contrary to the account given by others (e.g. Lisle, 1989), however, Alonso et al. observed that the infiltrating fines did not create a seal within the bed subsurface, and that their amount was directly related to the suspended load. The different depositional behavior of the fines infiltrating the Tucannon River substrate is mainly due to the large difference in size between the framework gravel and the matrix material, a situation that is not commonly encountered in natural streams.

Laboratory experiments on the characteristics of fines intrusion and retention into gravel beds have been carried out by Einstein (1968), Beschta and Jackson (1977), Dhamotharan et ?'. (1980), Carling (1984), and O'Brien (1987).

In his pioneering work, Einstein studied the intrusion of silica flour (3.5 to 30 microns) into a gravel bed. He used two flumes with gravel mixtures as bed material, and water with fines were recirculated at flows not large enough to move the gravel. He noted that the fines settled into the gravel reaching the bottom of the flume and gradually filled the pores from the bottom up. Einstein's experiments represent an important contrib: ion in the siltation of gravel matrices when almost all the sediment moves in suspension, and the bed material is considerably coarser and immovable. They give, however, a simplistic and probably misleading description of the phenomenon in the light of field data. The main deficiencies of these experiments are the lack of pavement, the absence of bed load motion and pool-and-riffle structure, and the great size difference between the coarse and fine sediment phases. The last factor, which is probably the most significant, dictated the mode

of fines deposition. The first factor may also be crucial in the creation of a seal within the substrate. The second and third factors may provide important cleaning mechanisms of the gravel bed from the fines (Vaux, 1968; Alonso et al. 1988; Lisle, 1989).

In their experiments, Beschta and Jackson (1979) used well-sorted gravel with D_{50} = 15 mm as bed material, and two uniform sands with D_{50} = 0.5 mm and 0.2 mm as fines. Water and fine sediment were fed at the upstream end of the flume, and gravel bed load, except for a jiggling motion, was essentially absent. When the fines with 0.2 mm median size were fed into the channel, the gravel pores filled from the bottom up in the Einsteinian mode. The use of the coarser fines, however, resulted in a sand seal. Increasing Froude numbers forced the seal deeper into the gravel substrate.

Dhamotharan et al. (1980) conducted a physical model study of Oak Creek to examine whether phenomena characteristic of natural gravel bed streams can be reproduced in a laboratory flume. The flow rate in the experiments was capable of entraining even the coarser grains of the bed material. The resulting transported material was recirculated with the water through the flume. A pavement and a weak pool-and-riffle structure, similar to those observed in the field, were developed in the flume. After the equilibrium stage was reached, some preliminary experiments were carried out with fines feeding. It was found that the fines collect predominantly between the subpavement and pavement, and that the accumulation of fines resulted in a reduction of the bed load transport rate when compared with the same experiment but in the absence of fines.

The purpose of Carling's (1984) flume experiments was to determine the role of gross flow parameters on the siltation rate of sand into a gravel bed. He used well sorted gravel, with median size of 15.6 mm, as bed material and three different grades of sand (D_{50} = 0.15 mm, 0.19 mm, 1.4 mm) as fines. The flow was not capable of entraining the gravel, so no self-formed pavement was present. The coarsest sand was the only one to form a seal. He concluded that the initial flow concentration of sand was the only mean flow parameter that correlated well with the siltation rate. He observed that the siltation rate was high even at low flow concentrations, and that the surface bed layer remained free of sand.

O'Brien (1987) carried out field work in Yampa River and subsequently a full scale physical model study of the same river. His emphasis was on determining minimum streamflow requirements necessary for the maintenance of fishery habitat. In agreement with other researchers, he observed that in the absence of gravel mobilization the bed could be cleaned from the fines to a depth close to the median size below the bed surface.

In addition to the above mentioned field and laboratory studies, several researchers have suggested methodologies for flushing gravel matrices from fines that would result in acceptable quality of aquatic habitat. Most of these methodologies are purely empirical in nature and have been reviewed by others (Reiser et al., 1985; Milhous, 1986).

PRESENT EXPERIMENTS

The purpose of the present study was to study the process of deposition and removal of fines in a gravel-bed stream under conditions encountered in natural streams. Two flumes were used to meet this objective. The first flume was 16.75 m long, 0.3 m wide, and 0.61 m deep, and the second one was a tilting flume 12 m long, 0.91 m wide, and 0.38 m deep. They will be referred to as the long and the tilting flume, respectively.

Most of the experimental work was carried out in the long flume, which recirculated both sediment and water. The major advantages of this flume are its length and its transparent glass walls. The bed load motion, and the infiltration of fines into the bed can be observed closely through the walls. The narrow width of the channel, which suppressed any tendencies for the formation of bars, is its major disadvantage.

The tilting flume facility operated as a feed system in both water and sediment. Two sediment feeders are used at the upstream end of the channel; one is introducing into the flume material identical to the original bed mixture, and the other feeds the flume with fines. The important feature of the tilting flume is its width, which was sufficient to allow for bars to develop. The short length of the channel and its opaque walls are its major drawbacks.

The mixture used as bed material was poorly sorted, with median grain size of 2.44 mm and standard deviation of 2.75. Two sizes of silica flour were used as fines, one with D_{50} = 0.08 mm and the other with D_{50} = 0.11 mm, both being well sorted. The white color of the fines was very helpful for observation purposes.

Six series of experiments were conducted in the long flume and a single series in the tilting flume. The initial bed slope for the long flume ranged from 0.003 to 0.008, while for the tilting flume it was 0.012. The dimensionless design Shields stress varied from 0.10 to 0.15, based on the mean size of the original mixture. On the average, six experiments were conducted for each series.

Before the beginning of each series a fresh mixture of bed material was introduced into the channel and was screeded with a template to the desired initial slope. The first experiment of each series in the long flume was performed in the absence of fines. During the rest of the experiments in each series, fines were introduced into the flume, with their amount being progressively increased with each experiment.

Seven experiments were performed in the tilting flume. No fines were fed during the first experiment. During the rest of the experiments, the impregnation of fines into the gravel bed was examined, as well as the ability of the flume to purge itself of the fines accumulated in the bed.

Each experiment was assigned a symbol, e.g. S2:E3, where S stands for series and E for experiment. The above symbol thus refers to the third experiment of the second series. The tilting flume experiments belong to the seventh series.

RESULTS

The field studies reviewed earlier indicated that gravel bed streams are typically characterized by the presence of a pavement, bed material that is poorly sorted, bed shear stress that moderately exceeds the critical value for particle entrainment, and by frequent bed activation. In addition the size of the finer matrix material overlapped, at least partially, with the size of the coarser framework particles. The same studies demonstrated the importance of these factors in the process of fines infiltration into and their removal from a gravel bed. Nevertheless, most of the laboratory studies did not simulate several of these features in the experiments of the deposition of fines in gravel matrices.

The purpose of the first experiment of each series was to allow for a self-formed pavement layer to develop from the original bed material. In the tilting flume case a pool-and-riffle structure was also present at the end of the first experiment. The time required for the channel to reach equilibrium ranged from 30 to 54 hours. During this period a significant part of the finer material ($D < 0.5$ mm) infiltrated below the surface layer and collected within the subpavement. At the completion of the experiment the bed material was sampled by removing three consecutive layers. Wet clay was forced against the bed surface to remove the top layer, or pavement, whose thickness was about 0.65 cm ($\approx D_{90}$). The subsequent two layers, called subpavement and bottom layer, were extracted with a spoon, both being about 1.5 cm ($\approx 2.3 \ D_{90}$) thick. The top sample is close to an areal while the other two are close to a volumetric (Diplas and Fripp, 1990). The size distribution of these three layers and that of the bed load material obtained during the fourth series are shown in Fig. 1. Although areal samples cannot be compared directly with volumetric samples, Fig. 1 indicates that the pavement is coarser than the subpavement (Diplas and Sutherland, 1988). The latter was also evident from side by side visual comparison of these two layers. Typical relations among subpavement median size, D_{50}, bottom layer median size, D_{b50}, and bedload median size, D_{l50}, are $D_{50} \approx D_{b50}$ and $D_{l50} \approx 1.14 \ D_{50}$. In addition, the subpavement contains considerably higher amounts of finer particles ($D < 1.0$mm). At the end of the first experiment in the tilting flume, a well developed pool-and-riffle sequence in conjunction with three bars, that were alternatively attached to the channel side walls, had formed. Detailed sampling of the bed material composing the surface of a bar revealed a gradual decrease in grain size in the downstream direction. The resulting bed configuration and variation of the grain size along a bar are in good agreement with field observations (Lewin, 1976; Bluck, 1982).

The characteristics of the subpavement material for both the tilting and long flume experiments were essentially identical. The same holds true for the material of the corresponding bottom layers. Thus, the only difference, in terms of bed material, between the experiments without fines in the two flumes was restricted to the surface layer.

Experiments with Fines in the Long Flume

Starting from the second experiment of each series, fine material was introduced into the flume. The following program of incremental increase of fines supply was used for the experiments of series two to five. The amount of fines introduced into the flume during the second experiment accounted for a system concentration of 2,000 mgL^{-1}, where the system concentration is defined as the ratio of the mass of fines and the total volume of water in the system (flume and pipes). The fines added during the third, fourth, fifth and sixth experiments corresponded to system concentrations of 3,000 mgL^{-1}, 5,000 mgL^{-1}, 10,000
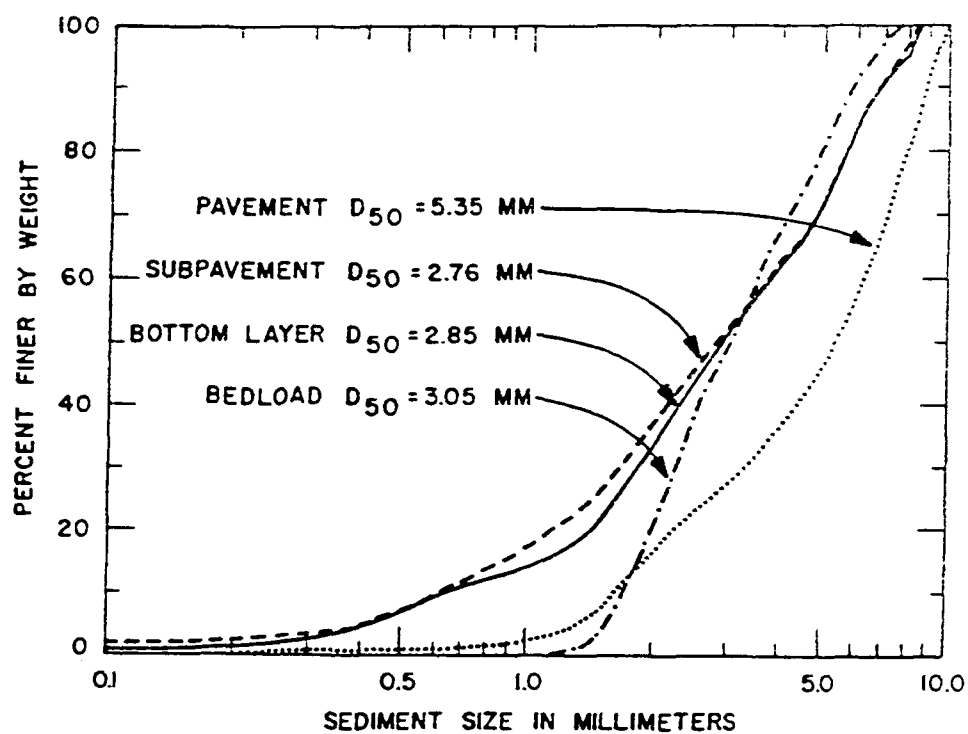
Fig. 1.   Pavement, subpavement, bottom layer, and bed load
size distributions for the experiment S4:E3.

mgL^{-1}, and 10,000 mgL^{-1} respectively. The total amount of fines introduced into the long flume during a single series of experiments ranged from 19.5 to 27 kgr. During series four and five, fines with 0.11 mm median diameter were used. For the rest of the series (one, two, three and six), fines with a median diameter of 0.08 mm were used.

Fines have been observed to infiltrate into gravel beds in two different ways. The first one is characterized by unobstructed settling through the voids of the framework material. In the second way the fines are inhibited from settling freely, and instead they bridge the gaps among particles of the gravel bed, thus creating a seal that limits any deeper penetration of fines. The mode of deposition is mainly controlled by the relative size of the settling fines and the voids of the framework gravel (Sakthivadivel and Einstein, 1970; Sowers and Sowers, 1970). The presence of a coarser surface bed layer promotes the latter mode of deposition (Frostick et al., 1984) with clogging taking place close to the bed surface. When the subpavement material is poorly sorted, the possibility of a seal forming near the bed surface is further enhanced.

To demonstrate visually the infiltration of fines into the channel bed, approximately six bed layers were cleaned of their finer particles over a segment of the channel bed during the first series. Figures 2 and 3 show two views of the channel bed from the side wall at the end of the sixth and seventh experiments. In Figure 2, the white contour identifies the maximum depth of infiltration of the silica flour. As shown in Figure 3, fines subsequently added to the flow deposit at all levels above the white contour, with the bulk of the filling progressing from this contour upwards to the gravel surface. Futhermore, these two figures show that the difference in size between the depositing fines and the coarse framework material dictates the depth of deposition of the fines.

Prediction of the seal thickness would be important for determining the effects of fines on spawning grounds and the possibility for their flushing. During the present experiments the added fines never infiltrated deeper than the bottom layer of the channel bed in any appreciable amounts. Most of the times the top of the bottom layer corresponded to the deepest penetration of the fines. This depth was about 2.1 cm from the bed surface or in terms of size of bed material $\approx 3D_{90}$. For a more conservative estimate it is considered that the maximum infiltration depth is $\approx 5D_{90}$, which corresponds to the bottom of the bottom layer. From his field study, Lisle (1989) found an average value of $2.6D_{90}$ for the thickness of th. seal, while Beschta and Jackson (1979) suggest values ranging from 2.5 to 5.0 D_{90} based on their laboratory experiments. Figure 4 shows the pavement, subpavement and bottom layers for the experiment S5:E6.

The sixth series was designed to examine the influence of the added fines on the bed load transport rate. During this series considerable variation in the bed load transport rate was observed. The water surface slope and the velocity profile, however, did not exhibit any noticeable changes during this series. It is therefore suggested that the factor likely to be responsible for the otherwise unexplained bed load variation is the accumulation of fines in the channel bed.

The grains that participate in the bed load transport are typically entrained from the surface layer of the bed material. Small amounts of added fines normally settle in the subpavement and are not expected to interfere with the bed load motion. As more fines accumulate in the bed, they start impregnating the pores of the surface layer and thus increasing the stability of surface grains, resulting in a gradual reduction of the bed load transport rate. As even more fines deposit into the channel bed, the subpavement gradually becomes saturated with them, and the interstices between the grains of the

Fig. 2 A view of the fines infiltrating into the channel bed from the sidewall of the long flume for S1:E6. The bed material has been locally cleaned of its finer grains for a few coarse layers.
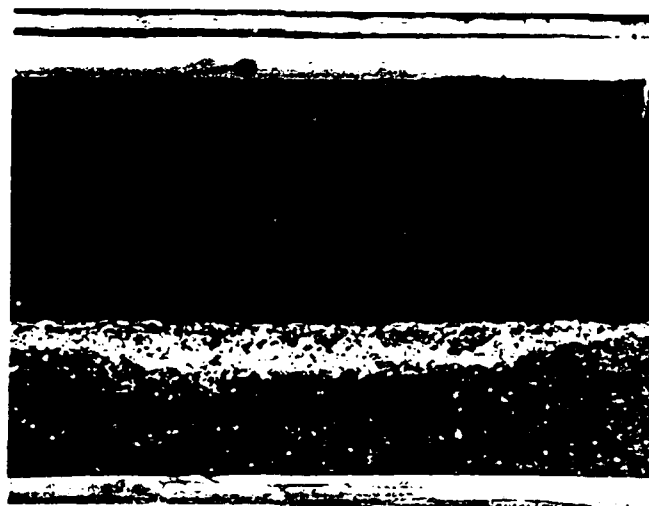

Fig. 3 A view of the fines infiltration from the sidewall, of the flume site shown in Figure 2, at the end of run S1:E7.

Fig. 4 Photographs of (a) the pavement and subpavement layers and
(b) the pavement and bottom layers, from S5:E6.

pavement start to fill. During this process the mobility of the pavement particles is further reduced, while the bed surface becomes smoother, thus increasing the mobility of the grains moving on top of it. Eventually these opposing tendencies balance each other and the bed load rate stops declining. A further increase of the amount of fines in the pavement enhances the mobility of the moving grains resulting in an increase of the bed load transport rate. In this state, the number of moving grains has been significantly reduced, but the rest periods between motions are much shorter compared to grains moving on a pavement in the absence of fines. By adding even more fines, the bed gradually changes from gravel to sand as the gravel is buried. At this state ripples, bed forms characteristic of sandy streams, appear on the bed of the channel (Figure 5).

From the present experiments it was roughly estimated that when the amount of the deposited fines comprised about five percent of the subpavement material, the bed load transport rate started to decline. The lowest bed load transport rate observed represents a 57 percent reduction in the bed load value in the absence of fines. Dhamotharan et al. (1980) observed a 30 percent reduction under similar conditions. When the deposited fines comprised about five percent of the pavement material the bed load rate started to increase.

Experiments with Fines in the Tilting Flume

Fines, with a median size of 0.11 mm, were fed into the tilting flume during the second experiment at a rate of 2.0 gr/sec, which corresponds to a feed rate concentration of 156 mgL^{-1}. The feed rate concentration is defined as the ratio of the fines feed rate and the water volume discharge. The water discharge and coarse sediment feed rate of the second experiment were identical to those of the first experiments. The only thing that changed during the third experiment was the fines feed rate concentration, which became 350 mgL^{-1}. The coarse sediment feed rate was interrupted during the fourth experiment, the flow rate was decreased to the point that bedload transport become negligible, and the feed rate concentration of fines was adjusted to 430 mgL^{-1}. No fines were fed into the flume during the fifth experiment, keeping everything else the same as for the previous experiment. The sixth experiment was identical to the first one. In the seventh experiment the flow rate and coarse sediment feed rate were considerably increased, while no fines were fed into the flume.

The purpose of the second, third and fourth experiments was to examine the despoitional pattern of fines in a steam with pool-and-riffle structure, under various flow conditions. The ability and extent to which the stream could purge itself of the fines was tested during the last three experiments.

At the end of the second and third experiments samples obtained from the channel bed indicated that higher amounts of fines were deposited at the bar tail (downstream end of the bar) and in the pool area, or in those parts of the channel bed where the flow is fairly tranquil. Reflecting the higher feeding rate of fines during the third experiment an increase in the deposited fines was observed in this experiment. Further increase in the deposited fines occurred during the fourth experiment in the absence of bed load motion. Fine material was present in the pavement layer, with larger quantities in the bar tail and pool areas. Frostick et al. (1984) also found larger quantities of fines being deposited in the pool compared to the deposition in the bar head during their field tests.

Fig. 5 View of the channel bed at the end of run S6:E7. A rather well developed ripple configuration is present on the channel bed. The flow direction is from the top to the bottom of the photograph.

The ability and extent to which the channel was able to clean its bed of fines, in the absence of bedload motion, was examined in the fifth experiment. The fines that were deposited in the subpavement and bottom layers remained intact. The pavement was partially cleaned. In agreement with observations from the previous experiments, areas of swift flow, such as the bar head, experienced more thorough cleaning of the pavement.

Other researchers (Milhous, 1973; O'Brien, 1984) have also emphasized the necessity of pavement mobilization for removing fines below the pavement layer.

The cleaning process under conditions of broken pavement was examined during the sixth experiment. Both the pavement and subpavement layers were effectively purged from the fines during this experiment, while the fines in the bottom layer were not affected.

The last experiment was designed to simulate a rather infrequent flood, and examine the effectiveness of such a flood in the process of cleaning the bed from fine material. The bed load transport rate in this experiment was about two and one half times the bed load transport of the sixth experiment. The pool-and-riffle structure was obliterated during the seventh experiment, rendering the channel bed flat. It is believed that this experiment modeled the scour and fill process of streambeds that is encountered in natural streams (Alonso et al. 1988; Lisle, 1989). Samples of the channel bed material revealed the removal of fines from the pavement, subpavement, and bottom layers.

Finally, it was observed that the behavior of the added fines ($D_{50} = 0.11$ mm) was different from the behavior of the finer particles (0.177 mm $< D <$ 0.42 mm) of the original bed material. While the amount of fines in the channel bed increased during the first four experiments and decreased during the last three, the percentage of the finer portion of the bed material remained roughly constant throughout the series. In addition, the added fines can be transported in suspension by the channel flow while the finer bed material grains are too heavy to be suspended in significant quantities. It is therefore suggested here that the grains that are likely to be removed by the flow from the subsurface bed material are only those that can be suspended by the flow.

SUMMARY AND CONCLUSIONS

The results of the present experiments and those of earlier studies provide a good description of the process of fines settling to and removal from coarse bed material. As soon as fines are introduced into the channel, they start depositing on the bed, infiltrating below its surface. The depth of infiltration depends on the difference in size between the penetrating grains and the coarser bed material, and the boundary shields stress. Both types of fines used during the present experiments created a seal within the substrate. This seal was deeper into the substrate for the fines with median size of 0.08 mm, and it never exceeded a depth of 5 D_{90}. The seal is pushed deeper in areas of higher shields stress. The amount of fines than can ultimately deposit within the subpavement layer is independent of the boundary shear stress or other flow parameter. As long as there are fines in the channel flow moving in suspension or as bed load, they will keep infiltrating in the bed until they have saturated the subpavement layer. The fines that settle in the surface bed layer do interact with the flow. It is suggested that their amount depends on the boundary shields stress, the fall velocity of fines, the particle Reynolds number, and the mean flow concentration of fines (Diplas and Parker, 1985).

In the presence of an alternate bar structure on the bed, and when the flowing water covers these bars over their entire length, then the fines first removed from the flow are most likely to be deposited in the subpavement of the bar tail and the pool. As more fines deposit into the bed, they progressively infiltrate the subpavement throughout the channel bed, while fine grains start appearing in the pavement of the pool and the bar tail. With additional fines being introduced into the flume, the subpavement eventually becomes saturated, and the fines content in the surface layer increases considerably. The maximum mean flow concentration of 3,000 mgL^{-1} in the tilting flume experiments exceeds the fines concentration typically encountered in natural streams during peak discharges (Walling and Webb, 1987; Lisle, 1989).

The bed load transport varied as a result of the fines that accumulated in the channel bed. This bed load variation correlates well with the amount of fines in the subpavement and pavement.

In the absence of bed load motion, a channel flow without any suspended material can at most remove the fines from the surface layer of the bed. Conditions of mobilized pavement are required if layers below the surface are to be purged of the fines. Sporadic bed load motion does not affect the bed configuration and removes the fines to the bottom of the subpavement ($\approx 3D_{90}$). During the cleaning of the bed, fines are first removed from the bar head and subsequently from the pool and bar tail. General bed load motion, which occurs during major floods, tends to efface the pool-and-riffle structure and removes the fines from up to the bottom layer of the channel bed.

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THE SIZE DISTRIBUTION OF SALMONID SPAWNING GRAVELS

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for presentation at the
Third International Workshop on Gravel-Bed Rivers
Firenze, 24-28 September 1990

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ABSTRACT

Salmonids (salmon and trout) require freshwater stream gravels for spawning. Spawning gravels have been identified by biologists as limiting salmonid populations in many settings, either because of a shortage of gravels with suitable framework sizes or because of a large proportion of interstitial fine sediment. As a result, many studies have been conducted on spawning gravels since the 1950's, mostly by fisheries biologists, and mostly appearing only the the grey literature. Unfortunately, results have been inconsistently reported in many cases, with many studies reporting only one summary statistic (different for each study) from the gravels, rendering comparisons among studies difficult. This study compiled data from many previous studies and from original field work to determine what the distinguishing characteristics of spawning gravels might be, how they vary with the size of the spawning fish, and whether they differ in sedimentologic characteristics from fluvial gravels reported in the geologic literature.

A total of 135 spawning gravel size distributions were compiled by species. Most of these distributions represent the average or composite of numerous individual samples. The median grain sizes of these distributions ranged from 5.4 (for coho salmon redds in Flynn Creek, Oregon) to 78 mm (for chinook salmon in the Columbia River near Vantage, Washington, with a grand median of 22 mm. Values of geometric standard deviation range from 1.6 to 21, with half the values between 3.3 and 5.2. Nearly all of the distributions are negatively skewed, with half of the skewness coefficients falling between 0.24 and -0.39.

These spawning gravels appear to be somewhat coarser than most gravel size distributions reported in the geologic literature, but they do not differ in sorting or skewness. Contrary to Pettijohn's (1975) suggestion that most gravels are bimodal, very few of the spawning gravel size distributions analyzed here were bimodal.

The results also confirm the widely held assumption among fisheries biologists that bigger fish can spawn in in bigger gravels. However, although these larger fish are capable of using larger gravels, they are not obligated to do so and often spawn in smaller gravels. The relation between fish size and spawning gravel size is thus described by an envelope curve, which can be used as a general guide to the likely gravel requirements of various fish. Fish 30-cm long can use gravels up to about 30 mm in size, while 90-cm-long fish can use gravels up to 80 mm in size.

PROBLEM STATEMENT

The availability of gravels of sizes suitable for spawning has been recognized as a factor limiting salmonid populations (Allen 1969, Reiser and Bjornn 1979). Although experienced field biologists can usually identify a suitable spawning gravel when they see one, prior to this study, there has been no systematic attempt to compile a wide range of gravel size distributions from the literature for purposes of comparison, to determine what their distinguishing characteristics might be, how they vary with the size of the spawning fish, and whether they differ in sedimentologic characteristics from fluvial gravels reported in the geologic literature.

A great deal of information on spawning gravel sizes has appeared in the fisheries literature since the 1950's, but it has been inconsistently reported. Most publications have not reported the same measures, nor included the complete size distributions (from which other descriptors could be computed), so comparisons are difficult. A number of papers have been occupied in a quest for a single variable descriptor (one statistic drawn from a gravel size distribution) for use as an index of gravel quality, and following these leads, some authors have reported only one statistic from their gravel samples (as reviewed by Kondolf 1988). This trend has been unfortunate, because the gravel size requirements of spawning fish vary with life stage, and the appropriate measures will vary accordingly. Thus, no single statistic is perfectly suited as an index of gravel quality, and the failure to report the entire size distribution makes comparisons difficult among studies.

It is widely acknowledged among field biologists that big fish can spawn in bigger gravels. This is due not only to the greater upward pull that larger fish can exert on the stream bed, but also to their ability to hold in faster currents, which assist them in moving larger particles. Larger gravels also tend to be associated with higher velocities, so these variables are not independent. Behavioral patterns also confound the relation between fish size and gravel size. For example, pink salmon typically spawn in low-gradient reaches of coastal rivers, where gravels are small and contain abundant fine-grained sediment. In contrast, silver and chinook salmon may ascend rivers to headwater reaches, where gradients are high and bed material is larger.

This study examined the range of size distributions reported in the literature orm spawning gravels and other gravels in the geologic record. The literature-derived data were augmented by original field work.

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METHODS
Compilation of Size Distributions and Redd Characteristics
   At the outset of this study, there appeared to be relatively
few published spawning gravel size distributions in existence, so
their compilation promised to be a relatively simple task. As
the literature search and train of telephone contacts and
correspondence proceeded, however, the existence of more and more
gravel size distributions came to light. Most of these have
appeared in the "grey" literature, in such outlets as state and
federal fish and game agency reports, theses, and consultant
reports; others exist only as unpublished data. The data
presented here were drawn from size distributions of spawning
gravel samples collected for this study and from 22 other
sources. Of these other sources, only four in the open
literature; 18 were in the grey literature or unpublished, and
many were obtained only by direct request of authors or
originating agencies. Besides those mentioned above, many other
reports were obtained and examined, but were not included in
these tables, either because full size distributions were not
presented in them or because too few sieves were used to
adequately characterize the size distribution.

   In many sources, the size distributions presented were averages
or composites of many samples collected in the study area. This
was a constraint that dictated, in large measure, the approach
used here. Because only average values were available from many
studies, for purposes of comparison, average values were computed
from studies in which individual size distributions were
reported. Individual size distributions were utilized in other
analyses (Kondolf 1988).

   Some published gravel size distributions are for samples
obtained in redds, others from potential spawning gravels. The
latter are unspawned gravels known to be representative of those
used for spawning either because of historical observations of
spawning at the site or because they were collected adjacent to
redds. The spawning female can effect a change in the size
distribution of gravel, so the two populations (redds and
unspawned) are not strictly comparable. Curiously, many
published studies refer only to "spawning gravels" without
explicitly stating whether the samples were obtained from redds
or from potential spawning gravels. In this study, potential
spawning gravels and gravels in redds were treated as two
distinct populations. Data collected for this study included
samples of both redds and adjacent, unspawned gravels (Kondolf
1988).

   To analyze the size distributions, the general procedure was to
plot the cumulative size distribution curve and to read off
percentile values needed to calculate size descriptors for

comparison. Size descriptors compiled from all sources were the median, D50, geometric mean, dg, sorting index, sg, skewness, sk (Inman 1952, Vanoni 1975), and graphic mean, mg (Folk 1980). These are computed as:

$$dg = [(D84)(D16)]^{0.5}$$
$$mg = [(D84)(D50)(D16)]^{0.333}$$
$$sg = (D84/D16)^{0.5}$$
$$sk = \log(dg/D50)/\log(sg)$$

In some cases, cumulative size distributions were presented in the source; in others they were computed from the presented values of weight or percent retained on each sieve. Only histograms or plotted cumulative curves were presented in some sources, so values were read from the graphs, and two sources presented values of D50, dg, or sg, which were directly incorporated in Tables 1-3. Length data for the spawning fish obtained from published sources or by contacting the authors or other experts familiar with the species in the region (Kondolf 1988).

Sorting and skewness for averaged and individual cumulative frequency distributions were plotted against mean size to ascertain if they varied systematically with grain size. Redd characteristics (mean grain size, redd area, and velocity and depth at the redd site) were plotted against fish length to define envelope curves of these redd characteristics as functions of fish size (Kondolf 1988). The degree to which the gravels are bimodal, as suggested by Pettijohn (1975) was assessed by inspection of cumulative size distribution curves. The degree to which gravels are lognormally distributed was assessed by plotting selected curves on log-probability paper and by comparing attributes of the sample distribution with values of the attributes expected for a normal distribution (Kondolf 1988).

LIMITATIONS OF THE DATA
Definition of Tails

In many studies, size distribution curves were not well defined at the tails because only a small number of sieves were used. In fact, data from a number of sources could not be used in this study for that reason (e.g., McNeil and Ahnell 1964, Burns 1970, Peterson 1978, and many samples of Platts et al. 1979). Other studies did not define the upper limit of the largest size class, denoting the largest class only as "larger than" some size. Inadequate definition of the fine tails of distributions was encountered in fewer studies (e.g., Chambers et al. 1953,1954), probably because assessment of fine sediment has been an objective of most recent studies.

Failure to define the upper limit of the largest class produces

a problem when plotting the cumulative curve because one doesn't know what size should correspond to "100 percent finer than." In compiling data for this study, the observed progressions of sieve sizes were generally extrapolated to set the maximum size, but this often required some judgement. For example, if 1-inch (25 mm) and 2-inch (50 mm) sieves were used, is the next logical step 3 inches (75 mm) or 4 inches (100 mm)? (Smaller sieves often followed neither an arithmetic nor geometric progression, so could not be used as a basis for making the judgement). Depending on the percentage caught in the coarsest sieve, the choice could influence the value of D95 and perhaps D90. If the coarsest sieve contains more than 16 percent of the sample, the value of D84 will also be influenced.

The problems of inadequate definition of the distribution (because of too few sieves) and failure to define the upper limit of the largest size class are compounded by the potential errors introduced by the small sizes of many samples of coarse gravels, which may lead to substantial errors, especially in calculation of the D95, D90, and D84 (Church et al. 1987). It is often not possible to obtain a rigorously large sample of coarse gravel in a redd, because the redd itself may not contain enough gravel.

A number of authors have reported that they discarded large rocks from their samples. Of the data sets used in Tables 1-3, the only authors who reported excluding large rocks were Chambers et al. (1954,1955), who excluded rocks larger than 152 mm, Helle (1970), who excluded rocks larger than 100 mm, and Hobbs (1937), who excluded rocks larger than 76 mm. It is possible that other workers also excluded large rocks, but made no statement to that effect in their description of methods. Exclusion or inclusion of large rocks in samples is a potentially large source of variability among studies because exclusion of large rocks affects the percentages assigned to all other size classes.

D95, D90, D10, and D05 values were read from cumulative curves and kurtosis values were computed by Kondolf (1988). However, these percentile values were frequently poorly defined, so kurtosis values may not be meaningful and are not presented here.

Spatial and Temporal Variability

Most entries in Tables 1-3 represent averages or composites for a number of samples collected in one river or stream. The number of samples represented by the entries ranged from 1 to 310. While the average values are probably most representative of typical conditions at the site, they may mask substantial variability among individual samples arising from differences in microenvironment of deposition among sample sites.

In addition to this spatial variability, temporal variability in gravel size at one site may be pronounced. Variations over

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time in fine sediment content have been documented by Cederholm and Salo (1979), Adams and Beschta (1980), and Scrivener and Brownlee (1982). These temporal variations presumably result from seasonal variations in sediment transport and may reflect, to a large extent, changes in the population of mobile fine sediment in comparatively stable gravel beds. However, gravel deposits themselves may be subject to wholesale replacement at some sites in some years. Kondolf et al. (in press) painted potential spawning gravels at seven sites in four streams in eastern California and documented that all particles had been scoured and washed downstream during the high flows of 1986 (recurrence interval about 7 years). At three of the nine sites, new gravel of size suitable for spawning had moved in to replace the washed-out gravel, but at the other sites, the suitably-sized gravel had been replaced by cobble and coarse gravel, which was probably too coarse for spawning by resident brown trout. This wholesale gravel washout does not occur every year; a new set of tracer gravels emplaced at the same sites were undisturbed over the 1987 season (recurrence interval < 1 year).

Thus, the gravel size descriptors shown in Tables 1-3 (except for samples from stable terraces) reflect conditions at the time of measurement. In another year or season, bed material composition might differ.

Influence of Study Site Selection

Finally, it should be recognized that the gravels sampled and the observed hydraulic conditions at sample sites (reported in Kondolf 1988) also reflect the choice of study areas by investigators and the limitations imposed by working conditions. Observations made in typical chinook salmon stream in the Pacific Northwest can be reliably accepted to reflect typical conditions for many populations. However, fish may adapt to different environments and use very different hydrologic conditions. For example, the large chinook salmon of the Kenai River, Alaska, utilize velocities and depths far greater than recorded elsewhere (entry 28, Table 1). Similarly, chum salmon of the Susitna River utilize sites with extremely fine-grained surface material but which are characterized by upwelling groundwater (entry 131, Table 2). If criteria developed from observations elsewhere were applied to these sites, they would be deemed unsuitable for spawning, but both these populations reproduce very successfully. The conditions most commonly studied are likely to be those most commonly encountered by fishermen and fishery biologists. Sites with difficult access may be avoided for logistical reasons, and sites with unusual conditions may be avoided simply because they are not "representative" of most spawning by the species. Thus, the unusual circumstances in which the adaptability of the fish

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is displayed will be less often studied. Finally, field studies
in fisheries research may follow the same patterns as field
studies in geomorphology, which tend to concentrate in areas near
major research universities or field research centers (Graf
1984).

In addition to the potential bias introduced in the choice of
study area, the limitations of working conditions may impose a
further bias. In streams and rivers whose range of available
habitat includes deep and swift sections, the maximum depth
recorded at redds is typically just under the height of the
observer's chest waders. Redds in deeper parts of the channel
may not be visible, or, if they are, it may be impossible to make
measurements there by wading. Similarly, samples may be
unobtainable with conventional bulk core or freeze core samplers.

Burger et al. (1983) radiotagged schinook salmon in the kenai
River and used boats to measure velocity and depth over redds in
deep waters; to collect gravel samples, they returned during
lower flows. Without the technical advance of radiotagging,
biologists might still not know that fish could spawn in such
deep waters.

To measure particle size in deep water, divers have been used
to collect samples (Klingeman and Emmett 1982) or to take scaled
photographs (Chapman et al. 1983). Clearly, these techniques
require a substantial investment of time, effort, and equipment.
Little wonder that most studies are conducted on sites with good
access and shallow flows.

An illustration can be drawn from field work conducted for this
study in the Grand Canyon, where portability was an important
consideration because access to sites was by river raft with
limited cargo space. Samples of rainbow trout spawning gravels
were collected with a portable core sampler that cannot operate
in water deeper than about 50 cm. At one site (Tapeats Creek),
redds in deeper water could not be included in the sample
population.

RESULTS AND DISCUSSION
Presentation of Data

Background information (reference, location, number of samples,
species and fish length) and size descriptors (D50, dg, mg, sg,
and sk) for studies of redd gravels is presented in Table 1.
(Redd length, width, and area; and mean velocity and depth at the
redd site are presented in Kondolf 1988). The entries are
grouped by species; the species code used in these tables is
explained in Table 4. Comparable data for samples of potential
spawning gravels are presented in Table 2, and Table 3 presents

background data and size descriptors for fluvial gravels reported in the geologic literature.

Because most entries in Tables 1-3 represent average or composite values for a number of samples collected in a river or stream, it is probably not appropriate to compute grand means of these entries. However, one objective of this study was to examine general trends in these data. If the entries are assumed to reflect typical conditions at the sampled sites, it is appropriate to examine the spread of these entries as indicative of the range of spawning gravel sizes encountered in the field.

Size distributions for selected entries of trout and salmon redds are presented in Figures 1 and 2 using box-and-whisker plots modified after Tukey (1977). For each entry, the rectangle (box) encompasses the middle 50% of the distribution, from the D25 to D75 values, termed the "hinges". The median size, D50, is represented by a horizontal line through the box. Above and below the box are lines (whiskers) extending to the D90 and D10 values. Unlike conventional box-and-whisker plots, the ordinate in Figures 1 and 2 utilizes a logarithmic scale, reflecting the similarity between these distributions and the lognormal distribution. Also, in standard box-and-whisker plots, the whiskers extend to the extreme values, or extreme values are plotted and identified and the whisker extends to the innermost identified value. Because of the wide range of grain sizes in natural gravels and the poor definition of tails in many studies, it would be impossible to extend the whiskers to the absolute extremes. Even if all distributions were completely defined, it can be imagined that there is some small amount of clay present in almost any gravel, and it would be impractical to extend whiskers into the clay range.

These plots provide an easily comprehended depiction of the range and central tendency of sediment size distributions and permit the reader to compare multiple samples graphically. This mode of presentation solves a problem encountered when one attempts to present multiple cumulative size distribution curves in the conventional format, namely that the lines overlap and become indistinguishable unless the distributions are all very different.

General Properties of Spawning Gravel Size Distributions

Taken as a whole, median diameters for all 135 spawning gravel entries in these tables have a median value of 22 mm. The range is quite large, from 5.4 mm for coho salmon redds in Flynn Creek, Oregon (Koski 1966), to 78 mm for potential chinook salmon spawning gravels in the Columbia River near Vantage, Washington, a site now inundated by Priest Rapids Dam (Chambers et al. 1954). This range is even greater if the 0.1 mm value from unspawned,

silty substrates adjacent to chum salmon redds in sloughs of the
Susitna River, Alaska (entry 131), are included. As documented
by Vining et al. (1985), chum salmon utilize sites of upwelling
groundwater in the side sloughs of the river, often excavating 30
cm of this fine sediment before encountering gravel in which to
deposit their eggs. Thus, the 0.1 mm value is an outlier that
accurately represents the surficial substrate where chum salmon
excavate redds but not the particle size in egg pockets.

Of the median grain sizes presented for entries in Tables 1-3,
50 percent lie between 14.5 and 35 mm. Similarly, half of the
entries for graphic mean diameter fall between 10 and 27 mm, for
geometric mean, between 8.7 to 24 mm (Table 5). Values of
geometric standard deviation range from 1.6 in chinook salmon
spawning gravels of the Yuba River, California, to 21 in poorly-
sorted chum salmon spawning gravels of Susitna River Slough 10.
However, half the entries lie between 3.3 and 5.2, with a median
sg of 4.0 mm (Table 5). The vertical spread of the box-and-
whisker plots (Figures 1 and 2) provides an indication of the
sorting of the plotted entry. (The spread of these box-and-
whisker plots is from the 10th to 90th percentiles, whereas the
sorting index using the 16th and 84th percentiles, so the two
are not ex comparable.)

Nearly all of the average spawning gravel size distributions in
Tables 1-2 are negatively skewed, with half of the skewness
coefficients falling between -0.24 and -0.39 (Table 5). Only 3
entries did not show any negative skewness: chinook spawning
gravels in Benjamin Creek, Alaska (sk = 0.06), pink salmon
spawning gravels in Little Creek, Alaska (sk = 0.0), and the silt
deposits of the Susitna River sloughs (sk = 0.10). Negative
skewness can be detected in the box-and-whisker plots by longer
whiskers below the box, reflecting extended fine sediment tails.
The negative skewness is characteristic of these distributions on
a log-transformed basis; if the distributions were considered
without transformation, they would be positively skewed.

Differences in gravel sizes used by different species are
indicated in Table 6, which presents the spread (median, lower
and upper extremes) of entries by species. For median grain
size, the median value of entries ranges from 8.5 mm for brook
trout to 34.5 mm for chinook salmon. For geometric mean, median
value of entries ranges from 4.2 mm for brook trout to 24.5 mm
for chinook salmon. These interspecific differences are also
visible in the box-and-whisker plots presented in Figures 1 and
2.

Comparison of Spawning Gravels with Fluvial Gravels in the
Geologic Literature

Because of the intense interest in spawning gravels, it is of

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some interest to ask whether spawning gravels are distinctive
from fluvial gravels studied by geologists and engineers.
Results of this study indicate that spawning gravels tend to be
coarser than most gravels reported by geologists, but no
different in sorting and skewness. When the 135 entries for
spawning gravels are considered, the median values of D50, mg,
and dg are all higher than median values for these descriptors
drawn from the eleven entries for fluvial ("other") gravels drawn
from the geologic literature (Table 5). Median values of sg and
sk, however, are virtually identical for the two populations.
(Differences in extreme values of sg and sk can be attributed to
the different number of entries of the groups.)

The size distributions presented by Morris and Johnson (1967)
as "typical" for "water-laid gravel" (entries 143-145) are
smaller still than the average size of spawning gravels, but are
less negatively skewed (-0.02, 0.08, and 0.02) than the other
fluvial gravels or the spawning gravels.

Thus, it appears that the natural gravels used by spawning fish
are not distinctive from other fluvial gravels in sorting or
skewness, but they tend to be coarser than many of the gravels
that have been sampled by geologists.

Modality of Gravels

Pettijohn (1975) reviewed the studies of Conkling et al.
(1934), Krumbein (1940,1942), and Plumley (1948), and concluded
that most gravels are bimodal, distinct from sands, which tend to
be unimodal. The cumulative frequency distributions of bimodal
gravels should have a steep fine sediment tail, a flatter middle
portion, and another steep leg in the gravel range.
Distributions that define a smoothly concave-upward curve are not
bimodal. Inspection of numerous cumulative frequency curves by
Kondolf (1988) indicated that most were not obviously bimodal.
The curves that best displayed bimodality were the fluvial
gravels cited by Pettijohn (1975), although the secondary mode
(in the sand range) is often not pronounced. Most spawning
gravel curves are not bimodal. Many have extended fine tails,
but there is no indication of a secondary peak in the sand range
in most. Less than one fourth of the size curves examined by
Kondolf (1988) could be considered bimodal.

Thus, bimodal curves occur, but are not the rule, among
spawning gravels. Many more of the fluvial gravels reported in
the geologic literature appear to be bimodal, but, again, this
characteristic is not universal.

Influence of Mean Size on Other Properties

Church et al. (1987, p.51-52) noted that "mixtures of clastic
material frequently display variance proportional to the mean

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size, which is not surprising since all sizes may be present up to some cutoff point." To det ermine if sorting and skewness were related to grain size, the values of these descriptors in Tables 1-3 wer plotted as functions of graphic mean diameter. Values of sg have a smaller spread and tend to be lower for larger gravels (Figure 3), indicating the sorting is less variable and somewhat better for larger gravels. (The outlier is the value for silt deposits used by chum salmon in the Susitna River sloughs.) Skewness and graphic mean (of the entries in Tables 1-3) appear to be unrelated (Figure 4).

An inverse relationship (albeit weak) between sg and mean size is inconsistent with the finding of Church et al. (1987) that standard deviation increased with mean grain size in 78 surface samples from one river. The data set used here is not really comparable to that of Church et al. (1987) because it comprises samples from a wide range of environments, which can be expected to differ in many respects besides mean grain size. If the only independent variable were framework grain size, then sg (which is defined as geometric standard deviation) could be expected to increase with increasing grain size. The (weak) opposite trend may reflect the influence of fine sediment content on both the graphic mean and sorting: if framework gravel size is held constant, higher percentages of matrix sediment should decrease mean grain size and increase the standard deviation. Such a relation is suggested for the entries in Tables 1-3 by the relation between sorting index and D16 shown in Figure 5. The sorting index shows no relation to the value of D84 (Figure 6), indicating that the variations in sorting index are controlled primarily by the amount of fine sediment rather than the presence of large particles. In large part, this could be due to the practice of many investigators to exclude large particles, variously defined as particles larger than 51 to 152 mm, from the sample (e.g., Helle 1970, Chambers et al. 1954,1955).

The dependence of sorting and skewness on mean size was also examined for individual size distributions by Kondolf (1988) and found to be weak or absent.

The observed relationships between sorting and skewness, and mean size can be summarized as follows: larger gravels may be better sorted and more negatively skewed, but the relation is weak.

Influence of Fish Size on Redd Characteristics

Numerous conversations with fishery biologists have indicated a widespread recognition that gravel size, water depth and velocity, and redd size vary with the size of the spawning fish. Despite its widespread recognition, little has been published that describes this relation, and many studies of spawning

preferences do not include information on size of the spawning fish. Differences in gravel size used by various species have been described and attributed (explicitly or implicitly) to differences in size of the species (e.g., Burner 1951; Chambers et al. 1954,1955). Hunter (1973, p.26) recommended that the size of the spawning fish be recorded whenever spawning habitat investigations were conducted, noting that, "Interspecies spawning preferences for salmonids of the same size will be closer than intraspecies requirements for fish of varying size." Van den Berghe and Gross (1984) measured depth of egg burial in a population of coho salmon and found that depth of burial was strongly correlated with size of the spawning female.

A relation between fish size and gravel size is intuitively satisfying because larger fish are capable of moving larger gravels by virtue of the greater upward pull they can exert on the stream bed, and by virtue of their ability to maintain themselves in deeper and swifter flow, where the greater force of the current assists in entrainment of particles. However, while a large chinook salmon may be capable of spawning in a steep, swift stream with a coarse gravel bed, she is not obligated to do so. She may just as easily utilize smaller gravels in lower gradient streams. Thus, the ability of larger fish to utilize larger gravels can be viewed as determining an envelope curve rather than a linear relation between fish size and gravel size. The actual gravel sizes used by fish will be determined in large measure by the sizes available in the channel at any given site. The same principle should apply to other redd characteristics such as redd size, and the depth and velocity at the redd site. Vronskiy (1972) found no relation between the size of individual chinook salmon and the size of their redds in the Kamchatka River.

The data in Tables 1 and 2 were used to construct Figures 7 and 8, in which gravel size is plotted as a function of size of the spawning fish. The maximum gravel size and the range of gravel sizes used for spawning increase with fish size, but larger fish use smaller gravels as well as larger gravels. Envelope curves have been drawn through the maximum gravel sizes utilized by different size fish, but excluding some outlying points. The envelope curve in Figure 7 shows that 30 cm fish can generally use gravels up to about 30 mm in median size, while 90 cm fish can use gravels up to 80 mm in median size. Similarly, the envelope curve in Figure 8 shows that 30 cm fish can use gravels up to about 28 mm in graphic mean diameter, while 90 cm fish can use gravels about 50 mm in graphic mean.

Kondolf (1988) compiled data on redd area and velocities and depths at the redd sites. In general, larger fish construct larger redds, but the relation is poorly defined. Fish 30 cm

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long can spawn in water velocities of up to 0.5 m/s and depths as great as 0.25 m, while 100 cm fish can spawn in velocities up to 1.25 m/s and depths as great as 2.0 m.

These results indicate that the maximum values of the redd attributes of gravel size, depth, and velocity are related to fish size. However, confounding factors exist. These factors include variability in the sizes of fish in a spawning run, differences in the availability of different sizes of gravel from site to site, and the presence of upwelling or downwelling currents that may make a given site attractive to spawners.

There can be wide variations in fish size within a given spawning population, as illustrated in tributaries to the Owens River, California, where a single stream may have spawning brown trout ranging in size from 20 to 45 cm (D. Wong, Calif. Dept. Fish and Game, Bishop, personal communication 1987). If the gravel sizes available for spawning in such a stream were limited, it would mean that fish of very different size were utilizing gravel of the same size. The size of spawning fish in a given stream may fluctuate from year to year reflecting variations in the age structure of the spawning population, or there may be a long term secular change in the age structure of the spawning run. For example, in the Upper Sacramento River system, the spawning run of chinook salmon had a ratio of ocean age 3:4 year fish of about 50:50 in the early 1960's, when the spawning gravel samples of Van Woert and Smith (1962) were collected. However, increased harvesting of the run by commercial fishing interests offshore has resulted in a change in the composition of the run, so that it now has a ratio of about 65:35. The ocean age 3 year fish average 73 cm in size, while the ocean age 4 year fish average 95 cm in size, so the average size of spawning chinook salmon in the upper Sacramento system has decreased (Fred Meyer and John Hayes, Calif. Dept. Fish and Game, Sacramento, personal communication 1987). Thus, the size of spawning fish in a given stream may vary substantially over time.

It is now widely recognized that utilization of habitat components is influenced by their availability (Baldrige and Amos, 1981), and the size of gravels available to spawning fish can vary substantially from site to site. The range of gravels sizes available in a stream is a function of geomorphic variables such a climate, drainage area, basin lithology, structure, and of the local channel slope that results from the interaction of the other variables. The gravel sizes selected by one species in one region may show a wide variation because of gravel availability. For example, individual samples of rainbow trout spawning gravels in the Colorado River and tributaries ranged widely in size, from 7 to 48 mm in graphic mean diameter, but showed no

systematic increase in gravel size with fish lewngth. The
smallest fish occurred in small tributary streams with relatively
limited habitat and low nutrient levels; these small streams are
also steep and have coarse bed material. In contrast, the
largest fish occurred in the mainstem where the gradient is lower
and bars are composed of finer gravel (Kondolf et al. 1989).

Fish select spawning sites based not only on gravel size, but
on other factors as well, such as water depth and velocity,
cover, and the presence of upwelling or downwelling currents.
Higher velocities would tend to be associated with coarser
gravels, but the other variables should, for the most part, be
independet of substrate size. Thus, a fish may select a
spawning ite with gravels much finer than the maximum size she
can move as indicated on the envelope curve in Figures 7 and 8.
For examle, chum salmon in sloughs of the Susitna River select
sites baed primarily on the presence of upwelling currents, and
spawn irextremely fine-grained substrates if upwelling currents
are prent (Vining et al. 1985).

In sumary, there are physical reasons to expect that larger
fish ca use larger gravels for spawning, and such a pattern is
widely ecognized by fisheries biologists. The data presented
here sw that the maximum size utilized by a given size fish is
relate to fish length, but fish commonly spawn in gravels
smalle than they are capable of using. Thus, the relation
betwee fish size and gravel size is an envelope curve defining
the mimum size movable as a function of fish size. Similarly,
water epth and velocity are related to fish size by an envelope
curve Kondolf 1988).

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Table 1

Fish

Entry[1]	Reference	Species[2]	Location	Length (cm)	n	D50	mg	dg	ig	sk
1	Hitzel 1980	BK	Sheldon Ck, Ontario	18	32	8.9	5.7	4.6	5.5	-0.39
2		BK	Skunk Ck, Ontario	18	6	8.2	4.9	3.8	5.7	-0.44
3		BK	Galt Ck, Ontario	18	3	7.2	4.8	3.9	4.7	-0.4
4		BK	Conger's Ck, Ontario	18	16	10.7	6.3	4.8	5.5	-0.47
5	Harner 1963	RS	Cross Lake Thoroughfare, ME	49	23	16.5	9.5	7.2	4.8	-0.52
6		RS	Long Lake Thoroughfare, ME	53	10	15.0	9.1	7.0	4.7	-0.49
7	Hobbs 1937	BT	Various rivers, New Zealand	43	5	18.5	16.0	14.8	3.5	-0.17
8	Reiser and Wesche	BT	Douglas Ck abv Cheyn Dv, WY	31	20	24.0	18.0	15.5	4.4	-0.29
9	1977	BT	Douglas Ck bel Cheyn Dv, WY	31	53	28.0	22.1	19.7	2.7	-0.36
10		BT	Lake Ck, WY	31	16	11.0	9.0	8.1	5.8	-0.17
11		BT	Pioneer Canal, WY	31	6	17.0	11.6	9.6	3.5	-0.45
12		BT	Laramie R at EP site, WY	31	60	32.0	15.8	11.1	7.9	-0.51
13		BT	Laramie R bel Pionr Cnl, WY	31	7	18.0	12.4	10.2	4.1	-0.40
14		BT	Hog Park Ck, WY	31	8	17.0	11.4	9.3	5.2	-0.37
15	Hitzel 1980	BT	Skunk Ck, Ontario	32	4	9.2	6.0	4.8	5.0	-0.4
16		BT	Galt Ck, Ontario	32	2	5.8	4.8	4.4	5.6	-0.16
17		BT	Beatty Saugeen R, Ontario	32	5	14.5	10.3	8.7	4.2	-0.36
18		BT	Syndenham R (1), Ontario	32	6	8.4	6.1	5.2	5.2	-0.29
19		BT	Syndenham R (2), Ontario	32	30	9.9	6.6	5.4	5.0	-0.38
20	Carling (unpub)	BT	Carl Beck, England	29	22	50.0	28.0	21.0	5.2	-0.52
21		BT	Eggleshope Beck, England	29	35	19.0	12.6	10.3	8.6	-0.29
22	Maddux et al 1987	RBT	Colorado R, AZ	45	10	10.5	6.9	5.6	5.3	-0.37
23	This study	RBT	Colorado R tributaries, AZ	40	10	32.0	26.7	24.3	2.9	-0.26
24	Orcutt et al	SH	N.Fk Clearwater R tribs, ID	76	60	42.0	30.3	25.7	4.3	-0.34
25	1968	SH	Salmon R tribs, ID	75	8	46.0	25.6	19.1	5.8	-0.50
26	Chamberstal 1954-5	SH	Kalama R, WA	81	3	31.0	25.8	23.5	3.9	-0.20
27	Hobbs 1937	CH	Hinding R, New Zealand	81	2	16.5	15.6	15.2	3.3	-0.07
28	Burger et al.	CH	Kenai R, AK	101	4	31.8	29.0	27.8	2.7	-0.14
29	1983	CH	Benjamin Ck, AK	94	4	22.0	22.8	23.2	2.7	0.06
30	Vronskiy 1972	CH	Kamchatka R, mainstem	90	4	47.0	34.9	30.0	3.7	-0.34
31		CH	Siberia: arm 1	90	2	26.0	23.3	22.0	3.9	-0.12
32		CH	arm 2	90	2	16.0	13.7	12.7	3.8	-0.17
33	This study	CH	Crooked Ck, AK	90	4	36.0	29.7	27.0	2.3	-0.34
34		CH	Yuba R, CA	81	1	34.0	28.3	25.8	2.2	-0.34
35	Chambers et al.	CH	Kalama R, WA	86	13	54.0	43.8	39.5	3.0	-0.28
36	1954,1955	CH	Snake R, ID	86	8	21.0	18.4	17.2	3.4	-0.16
37		CH	Cispus R, WA	82	7	50.0	39.5	35.1	3.2	-0.30
38		CH	Imnaha R, OR	82	4	41.0	36.8	34.8	2.9	-0.15
39		CH	American R, WA	82	5	35.0	28.4	25.6	3.0	-0.28
40		CH	Cowlitz R, WA	82	8	51.0	35.0	29.0	3.9	-0.42
41		CO	Spring Ck, WA	65	4	35.0	24.4	20.3	3.4	-0.44
42		CO	Toutle R, WA	65	4	16.5	15.6	15.2	3.2	-0.07
43		CO	Burns Ck, WA (1953)	65	7	29.0	23.4	21.0	3.3	-0.27
44		CO	Burns Ck, WA (1954)	65	4	33.0	25.3	22.1	3.1	-0.36
45	Koski 1966	CO	Deer Ck, OR	67	nr	12.0	5.4	3.6	10.6	-0.51
46		CO	Needle Branch, OR	67	nr	6.3	3.6	2.7	9.6	-0.37
47		CO	Flynn Ck, OR	67	nr	5.4	3.4	2.7	9.1	-0.31
48	Helle 1970	PN	Olsen Ck, upper intdl	43	12	8.8	8.0	7.6	4.2	-0.10
49		PN	AK, low grad	43	10	11.0	8.0	6.8	5.1	-0.29
50		PN	mid intdl	43	43	10.0	8.2	7.5	4.7	-0.19
51		PN	lower intdl	43	29	8.0	6.3	5.6	5.3	-0.21
52	Chambers et al	SO	Okanagan R, BC	65	12	25.0	20.1	18.0	3.3	-0.27
53	1954,1955	SO	Little Wenatchee R, WA	50	4	17.8	14.7	13.4	2.7	-0.29
54	This study	SO	Quartz Ck, AK	35	3	19.0	18.6	18.4	3.5	-0.03

1 = Most entries are averages of more than one sample. as indicated by n.

2 See Table 4 for explanation of species code

Table 2

Fish

Entry	Reference	Species	Location	Length (cm)	c	D50	mg	dg	sg	sk
55	Spoon 1985	BT	Missouri R, MT	50	11	22.5	18.9	17.4	3.0	-0.24
56		BT	Beaver Ck, MT	29	15	13.0	9.0	7.4	6.2	-0.31
57	This study	BT	Owens R tributaries, CA	23	12	18.0	14.7	13.3	3.2	-0.26
58	Hartman &Galb 1970	RBT	Lardeau R, BC	75	6	23.5	17.2	14.7	3.6	-0.37
59	Platts et al. 1979	RBT	N Fk Boise R, ID	30	45	20.0	14.5	12.4	6.5	-0.25
60	Spoon 1985	RCT	Missouri R, MT	44	27	12.5	9.5	8.3	4.6	-0.27
61		RBT	Beaver Ck, MT	44	19	15.0	10.9	9.3	4.9	-0.30
62	Maddux et al 1987	RBT	Colorado R, AZ	45	1	16.0	7.6	5.2	10.5	-0.47
63	This study	RBT	Colorado R tributaries, AZ	40	8	21.9	17.4	15.5	3.8	-0.26
64		RBT	Nantahala R, NC (<90mm)	21	14	24.5	18.8	16.5	3.8	-0.30
65		RBT	Nantahala R, NC (all rx)	21	14	46.3	32.0	26.6	4.0	-0.40
66	Chambers&al 1954-5	SH	Kalama R, WA	75	2	40.0	31.6	28.1	4.3	-0.24
67	Bjornn unpub	SH	Tucannon R, OR: mouth	66	2	12.4	9.4	8.2	2.9	-0.38
68		SH	upstream	66	4	25.4	18.8	16.2	3.7	-0.34
69	Cederhole and	SH	Stequaleho Ck, WA: site 1	65	34	19.5	13.2	10.9	5.4	-0.35
70	Salo 1979	SH	site 2	65	43	22.0	14.5	11.7	5.1	-0.38
71		SH	site 3	65	38	22.0	14.7	12.0	4.8	-0.38
72		SH	Clearwater R, WA: site 1	70	27	10.4	9.6	9.3	4.8	-0.07
73		SH	site 2	70	22	13.5	10.5	9.3	4.8	-0.24
74		SH	site 3	70	25	18.0	11.8	9.6	5.6	-0.36
75		SH	site 4	70	39	19.0	12.2	9.8	6.5	-0.35
76		SH	site 5	70	73	23.0	14.3	11.2	5.3	-0.43
77		SH	site 6	70	61	15.0	12.0	10.8	4.0	-0.24
78		SH	site 7	70	17	22.0	10.2	6.9	8.4	-0.54
79	Shirazi et al 1981	SH	Beaver Ck, OR	68	3	26.5	20.4	17.9	3.3	-0.33
80		SH	Three Rivers, OR	68	3	32.4	23.2	19.6	3.8	-0.38
81		SH	Gopher Ck, OR	68	3	33.7	26.7	23.8	2.7	-0.35
82		SH	Rock Ck, OR	68	3	36.7	32.9	31.1	2.2	-0.17
83	Chambers et al.	CH	Columbia R, WA	86	4	78.0	51.2	41.4	3.2	-0.55
84	1954,1955	CH	Snake R, ID	86	10	21.0	16.4	14.5	4.1	-0.26
85		CH	Kalama R, WA	86	7	49.0	35.8	30.6	4.0	-0.34
86		CH	Cowlitz R, WA	82	1	42.0	30.1	25.5	4.9	-0.31
87		CH	Imnaha R, OR	82	3	52.0	37.4	31.7	3.8	-0.37
88		CH	Cispus R, WA	82	10	37.0	27.1	23.2	4.7	-0.30
89		CH	American R, WA	82	8	34.0	27.2	24.3	3.2	-0.28
90	Van Hoert and	CH	Sacramento R, CA	84	3	44.0	29.9	24.7	4.7	-0.38
91	Smith 1962	CH	Cottonwood Ck, CA	84	12	31.0	22.1	18.6	4.8	-0.33
92		CH	Cow Ck, CA	84	5	52.0	37.7	32.1	3.2	-0.41
93		CH	Battle Ck, CA	84	20	66.0	44.8	36.9	3.5	-0.46
94	Platts et al	CH	S Fk Salmon R,ID: Stolle Me	86	145	22.0	11.7	8.5	7.1	-0.49
95	1979	CH	S Fk Salmon R,ID: Poverty R	86	310	11.2	8.0	6.8	7.5	-0.25
96		CH	S Fk Salmon R,ID: Glory Are	86	80	16.5	10.8	8.7	6.4	-0.34
97		CH	Johnson Ck,ID	86	100	24.5	14.9	11.6	5.8	-0.43
98		CH	Bear Valley Ck,ID	86	20	10.8	8.0	6.9	6.0	-0.25
99		CH	Elk Ck,ID	86	20	15.2	10.7	9.0	3.9	-0.39
100		CH	Loon Ck,ID	86	20	21.5	14.4	11.7	5.1	-0.37
101		CH	Salmon R, ID: Lower Decker	86	5	27.0	18.5	15.3	5.1	-0.35
102		CH	Salmon R, ID: Upper Decker	86	5	13.2	11.4	10.6	4.6	-0.14
103		CH	Alturas Ck, ID	86	20	14.5	11.9	10.7	4.3	-0.21
104	Shirazi et al 1981	CH	Grant Ck, OR	82	4	30.0	22.8	19.9	3.6	-0.32
105		CH	Rogue R, OR: Old Bridge	82	3	37.8	35.9	35.1	2.1	-0.1
106		CH	Rogue R, OR: Hatchery	82	3	39.7	32.2	29.9	4.4	-0.2
107		CH	Rogue R, OR: Sand Hole	82	1	69.3	64.6	62.4	1.6	-0.3
108		CH	Rogue R, OR: Dam Site	82	3	59.0	42.0	35.5	2.4	-0.59
109		CH	Rogue R, OR: Big Butte Ck	82	3	35.0	25.5	21.7	3.3	-0.40
110	Chapman et al 1983	CH	Columbia R (Vernita), WA	86	4	43.0	37.6	35.1	2.9	-0.19
111	This study	CH	Crooked Ck, CA	90	1	41.3	30.3	25.9	2.3	-0.58
112		CH	Yuba R, CA	81	1	35.0	28.1	25.1	2.2	-0.42
113	Chambers et al.	CO	Spring Ck, WA	65	2	13.0	11.2	10.4	1.5	-0.18
114	1954,1955	CO	Toutle R, WA (1953)	65	2	10.0	9.2	8.8	3.7	-0.10
115		CO	Burns Ck, WA (1954)	65	1	29.0	25.5	24.0	2.9	-0.18
116		CO	Burns Ck, WA (1954)	65	1	33.0	27.6	25.3	2.5	-0.29
117	Helle 1970	PN	Olsen Ck, AK: upper intdl	43	22	9.2	7.5	6.8	5.4	-0.18
118		PN	Olsen Ck, AK: Middle Slough	43	25	6.5	5.2	4.6	5.0	-0.21
119		PN	Little Ck, AK	43	2	9.6	9.6	9.6	3.3	-.00
120	Shirazi et al 1981	CH	Porcupine Ck, AK: mainstem	65	2	9.6	7.5	6.6	1.9	-0.57

Table 2 (Cont)

Entry	Reference	Species	Location	Length (cm)	n	D50	mg	dg	sg	sk
125	Vining et al 1985	CM	Susitna R, AK: mainstem	65	2	62.0	36.6	28.1	5.5	-0.46
126		CM	Susitna R, AK: Side chl 10	65	4	21.8	10.6	7.4	9.7	-0.48
127		CM	Susitna R, AK: Side chl 21	65	5	36.4	22.8	18.1	6.6	-0.37
128		CM	Susitna R, AK: Slough 10	65	3	42.3	10.2	5.0	21.1	-0.70
129		CM	Susitna R, AK: Slough 11	65	6	20.5	16.9	15.3	4.9	-0.19
130		CM	Susitna R, AK: Slough 21	65	3	42.7	25.4	19.6	6.4	-0.42
131		CM	Susitna R, AK: silt	65	4	0.1	0.1	0.1	2.2	0.10
132		CM	Fourth of July Ck, AK	65	4	25.7	21.7	19.9	4.2	-0.18
133	Chambers et al.	SO	Okanagan R, BC	65	5	48.0	37.1	32.7	2.7	-0.38
134	1954,1955	SO	L. Wenatchee R, WA	50	4	14.5	11.2	9.9	3.5	-0.30
135	This study	SO	Quartz Ck, AK	35	3	22.9	18.5	16.6	3.3	-0.27

Table 3 Size Descriptors (mm) for Samples of Fluvial Gravels in the Geologic Literature *

Entry	Reference	Location	n	D50	mg	dg	sg	sk
136	Conkling 1934	alluvial fans, CA	33	28.3	20.1	17.6	3.8	-0.30
137	Krumbein 1940	San Gabriel Cyn, CA	15	9.35	6.4	5.3	6.4	-0.30
138	Krumbein 1942	Arroyo Seco, CA	20	4.98	4.2	3.8	6.8	-0.14
139	Plumley 1948	Rapid Ck terraces, SD	10	23.0	15.1	12.2	5.1	-0.39
140		Bear Butte Ck terraces, SD	7	29.6	22.3	19.3	3.9	-0.31
141		Battle Ck terraces, SD	6	21.3	12.9	10.0	7.1	-0.39
142	Schlee 1957	upland gravels, MO	72	5.2	3.0	2.2	11.2	-0.35
143	Morris and	Arapahoe Couty, CO	1	2.7	2.7	2.7	1.5	-0.02
144	Johnson 1967	Douglas County, CO	1	4.1	4.2	4.3	1.6	0.08
145		Arkansas R Valley, AR	1	8.3	8.3	8.4	1.5	0.02
146	Knott and Lipscomb 1983	Chulitna R, AK	10	20.2	18.8	18.3	3.1	-0.05

* All values are in mm except n, sg, and sk, which are dimensionless.

Table 4. Species code used in Tables 1-3

Code	Common Name	Scientific name
AS	Atlantic salmon	Salmo salar
BK	Brook trout	Salvelinus fontis
BT	Brown trout	Salmo trutta
CH	Chinook salmon	Oncorhynchus tstscha
CM	Chum salmon	O. keta
CO	Coho salmon	O. kisutch
PN	Pink salmon	O. gorbuschka
RBT	Rainbowtrout	O. mykiss
SH	Steelhead	O. mykiss
SO	Sockeye	O. nerka

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Table 5. Spread of Size Descriptors for Entries in Tables 1-3.*

Size Descriptor	Spawning Gravels					Other Gravels		
	Extr	P75	P50	P25	Extr	Extr	P50	Extr
D50	78	35	22	14.5	0.13	29.6	9.4	2.7
mg	64.6	27.2	16	10.2	0.1	22.3	8.3	2.7
dg	62.4	24	14.8	8.7	0.1	19.3	8.4	2.2
sg	21.1	5.2	4	3.3	1.6	11.2	3.9	1.5
sk	0.1	-0.24	-0.31	-0.39	-0.7	0.08	-0.3	-0.39

* All 135 entries for redd gravels and potential spawning gravels
are presented together under "spawning gravels." Extremes are
listed in columns labelled "Extr", and P75, P50, and P25 designate
first quartile, median, and third quartile, respectively.

Table 6. Spread of Entries in Tables by Species

Species	number of entries	Median Diameter (mm)			Geometric Mean Diameter (mm)		
		Extr	Median	Extr	Extr	Median	Extr
Brook trout	4	7.2	8.5	10.7	3.8	4.2	4.8
Atlantic salmon	2	15	15.8	16.5	7	7.1	7.3
Brown trout	18	5.8	17.5	50	4.4	9.9	21
Rainbow trout	10	10.5	21	46.3	5.2	12.4	26.7
Steelhead	20	10.4	22.5	46	6.9	11.9	31.1
Chinook salmon	42	10.8	34.5	78	6.8	24.5	62.4
Coho salmon	11	5.4	16.5	35	2.7	15.2	25.3
Chum salmon	13	0.13	25.7	62	0.1	15.3	31
Pink salmon	7	6.5	9.2	11	4.6	6.8	9.6
Sockeye salmon	6	14.5	22	48	9.9	17.3	32.7
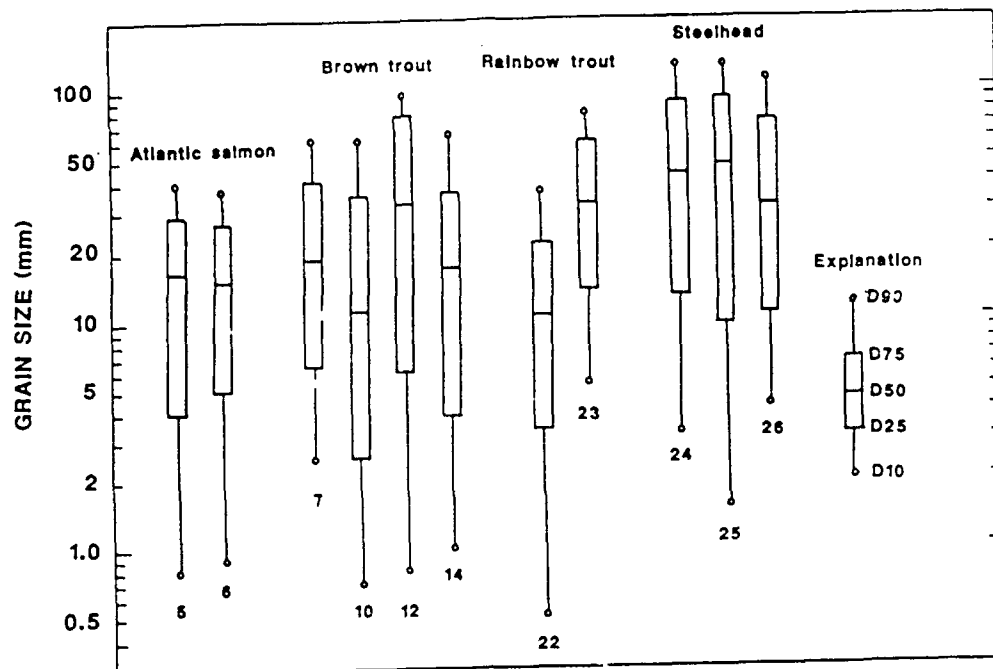
Figure 1. Box-and-whisker plots of size distributions of gravels from trout redds. Each plot is labeled with the number corresponding to its entry in Table 1 Box encompasses the middle 50 percent of the distribution, with the median indicated by horizontal line. Whiskers extend to D90 and D10.
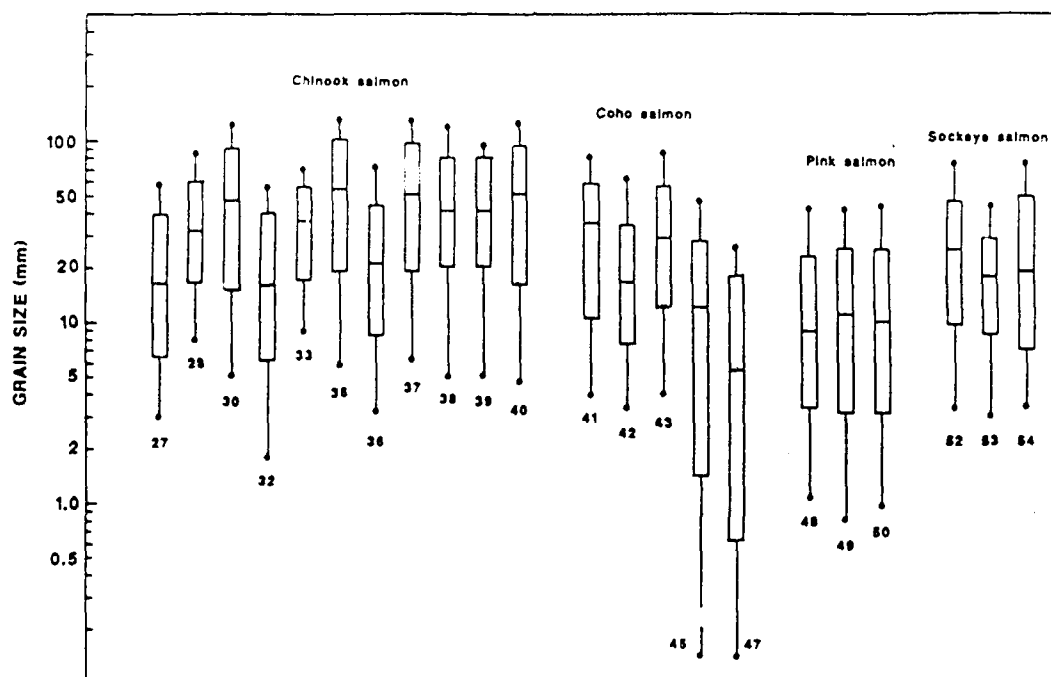
Figure 2. Box-and-whisker plots of size distributions of gravels from salmon redds. Each plot is labeled with the number corresponding to its entry in Table 1 Box encompasses the middle 50 percent of the distribution, with the median indicated by horizontal line. Whiskers extend to D90 and D10.

3
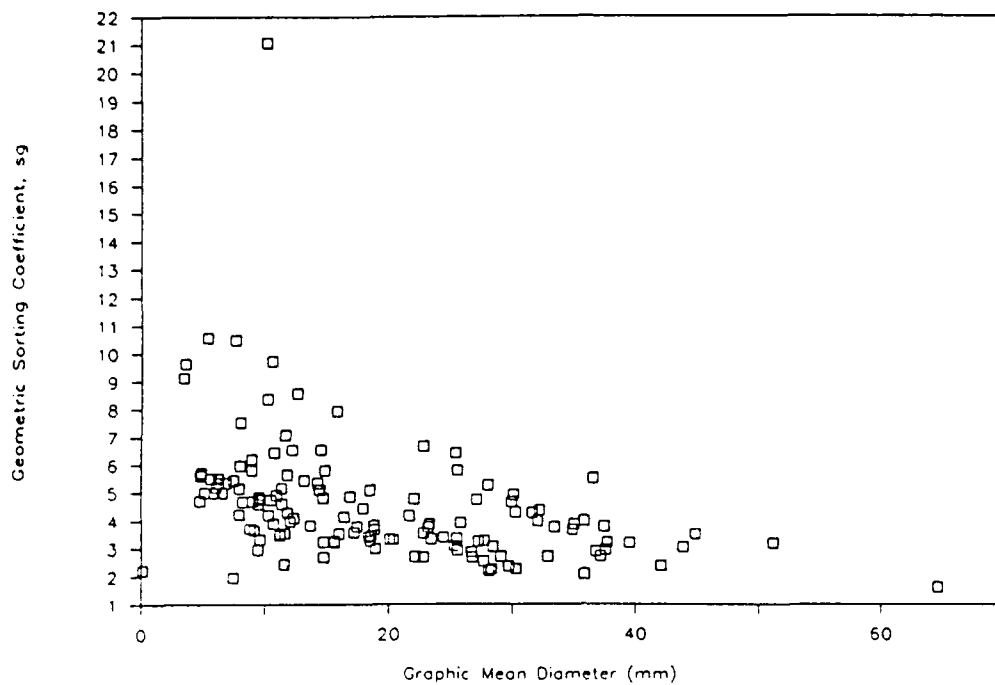
Figure ▮▮. Sorting coefficient vs graphic mean diameter for
entries in Tables ~~6.1 6.7~~. *1-3*.
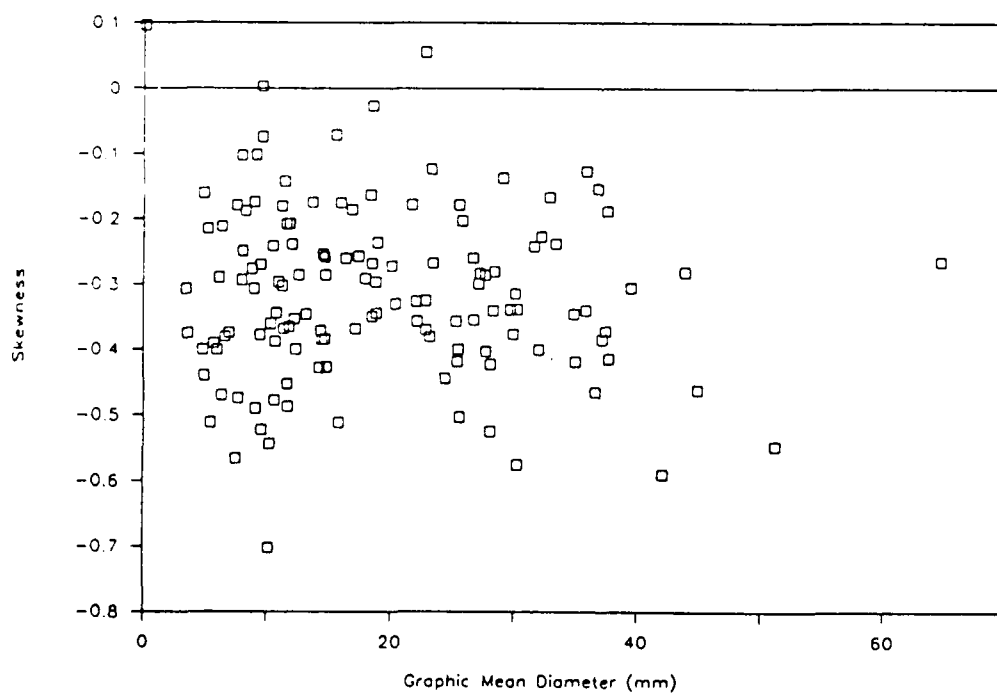
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Figure ▮2. Skewness vs graphic mean diameter for entries
in Tables ~~6.1 6.7~~. *1-3*.
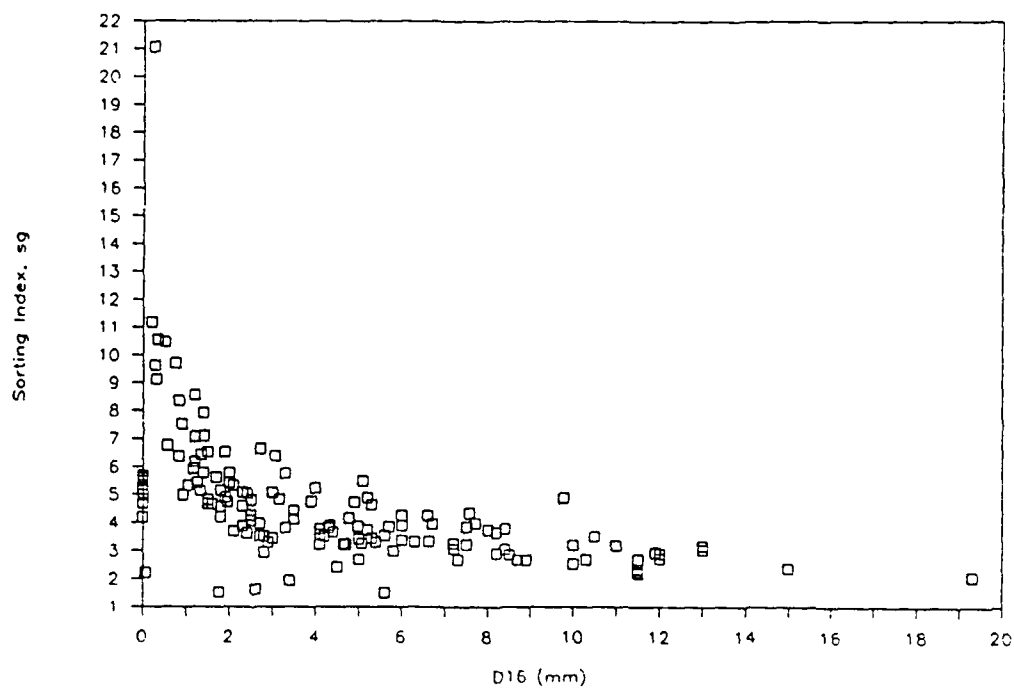
Figure 5. Sorting coefficient vs D16 for entries in Tables 1-3
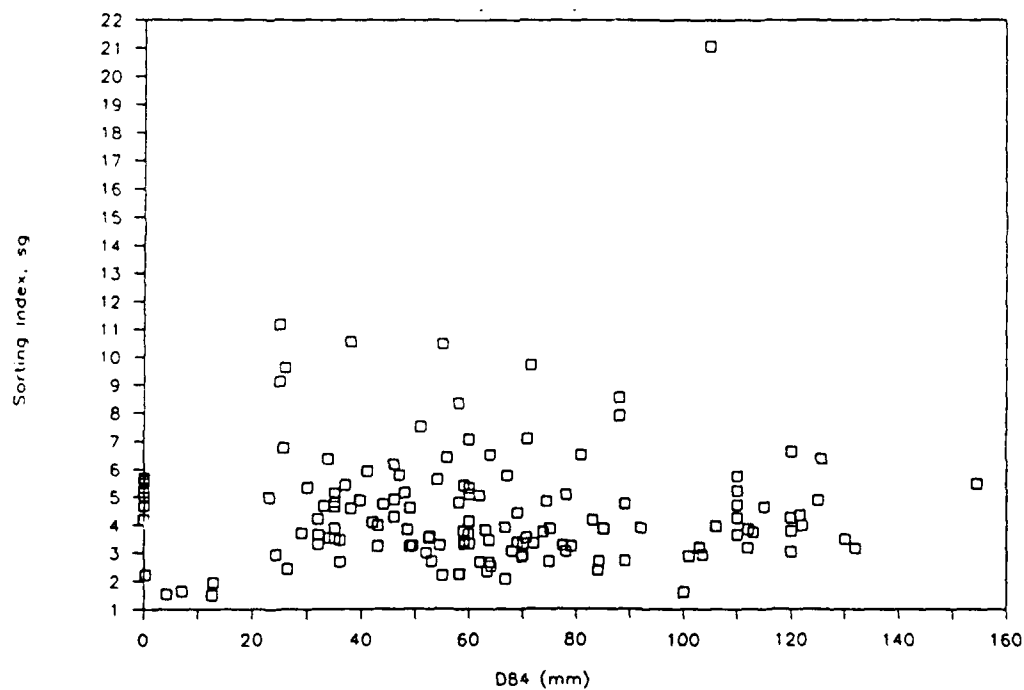
Figure 6. Sorting coefficient vs D84 for entries in Tables 1-3.
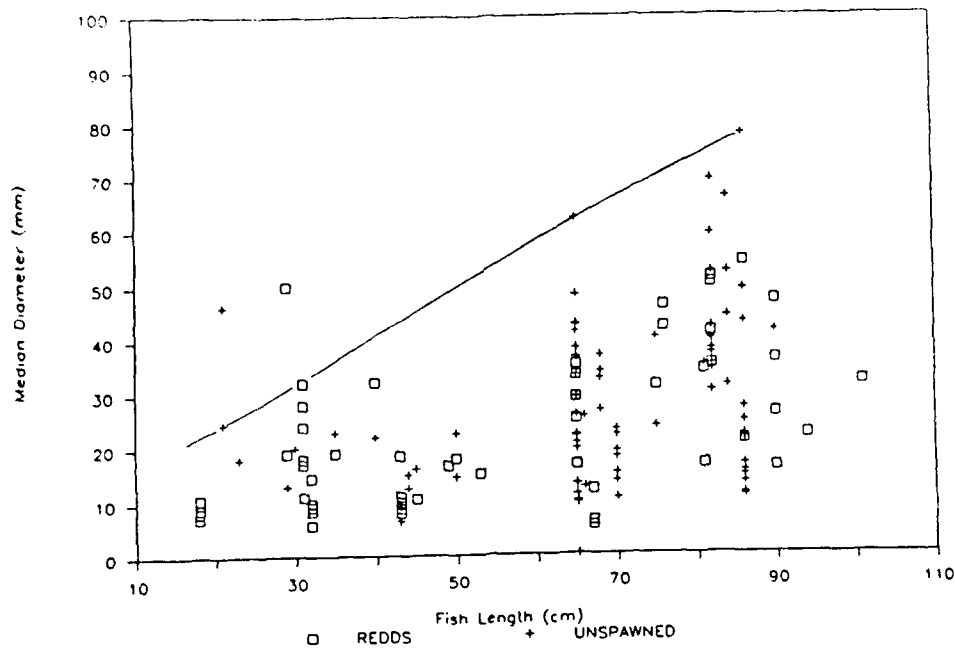
Figure 7**8**. Relation between fish size (total length) and median size of
spawning gravel for entries in Tables 1-**2**. The envelope curve defines the
probable upper limit of acceptable gravel size for most conditions, but excludes
two outlying points (entry 65 for rainbow trout in the Nantahala River, and entry
20 for brown trout in Carl Beck) so that the range of most entries is better
defined.
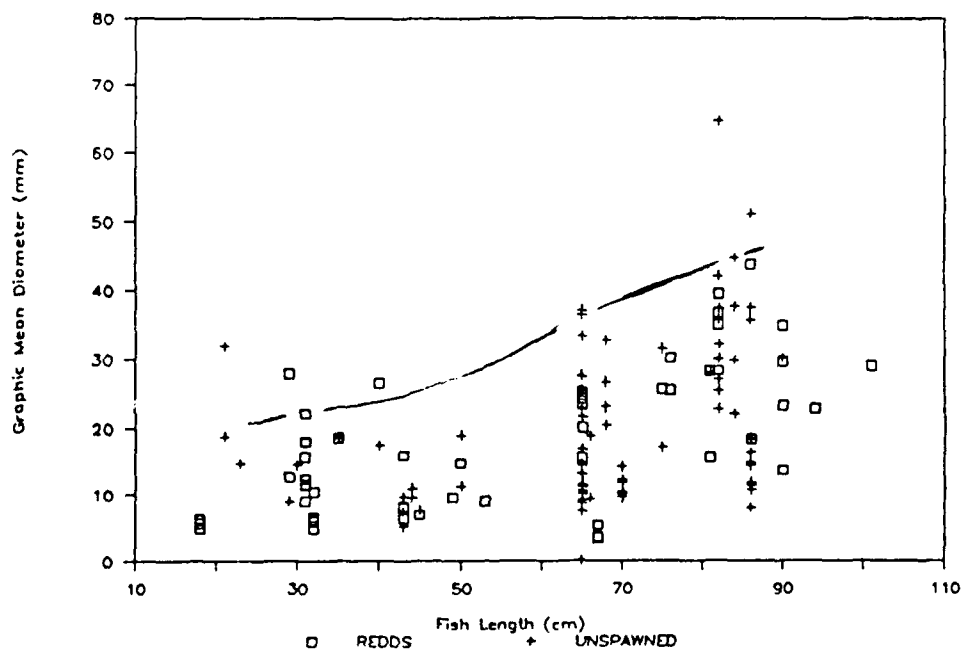
Figure **8**. Relation between fish size (total length) and graphic mean
diameter of spawning gravel for entries in Tables 1-**2**. The envelope curve
defines the probable upper limit of acceptable gravel size for most conditions,
but excludes entry 65 and entry 107 (for chinook salmon in the Rogue River) to
better define the range of most points.

LABORATORY OBSERVATIONS OF BED ARMOURING

AND CHANGES IN BED LOAD COMPOSITION

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Summary

A series of three laboratory experiments was carried out at a constant flow and initial bed slope. Continuous observations were made of the bed surface composition, of the bed load transport rate and of bed load composition. The bed load transport rate was found to diminish progressively after an initial period in which it varied slightly. Its size distribution was found to change in a broadly systematic fashion with time. Although observation of the bed surface was difficult, a tentative link was established between the characteristics of the bed load and of the bed surface.

1. **INTRODUCTION**

Most of the published sediment transport rate prediction methods have been developed using a "uniform" size of sediment. This was done so that the effect of sediment size on transport rates could be determined. However in the natural environment, river bed composition is rarely uniform. Non-uniform beds exhibit very different behaviour from that of simple sizebeds. If no sediment is introduced, the bed degrades and coarse material accumulates on the surface reducing the rate of transport significantly (Harrison, 1950). In some circumstances the transport rate is reduced to zero, and then the bed is said to be armoured.

Some of the single size prediction formulae have been amended to take account of mixed grain size (e.g. Bettes and White 1981). Empirical adjustments, determined from large sets of both laboratory and field data, have been proposed to reflect the interaction between grain sizes has occurred (Ranga Raju 1984). These methods frequently refer to a steady state equilibrium transport rate and are therefore unsuitable when considering the development of an armour layer as may occur during a flood.

It has been observed that the armour is a surface layer one grain diameter thick (Harrison 1950, Sutherland 1985). It is also generally accepted that all size fractions present in the original material are present in the armour layer.

It is clear that to understand the phenomenon of bed armouring detailed knowledge must be obtained of the processes which occur in the armour layer. These processes involve the entrainment of grains, sorting of size fractions and interaction between grain sizes. Observations of systematic changes in the composition of the armour layer could indicate the different processes present as the transport rate changes. Once this understanding has been acquired, reliable numerical models can be produced which will attempt to model actual physical processes and not just 'fit' known data sets. Therefore the ability to make observations of the changes in sediment composition of the armour layer is important. However there is considerable difficulty even in the laboratory in making sound observations of either the sorting of the armour layer or the composition of the bed load.

In this investigation periodic observations were made of both the bedload and the armouring bed. Observations of bedload were made using a sediment trap developed from a prototype used by Hardwick and Willetts (1989). The surface composition of the surface layer was studied by an "aerial" photography procedure (Bettes 1982). This paper reports experiments during which the bedload composition and transport rate and the bed surface layer were monitored during the development of a bed armour layer.

2. **EXPERIMENTAL WORK**
2.1 **Apparatus**

The experiments were carried out in a recirulating, tilting, glass-sided flume which was 12.5 metres long and 0.3 metres wide. The slope of the flume was adjusted to 0.001 for all the experiments. All water and bed level measurements were made relative to a datum plane, set by two parallel rails running along either side of the

flume. The distances from the datum plane to the bed and water surfaces were measured using a point depth guage with an accuracy of ± 0.1mm. The water depth was controlled by an adjustable sharp edged weir at the downstream end of the flume. This was adjusted so as to minimise the drawdown effects and give as large as possible a reach of flow at a uniform depth.

The bedload sediment trap consisted of an opening in the base of the flume, a valve and several interchangeable sediment collection boxes. (figs 1,2). The trap was installed 10.5 metres downstream from the flume inlet (Fig. 3). The opening in the base of the flume was 200mm x 9mm, the larger dimension being perpendicular to the direction of flow. These dimensions allowed the trap to collect the bedload across the whole width of the flume because even though the dimension of the opening in the flume base (200mm) is smaller than the overall width of the flume (300mm) the bed material lay around the trap at an angle of repose which funnelled the grains moving beside the flume walls down into the trap. The minimum opening dimension (9mm) restricted the maximum size of grain which could be trapped. This was accepted as 6mm and was the overriding limit as regards the maximum size fraction present in any of the bed mixtures used.

The bedload collection boxes were split into three sections laterally. Sediment was thus collected from three separate zones across the width of the flume, each 100mm wide. This was done after it was discovered that at the flow depth used, four secondary flow cells were set up dividing the flow into a central zone and two sidewall zones each 100mm wide.

The important feature of this bedload trap is that the collected bedload can be recovered easily and frequently without disruption to the flow or the gravel bed. This enables the amount and composition of the bedload to be monitored virtually continuously.

A camera was mounted on a carriage which moved along the rails. This was used to take photographs in plan view of the gravel bed surface at intervals of approximately 2 hours and at three points along the flume. Thus indicating changes in the bed surface were recorded at 3 metres, 6 metres and 9 metres from the flume inlet as the bed armour develop with time.

Flow was measured by integrating a velocity profile obtained at a particular cross-section using a Laser Doppler Anemometry system. Flow was held constant throughout each experiment by controlling the height of water above the downstream weir.

2.2 Procedure

Three experiments were carried out, keeping the hydraulic influences as constant as possible so that any reduction in sediment transport was attributable to bed sediment composition changes and not to changes in the flow environment. The object of each experiment was to obtain a stable armoured sediment bed by parallel degrading. Flow was monitored and any necessary valve adjustments made to keep it constant. Bed slope and water depth were also monitored in order to discover if they and the average bed shear stress calculated from these parameters remained sensibly constant throughout.

The base of the laboratory flume was covered to a depth of 100mm with a thoroughly mixed sediment mixture. The sediment was then slowly flooded for its whole depth and then drained to aid settlement. It was then scraped level using a scraper which ran on the datum rails. The excess material was discarded.

This gave a level sediment bed of a constant slope equal to 0.001. The surface contained all the sizes in the parent material. The only blemishes were a small number of drag marks, caused by dragging by the scraper of larger grains. However these were small in number and considered insignificant. The bed was then exposed to a low flow (below the estimated threshold of motion) to remove any unnaturally exposed grains arising from the bed-laying operations. The bed was then ready for the experiment to start. Photographs were taken and bed samples noted so that the initial state of the sediment bed was recorded.

A constant flow was now introduced and the weir adjusted to extend the uniform depth of flow. The flow in each of the three experiments was large enough to move the finest grains, but not to move every grain size. Selective erosion therefore occurred and the bed began to armour. Two experiments had a duration of 100 hours and the third of 50 hours. Experiments were stopped when either the sediment transport rate had declined to approximately 10% of its original value or 100 hours had elapsed.

During the experiments the flow was kept constant and the bedload, the surface of the bed and the flow parameters were sampled at intervals

2.2.1   Sampling Techniques

Flow Conditions

The water and the bed levels were recorded every 60mm across the width of the flume and at 1 metre intervals along the working length from 3 metres to 9 metres measured from the flume inlet - (fig. 3). The readings at each cross-section were averaged from seven measurements. From these measurements accepted values of bed depth, water depth and bed shear stress were calculated. The shear stress calculation employed a sidewall correction (Einstein 1942).

Bedload Sampling

Sediment transport rate measurements were made, every 30 minutes for approximately the first 6 hours. Then the time interval and the sampling period were increased gradually to hourly, then every 2 hours, every 4 hours and eventually before and after overnight runs. The elapsed time of the transport rate measurement was taken as the time to the middle of the sample period.

From the bedload recovered, the sediment transport rate, and the composition of the bedload were determined at intervals throughout the experiment. The bedload samples were dried, weighed and then sieved. All the sediment samples were analysed using a series of sieves at $\frac{1}{4}\phi$ intervals. A mechanical shaker with a shaking time of 10 minutes was used. Weights were determined to an accuracy of 0.01 grams.

Bed Armour Sampling

In order to record the development of the bed armour, non-destructive sampling was required at different positions along the flume and at different times. For this reason aerial photography was used. The information contained in these photographs enabled a bulk mass-size distribution of the armour to be determined.

At intervals throughout the experiment the flow was stopped, the water surface fell to a level dictated by the setting of the adjustable tail weir. Photographs were then taken with the camera over the centreline of the flume at three positions, 3m, 6m, and 9m downstream of the flume inlet. For analysis these photographs were enlarged to approximately four times the actual size — a square grid of one hundred points was superimposed on the photographs and the size of each stone under each point was measured. Sediment grains below 0.125mm in diameter could not be measured. The grain sizes were analysed on a size to number basis, this should give the same grading curve as a bulk mass-size distribution according to Hey (1983).

3. RESULTS AND DISCUSSION

Three experiments were concluded. It was intended to use the same mixture for each experiment. However the first mixture differed from the mixture used in the last two experiments because of unforeseen variations in a commercially supplied ingredient (Fig. 4). The experimental conditions are summarised in table 1.

Experiment No.	Flow L/s	Original Bedslope	Original Flow Depth mm	Average Shear Velocity m/s
1	7.5	0.001	65.7	0.0221
2	6.0	0.001	57.8	0.0205
3	9.0	0.001	72.8	0.0247

Table 1 Summary of experimental conditions

3.1 Bedload Results

Sediment Transport Rate

The rate of bedload transport decreased with time, (fig. 5,6,7) as it had in the previous work of Harrison (1950) and Proffitt (1980). For an initial period the sediment transport rate remained approximately constant or declined very gradually. The initial value of sediment transport rate must be interpreted with caution as the area around the trap stabilised, resulting in some initial disruption of the natural flow of sediment. There then occurred a period of rapid decline in sediment transport, clearly visible in fig. 5,6,7.

Comparison of fig. 6 and 7 suggest that the duration of the nearly constant transport rate is a function of bed shear stress (or shear velocity). These two graphs represent experiments with almost identical initial grain mixtures, yet in experiment 3 the onset of rapid decline of transport rate occurs much sooner than in experiment 2. It is clear that this factor also depends on the composition of the original bed. The duration of nearly constant transport rate observed in experiment 1 was much larger than in experiments 2 and 3. It was concluded that the longer fine tails of the size distribution of bed material in experiment 1 (fig. 4) resulted in more grains being available for transport, thus prolonging the zone of small change in the sediment transport rate.

Bedload Mass-size Distributions

The size distribution of the bedload also changed systematically with time. It was decided to fit hyperbolic curves to logarithmic plots of the sieve data (Barndorff-Neilsen 1980). This method was used as it can more adequately describe the distribution of grain sizes than the traditional log- normal distribution.

Figures 8,9,10,11 show the size distribution at various sampling periods during experiment 3 and fits a hyperbola to each set of points. It has been found that the selective transport which causes bed armouring involves a systematic change in the bedload which is shown by the changing characteristics of the hyperbolic distribution (Barndorff-Neilson, Christiansen 1986).

As the sediment transport rate declined it was noted that the modal grain size of the transported material changed dramatically (fig. 12). This plot indicates that the modal grain size of the bed load remained reasonably constant, fluctuating about a value of approximately 0.75mm. After about 600 minutes the modal grain size started to decrease until it reached a value of 0.425mm. The value of the modal grain size then fluctuated wildly from a maximum of 0.875mm to a minimum of 0.425mm. It seems that waves of coarse material are released at intervals which appear to increase with time. This can only be interpreted tentatively because of the small number of sample points in the later stages of the experiment.

It has been found that sediment transport processes which involve systematic change can be clearly shown on a plot of the domain of ξ and χ. ξ and χ give indications of the peakiness and skewness of the hyperbolic distributions respectively. Figures 13, 14 show the trend discovered in experiment 3.

The points in this plot are grouped in three areas. In area I the skewness of the distributions were small indicating symmetical distributions which had a modal grain size of approximately 0.75mm. ξ indicates that the distributions were neither excessively peaked nor flat. In area II the distributions have skewed towards the smaller size fraction and a lower value of ξ suggests that the distributions were less well sorted about the mean grain size than before. The bedload samples now jumped from area II to area III, indicating that the bedload had skewed to the larger fractions - the modal grain size was approximately 0.85mm and the sample was better sorted, i.e. a more "peaked" shape.

These plots (fig. 13,14) indicate that several processes were involved in a systematic fashion in the creation of an armoured bed. The results show that rather coarse material was transported initially. After depletion of this fraction in the bed, smaller material was transported in greater quantities. When enough material was thus removed the sediment transport rate declined because of the non-availability of suitable sediment and the sheltering effect of the larger more resistant grains. The fluctuations in modal grain size, skewness and kurtosis occurred when these larger grains lost their stability because of the winnowing of fine sediment from around them. The larger grains were then transported away, leaving unprotected fines to be entrained as bedload, resulting in another fluctuation in the bedload characteristics. These fluctuations continue for a considerable time. As the transport rate decreases, the fluctuation sequences involve a progressively smaller proportion of the original size fractions. Thus a comparatively small number of coarse grain dislodgements are required to sustain transport rate fluctuations in the later stages of armouring.

3.2 Composition of the Surface Layer

It was clear from the photographs taken of the bed surface (Figs. 15, 16, 17, 18, 19, 20) that, as the sediment mix coarsened, the flow boundary got progressively rougher. This roughening began at the upstream end of the flume and progressed slowly downstream. Consequently the uniform depth increased at constant flow as the experiment progressed.

The experiments produced a substantial collection of plan view photographs of the changes occurring in the sediment bed. It is intended to analyse them using the technique outlined above, in order to discover the link between the changes in the grain population of bedload with the changes in the grain population on the sediment bed. The changes in the sediment transport rate and the bedload composition need to be related to the changes in the composition of the bed in a synchronised manner.

4.   <u>CONCLUSIONS</u>

This series of experiments though limited in number has produced evidence from which useful and reliable conclusions were reached.

There is confirmation again that sediment transport rate declines with time and that this decline may be split into two zones (Proffitt 1980), one of small change in the sediment transport rate and one of rapid decline. The pattern of decline in sediment transport is a function not only of flow vigour but also of the original bed composition.

Continuous sampling of the bed load enabled us to observe the composition of the bedload as it changed with time. When the mass-size distributions were plotted in a logarithmic plot fitted with hyperbolic curves it was revealed that the characteristics of these plots behaved in a systematic fashion dependent on the processes present. This information is important, as bedload sampling is not encumbered with the great difficulties associated with the sampling of the bed surface composition.

As the sediment transport rate dropped the bed progressively coarsened due to the loss of fines. The rate of coarsening declined markedly as the sediment transport rate declined and the composition of the bedload fluctuated. These bed surface observations are generally compatible with the bed load measurements. Although it is clear that there is a relationship between the compositions of the bed and of the bedload, further work still needs to be done to match with respect to time the characteristics of the bed with the known characteristics of the bedload.

Schedule of Figures and Tables

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Water Surface

Flow

Accelerated Armouring

Bedload

Sediment Bed

Collar to Prevent Slic Planes Forming

Valve (open)

Interchangeable Sediment Collection Box

Collected Sediment
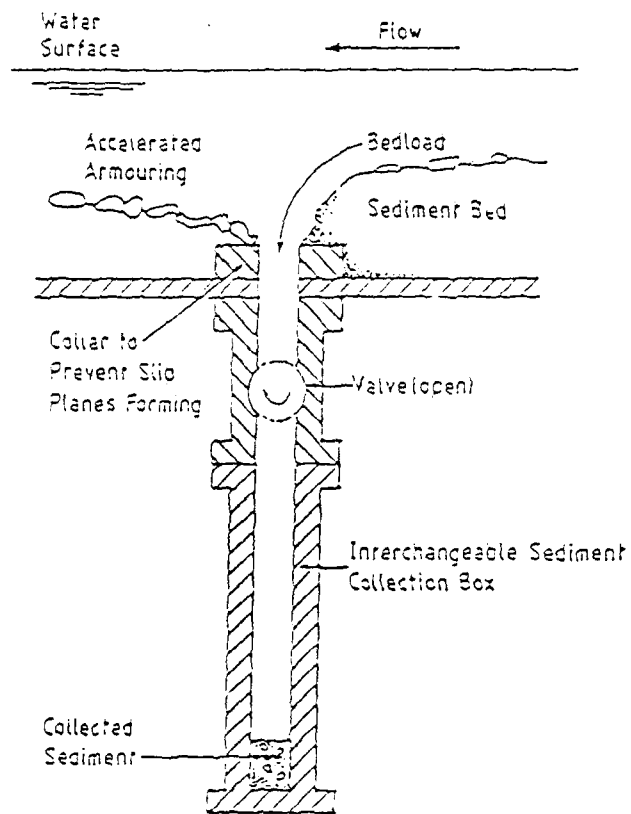
Figure 1   Side View of Bedload Sediment Trap Attached to Flume
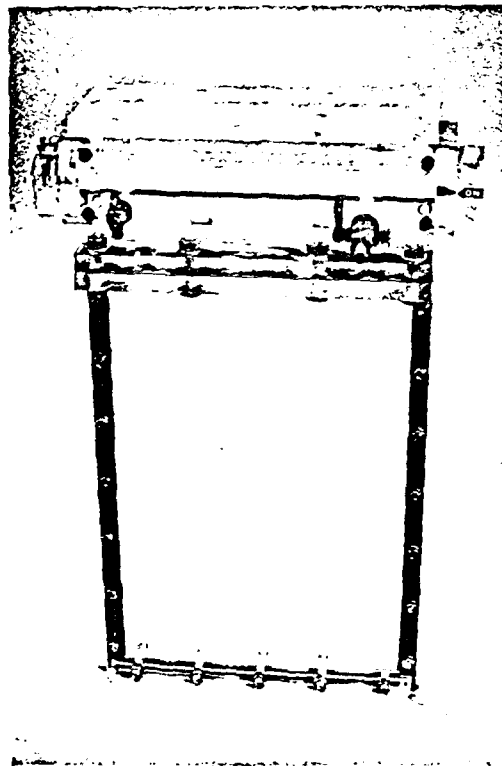
Figure 2   Photograph of Bedload Sediment Trap
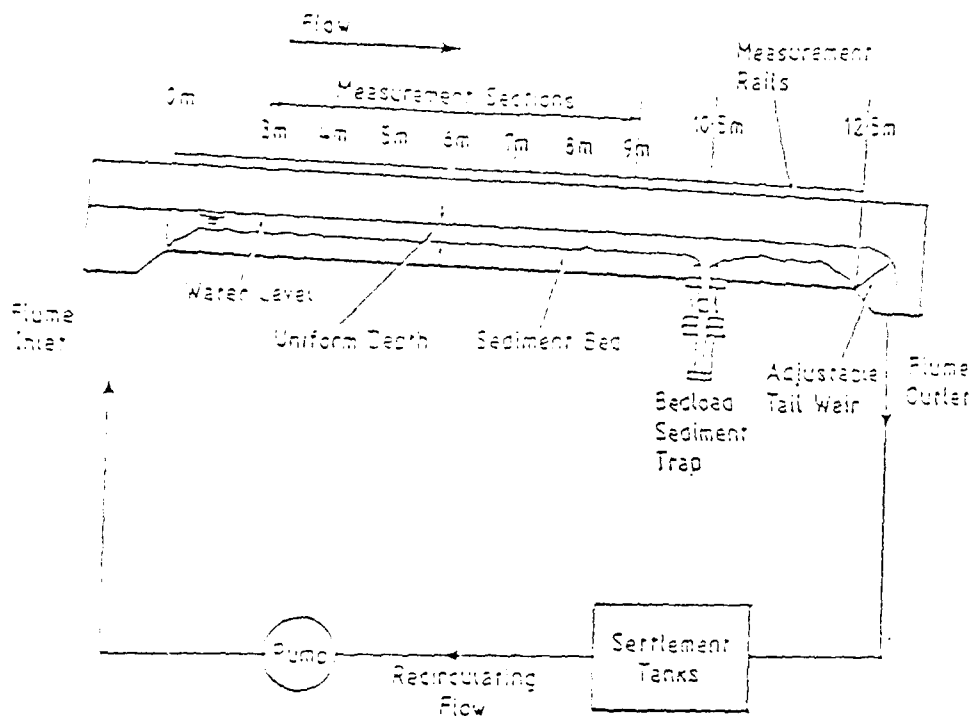
Figure 3   Sketch of Flume and Experimental Setup.

+   Experiment 1
o   Experiment 2
×   Experiment 3
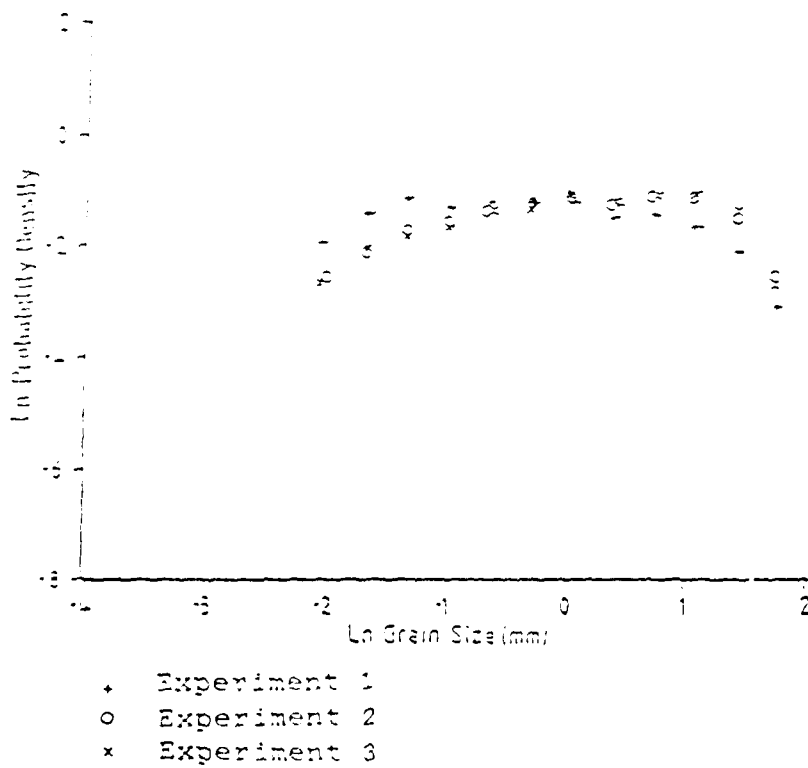
Figure 4   Size Distribution of Original Bed Mixtures
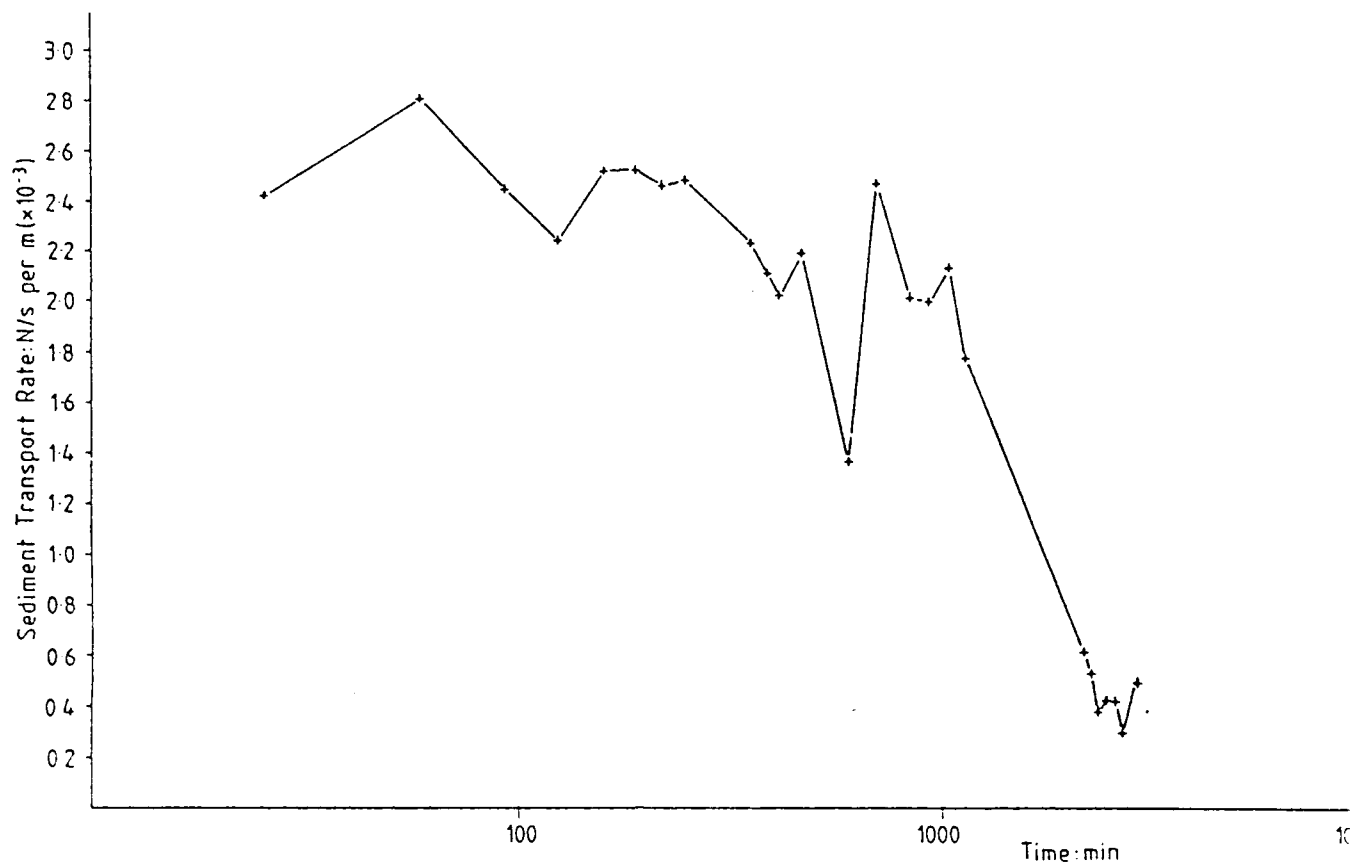
Figure 5   Decline in Sediment Transport with Time: Experiment 1
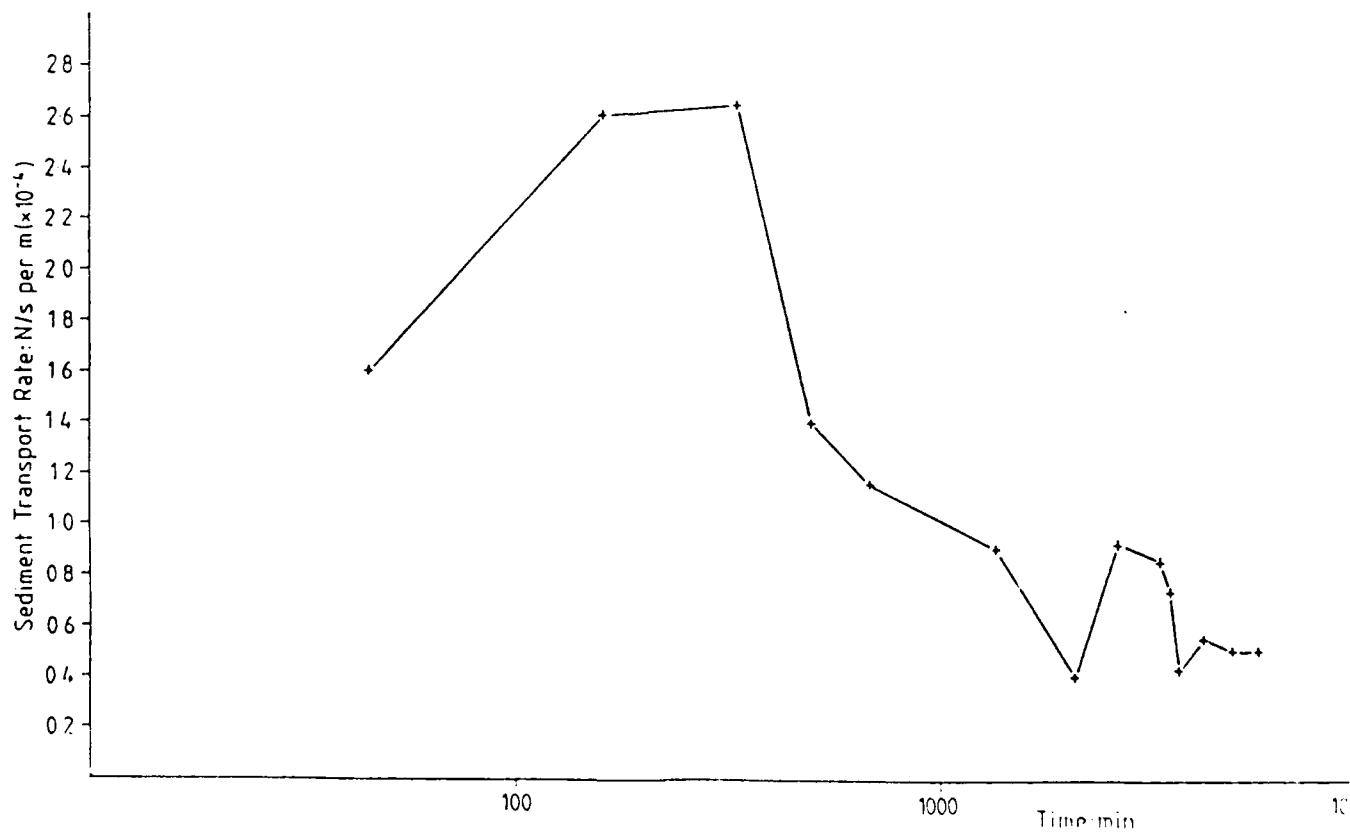
Figure 7   Decline in Sediment Transport with Time: Experiment 3
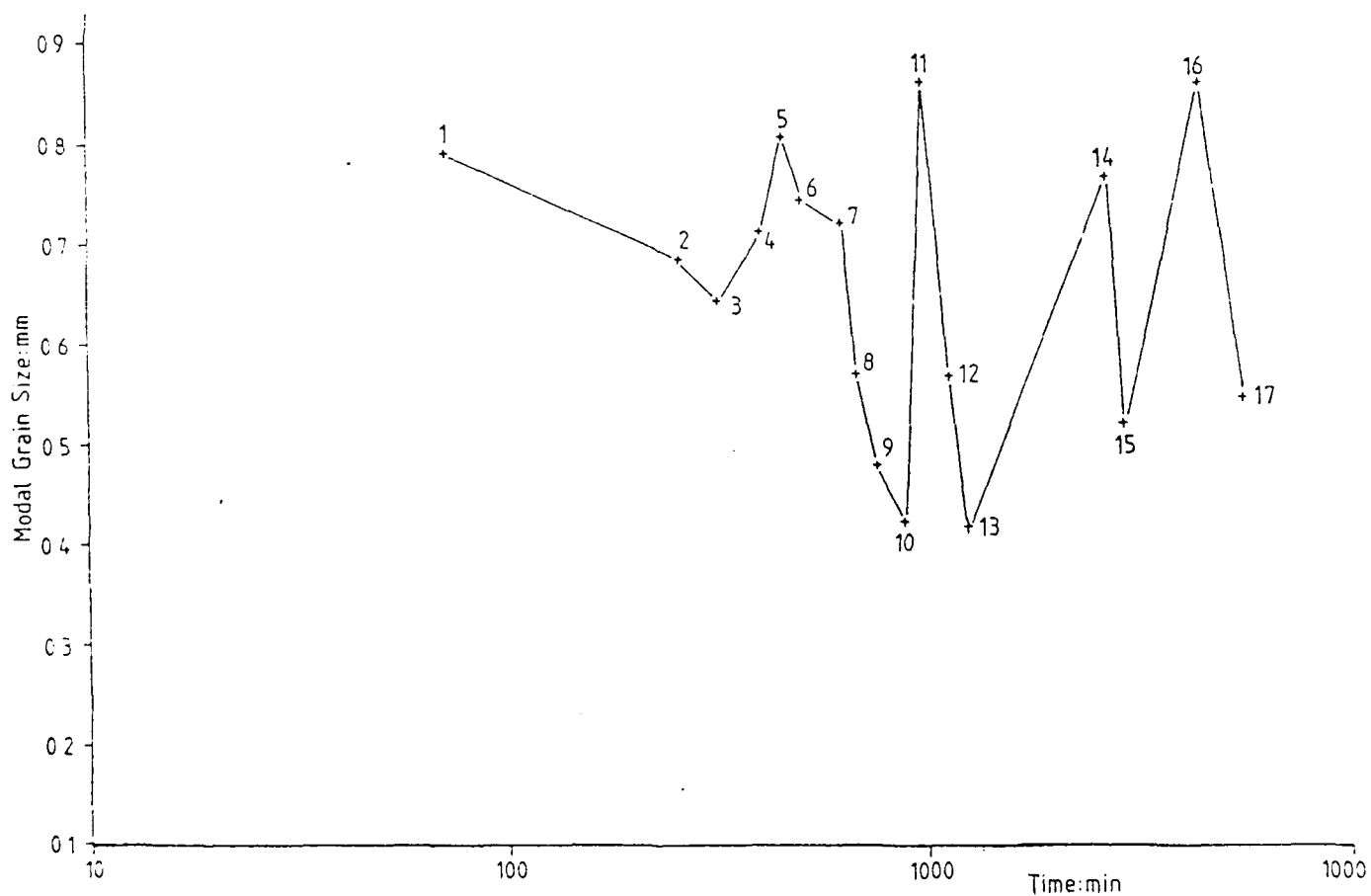
Figure 8   Modal Grain Size with Time: Experiment 3
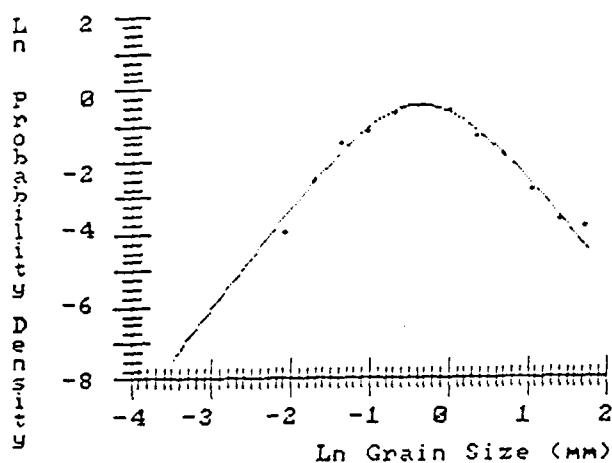
Figure 9
Size Distribution Bedload Sample
Experiment 3 Elapsed Time 70 mins.
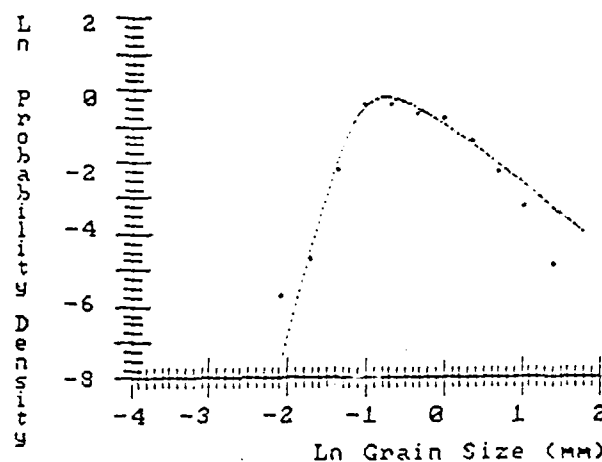
Figure 10
Size Distribution Bedload Sample
Experiment 3 Elapsed Time 755 min.
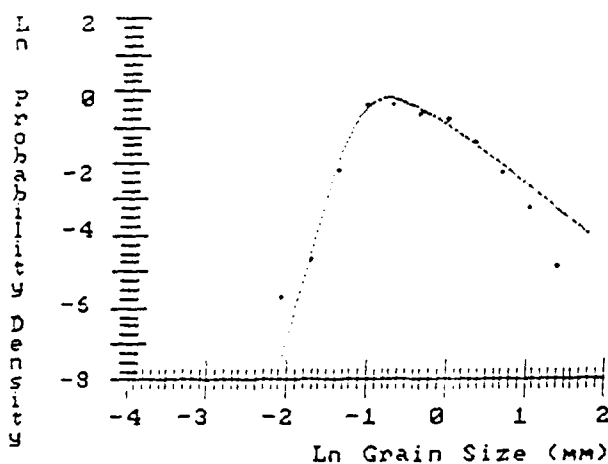
Figure 11
Size Distribution Bedload Sample
Experiment 3 Elapsed Time 1136 min.
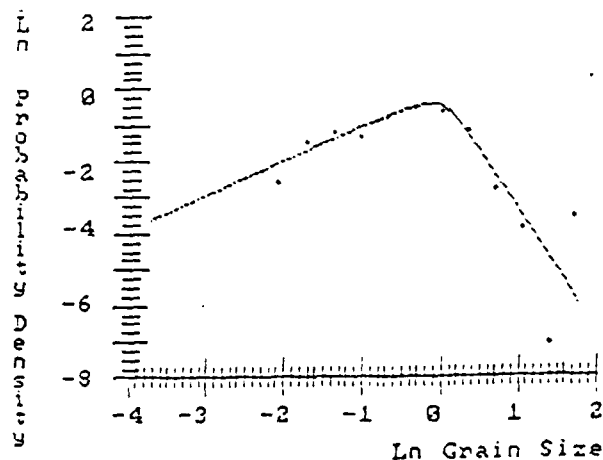
Figure 12
Size Distribution Bedload Sample
Experiment 3 Elapsed Time 4318 min.
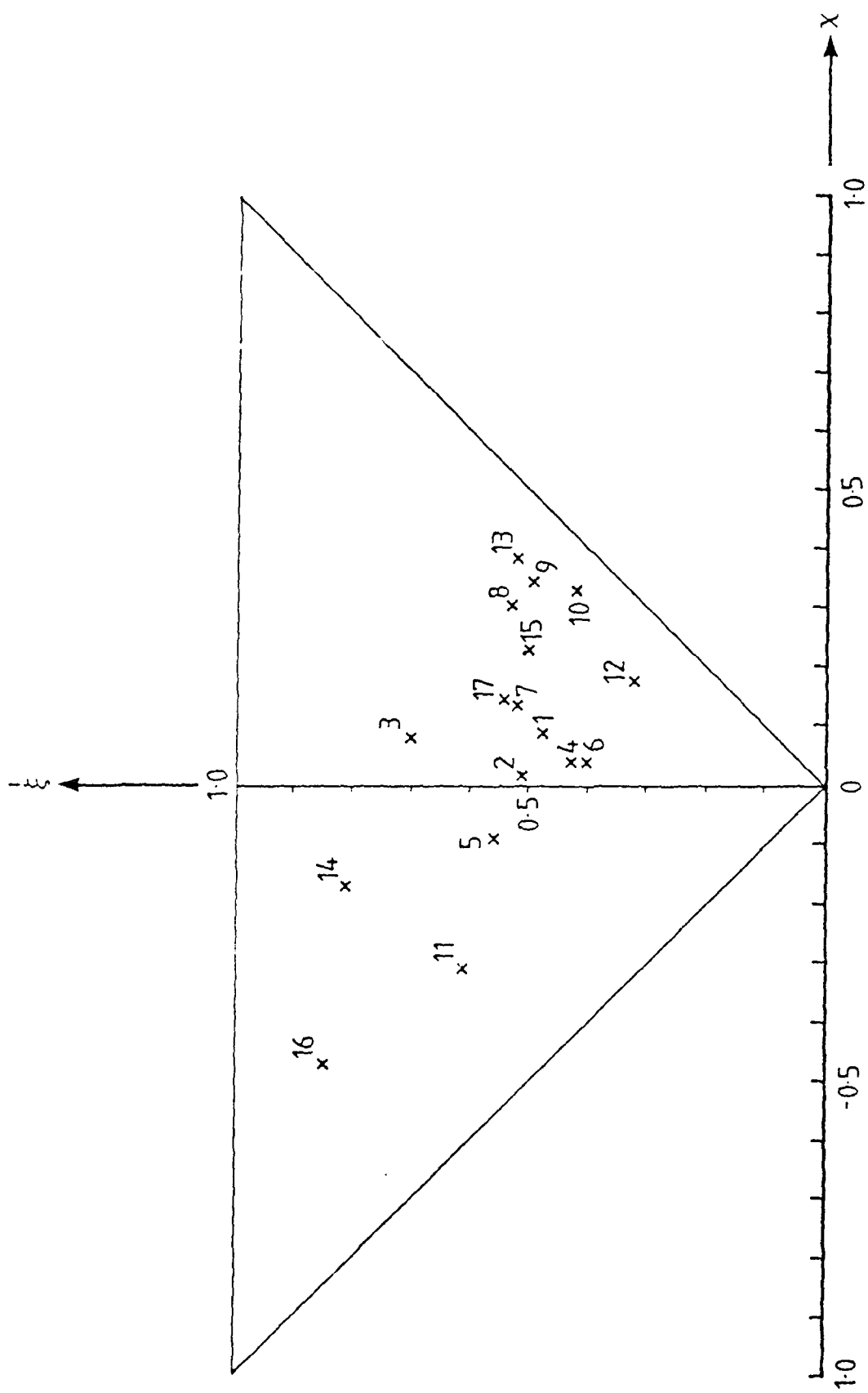
Figure 13   Plot of the domain of ξ and χ
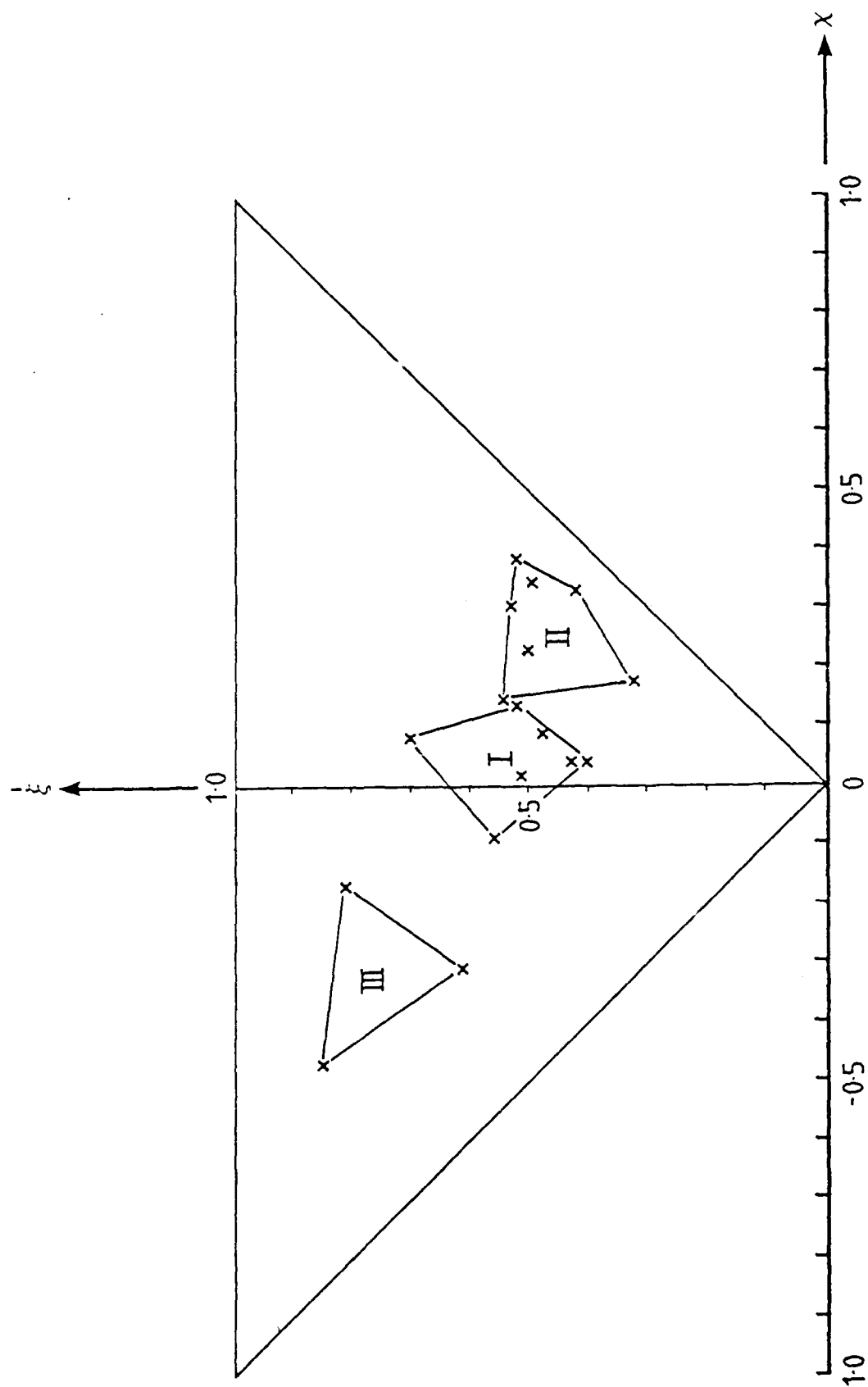
Figure 14 Plot of the Domain of ξ and γ outlining areas I, II, and III: Experiment 3.
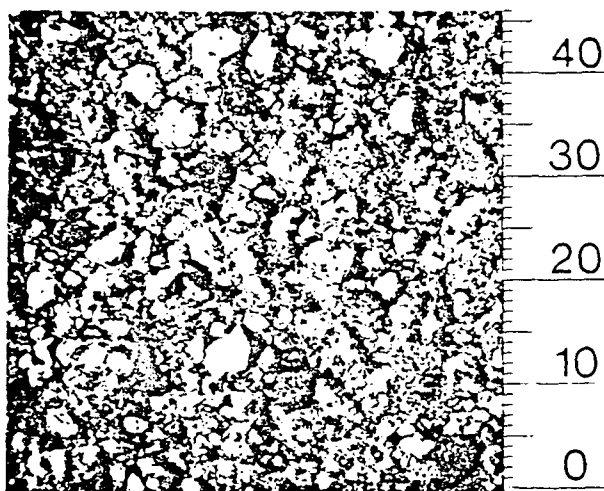
Fig. 15 .Plan View Photograph of Bed at 3 m
Time Elapsed 120 mins : Experiment 3
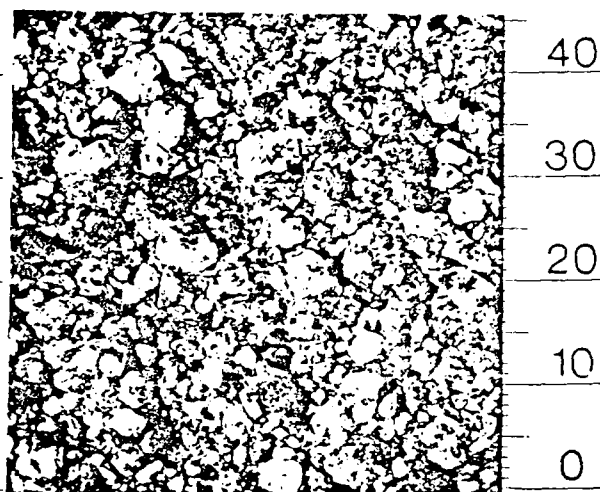
Figure 16 Plan View Photograph of Bed at 3 m
Time Elapsed 4157 mins ; Experiment 3
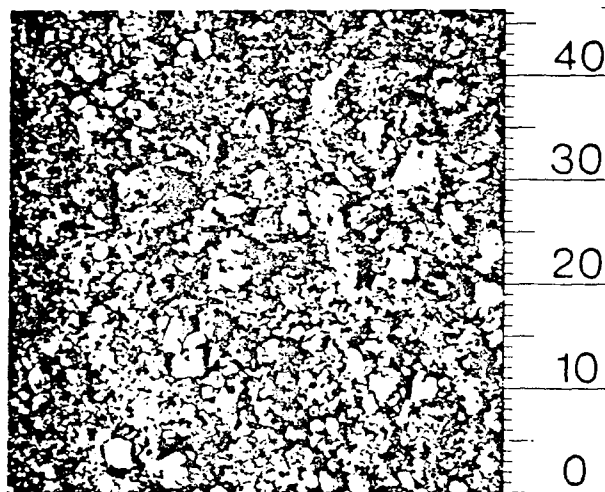
Figure 17 Plan View Photograph at Bed at 6 m
Time Elapsed 120 min. ; Experiment 3
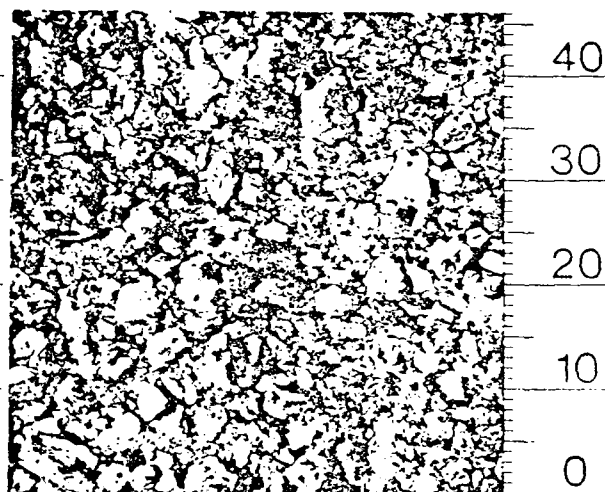
Figure 18 Plan View Photograph of Bed at 6 m
Eimt Elapsed 4157 min ; Experiment 3
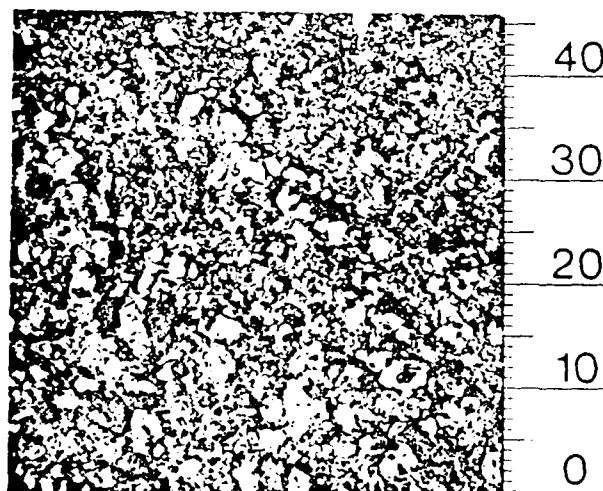
Figure 19 Plan View Photograph of Bed at 9 m
Time Elapsed 120 min : Experiment 3
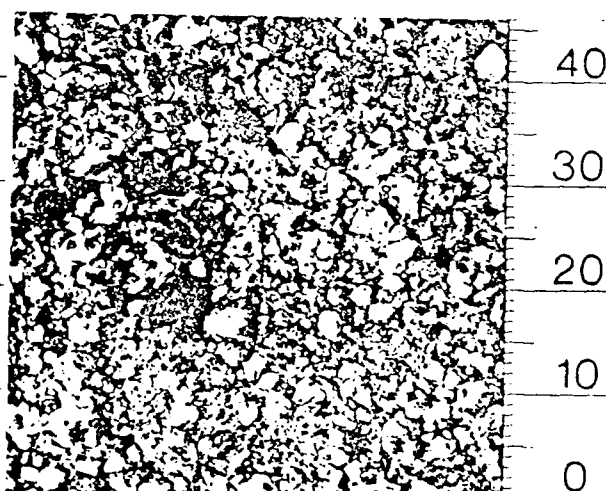
Figure 20 Plan View Photograph of Bed at 9 m
Time Elapsed 4157 min : Experiment 3

Flow Direction top to bottom of page

All scales in mm

3rd INTERNATIONAL WORKSHOP
ON GRAVEL-BED RIVERS

Firenze, 24-28 September 1990

ANALYSIS OF ARMOURING PROCESSES THROUGH LABORATORY EXPERIMENTS

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SUMMARY

The process of formation of the armour layer has been experimentally investigated. Eleven runs have been carried out using three different bed sediment mixtures combined with different hydraulic conditions. The characteristics of the armour layer are discussed through the analysis of the temporal evolution of bed surface composition and its effects on sediment transport process.

- 1 -

A.Lamberti E.Paris

1 - INTRODUCTION

Sediment transport in alluvial rivers originates from the mutual interaction between hydraulic and sediment characteristics; under certain circumstances the sediment transport process may lead to the formation of a protective cover generally consisting of a layer of grains on the bed surface (here called armour layer) which are coarser than the subsurface material.

This phenomenon has been investigated in several aspects:

- the sediment and hydraulic conditions under which the coarse layer develops (Lane & Carlson, Garde & Hasan, Little & Mayer, Knoroz, Proffitt, Sutherland);

- the effects on incipient motion and on sediment transport induced by the non uniformity of grains (Harrison, Einstein & Chien, Egiazaroff, Gessler '65, Day, Ashida & Michiue '73, Misri et Al., Proffitt & Sutherland, Samaga et Al., Falcon, Ribberink, Andrews & Parker);

- the relation between the final grain-size distribution of the coarse layer and the characteristics of the subsurface material (Gessler '65, Ashida & Michiue '71, Little & Mayer '72, Profitt & Sutherland, Saad)

- the erosion associated to the coarse layer development (Little & Mayer, Proffitt, Saad);

- the temporal evolution of the surface layer and/or transport rate during the formation process (Little & Mayer, Proffitt, Mosconi).

Mathematical simulation of the armouring processes has been also proposed by Hirano, Bayazit, Deigaard & Fredsoe, Borah et al., Shen & Lu, Karim & Kolly, Falcon, Ribberink, Lee & Odgaard, Willets et al., Palaniappan & Godbole, Armanini & Di Silvio.

In order to have a better understanding of the armour process, three set of experimental investigations has been carried out at the laboratory of "Istituto di Idraulica" of the University of Bologna (Italy). Preliminary results obtatined by the first set of experimental data have been already presented (Lamberti & Paris). Here, the total amount of experimental data are presented and discussed.

2 EXPERIMENTAL APPARATUS AND PROCEDURES

The experimental equipment has been already described in a previous paper (Lamberti & Paris); therefore only a brief description is given here.

A recirculating and tilting flume, 12 m long, 0.30 m wide and 0.50 m deep has been used to run sediment-free water over an initial flat bed made by non-cohesive sediment; the characteristics of the three sediment mixtures used in the experiments are summarized in table 1.

table 1. Sediment characteristics

Type of sediment	D_{16} mm	D_{50} mm	D_{84} mm	σ	Sg	ϵ
1	1.45	2.69	5.27	0.90	2.58	0.35
2	1.00	3.68	7.73	1.38	2.59	0.34
3	1.10	2.36	6.54	1.13	2.58	0.35

In each run, the formation of the armour layer was obtained under practically uniform flow conditions, during which measurements of liquid discharge, water temperature, piezometric and total head, velocity profiles, solid discharge, bed level variations and bed surface grain sizes have been carried out.

The runs were prolonged until a small fraction of the initial sediment transport (generally less than 5%) was collected in the sediment trap located at the downstream end of the flume; to this point the bed was considered to be armoured. Then, a sudden increase of discharge was

imposed until the armoured layer appeared completely removed.

Three series of runs have been carried out, one for each type of bed material, for a total number of 11 runs. The characteristics of each runs are summarized in tables 2-12 where symbols are: S_b, initial bed slope, t, run time (hours), Q, liquid discharge (l/s), T, water temperature (°C), h, water depth (cm), S_f, energy slope, U_v, mean velocity derived from velocity profiles (cm/s), U_p, mean velocity derived from Pitot measurements (cm/s), v_{*vp}, bed shear velocity obtained from velocity profile (cm/s), v_{*wc}, bed shear velocity obtained from side wall correction (cm/s), R_b, bed hydraulic radius (cm), Q_s, sediment discharge (kg/h), $D_{84}.D_{50}$, D_{16}, grain sizes for which the 84%, 50%, 16% of bed material is finer, while subscripts e and a refer to the eroded material (collected in the sediment trap) and the material of the armour layer surface, respectively.

2.1 BED SURFACE SAMPLING

One of the major problem to investigate the evolution of bed surface during an armouring process is the sampling technique; in order to preserve the integrity of the armour layer throughout the experiments , the size frequency distribution has been obtained by analyzing photographs of the bed surface according to the grid sampling method (Adams).

To this aim, photographs of the surface layer at the upstream, the middle and the downstream end of the testing reach have been taken at the beginning of the run, at the end of the armoured layer development and after its destruction.

Figure 1 shows a comparison between sediment diameters estimated from photograph analysis of bed surface and the sediment diameters obtained from sieve analysis, for the initial bed conditions. As it can be seen, there is a satisfactory agreement between the two set of diameters. Therefore, the same technique has been extended to survey the bed surface during the experiments.
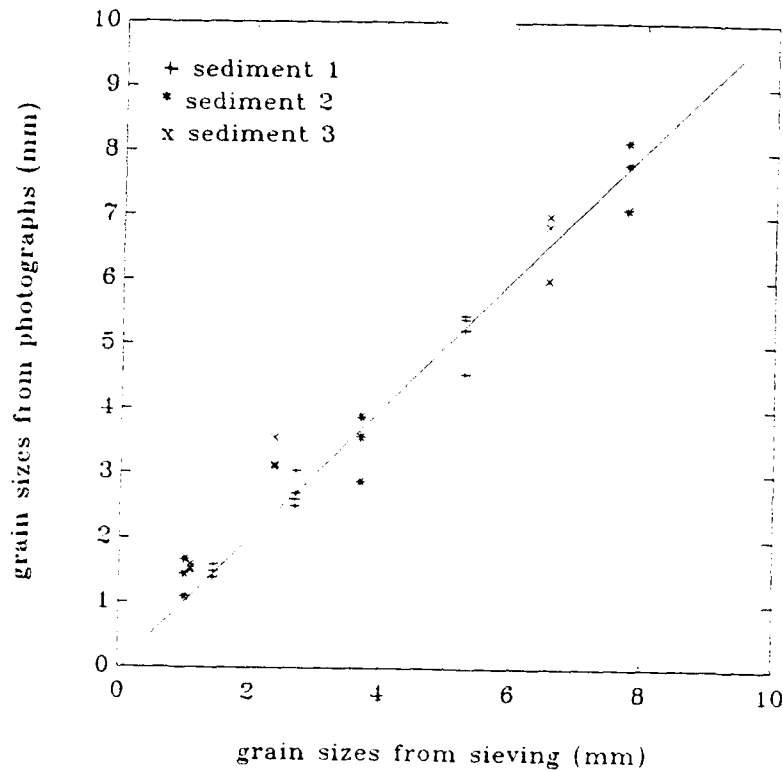
Fig. 1 - Sediment sizes estimated from photographs compared with sieve diameters.

2.2 SIDE WALL EFFECTS

During the experiments, runs have been conducted under a wide range of the width to depth ratio (from 1:1 up to 6:1); therefore, the side wall effects have been taken into account according to the procedure described in the previous paper (Lamberti & Paris). The reliability of the side wall correction has been verified by comparing the values of bed shear velocity $v_{*wc} = \sqrt{gR_bS_f}$ with the values of shear velocity obtained from analysis of velocity profiles, V_{*pv}. Results are listed in tables 2 - 12.

Table 2 - sediment 1

RUN 1

$D_{84} = 4.80$ mm $D_{50} = 2.82$ mm $D_{16} = 1.51$ mm $S_b = 0.002$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
hrs	l/s	°C	cm	×1000	cm/s	cm/s	cm/s	cm/s
0.00	40.8	18.8	18.8	2.45	-	-	-	5.4
2.00	"	"	18.9	2.40	-	-	-	5.3
5.00	"	18.9	19.1	2.25	-	-	-	5.2
11.00	"	19.4	19.4	2.15	-	-	-	5.1
13.00	"	19.3	"	2.10	-	-	-	5.1
18.50	"	19.6	19.8	2.08	-	-	-	5.1
21.50	"	19.9	20.0	2.01	-	-	-	5.1
28.50	"	20.0	20.0	2.00	-	-	-	5.1
43.50	"	20.4	20.2	1.95	-	-	-	5.0
59.50	"	20.8	20.3	1.85	-	-	-	5.0
84.50	"	21.1	20.3	1.95	-	-	-	5.0
85.50	68.7	20.2	30.0	1.45	-	-	-	4.5
86.00	"	20.3	-	-	-	-	-	-
87.00	"	20.2	-	-	-	-	-	-

	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
hrs	cm	kg/h	mm	mm	mm	mm	mm	mm
1.00	12.5	0.75	5.28	3.40	2.13	4.81	2.80	1.46
2.00	12.5	2.38	4.75	2.94	1.82	4.18	2.66	1.33
7.00	12.6	2.12	4.51	2.60	1.55	6.34	3.00	1.30
12.00	12.7	1.22	4.59	2.54	1.49	6.80	3.30	1.21
19.50	13.0	0.67	4.28	2.60	1.55	7.25	3.50	1.11
28.50	13.0	0.40	4.20	2.55	1.57	6.98	3.63	1.07
43.50	13.0	0.41	4.63	2.58	1.48	7.45	3.73	1.05
63.50	13.0	0.64	4.35	2.21	1.25	7.78	4.10	1.52
85.50	14.1	0.12	3.94	2.39	1.48	7.87	4.20	1.54
86.00	-	2.98	-	-	-	-	-	-
87.00	-	1.81	5.24	2.77	1.39	-	-	-

Table 3 - sediment 1

RUN 2

$D_{84} = 4.80$ mm $D_{50} = 2.82$ mm $D_{16} = 1.51$ mm $S_b = 0.002$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
hrs	l/s	°C	cm	× 1000	cm/s	cm/s	cm/s	cm/s
0.00	40.8	15.2	19.0	2.40	-	-	-	-
1.00	"	"	19.0	2.38	-	-	-	4.8
3.50	"	"	19.5	2.36	68	66	5.0	5.0
4.00	"	16.1	19.4	2.37	-	-	-	4.9
20.00	"	17.4	"	2.29	-	-	-	5.2
43.00	"	17.5	19.8	2.22	68	-	-	5.2
47.00	"	19.2	19.8	2.10	67	-	-	5.3
54.00	"	18.5	19.5	2.00	65	67	5.0	5.3
65.50	"	20.1	21.3	1.70	65	66	5.3	5.3
68.50	68.0	20.0	31.1	1.65	-	-	-	6.0
69.00	"	20.0	31.2	1.65	72	-	-	6.0
69.50	"	19.0	31.1	1.65	-	-	-	6.1
71.50	"	19.0	31.1	1.65	-	-	-	6.1

t	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
hrs	cm	kg/h	mm	mm	mm	mm	mm	mm
1.00	12.0	2.56	4.59	2.88	1.71	5.35	2.77	1.39
4.00	12.4	2.28	4.59	2.52	1.51	6.11	3.01	1.32
20.00	13.8	1.41	4.92	2.57	1.51	7.70	3.73	1.16
43.00	14.0	0.04	4.90	2.41	1.46	7.70	3.77	1.16
47.00	14.0	0.34	4.36	2.46	1.43	7.60	3.81	1.16
54.00	14.2	0.49	6.64	3.20	1.52	7.50	3.61	1.16
68.50	14.4	0.17	4.23	2.43	1.43	7.14	3.73	1.16
69.00	18.8	1.16	4.85	3.01	1.48	-	-	-
69.50	19.1	1.38	4.92	2.75	1.45	-	-	-
71.50	19.1	3.48	4.53	2.57	1.47	-	-	-

Table 4 - sediment 1

RUN 3

$D_{84} = 4.80$ mm $D_{50} = 2.82$ mm $D_{16} = 1.51$ mm $S_b = 0.005$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
h:s	l/s	°C	cm	×1000	cm/s	cm/s	cm/s	cm/s
1.00	11.9	18.5	6.20	7.00	53	54	-	6.1
2.00	"	"	6.40	6.90	53	52	-	-
3.00	"	-	6.40	6.75	-	-	-	6.1
4.00	"	-	6.40	6.50	53	-	-	6.0
5.00	"	-	-	-	-	-	-	-
7.00	"	-	-	-	-	-	-	-
8.00	"	18.7	-	-	-	-	6.5	-
9.50	"	18.8	6.60	6.30	53	-	-	6.0
11.00	"	18.9	-	-	-	-	-	-
13.00	"	19.2	6.60	6.30	53	-	-	-
15.00	"	19.3	6.60	6.30	53	-	-	-
18.00	"	19.4	6.60	6.30	53	-	6.7	6.0
18.50	15.6	19.5	8.00	6.00	55	-	-	-
19.00	"	19.0	-	5.50	-	-	-	-
19.50	19.6	19.7	9.40	5.00	64	-	-	6.3
20.00	"	-	-	-	-	-	-	-
20.50	"	19.7	9.60	-	-	-	-	-

t	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
h:s	cm	kg/h	mm	mm	mm	mm	mm	mm
1.00	5.40	17.56	4.96	2.93	1.74	6.16	2.40	1.08
2.00	-	18.16	3.83	2.23	1.41	7.75	3.78	1.04
3.00	5.70	13.80	4.22	2.44	1.46	7.83	4.24	1.02
4.00	5.70	4.66	4.00	2.34	1.38	7.94	4.49	1.08
5.00	-	2.21	4.31	2.63	1.52	8.01	4.63	1.03
7.00	-	2.21	4.03	2.30	1.32	8.18	4.86	1.19
9.50	5.80	1.71	4.20	2.30	1.30	8.13	4.93	1.35
11.00	-	1.25	4.16	2.58	1.69	8.17	4.98	1.19
13.00	-	2.71	4.53	2.49	1.47	8.20	5.02	1.02
15.00	-	1.30	4.69	2.75	1.58	8.38	5.02	1.01
18.00	5.80	0.54	4.25	2.41	1.44	8.22	5.07	1.00
18.50	-	0.56	5.65	2.77	1.29	-	-	-
19.00	-	5.24	5.20	2.77	1.47	-	-	-
19.50	8.0	4.03	5.24	2.43	1.26	-	-	-
20.00	-	23.98	4.56	2.51	1.39	-	-	-
20.50	8.1	30.61	4.44	2.39	1.36	-	-	-

Table 5 - sediment 1

RUN 4

$D_{84} = 4.80$ mm $D_{50} = 2.82$ mm $D_{16} = 1.51$ mm $S_b = 0.005$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
h-s	l/s	°C	cm	× 1000	cm/s	cm/s	cm/s	cm/s
1.00	11.9	17.6	6.00	5.30	56	57	5.8	5.2
2.00	"	17.8	6.50	5.30	56	57	-	5.4
3.00	"	-	-	-	-	-	-	-
7.00	"	18.0	6.50	5.00	57	-	-	5.3
10.00	"	18.2	6.50	-	-	-	-	-
22.50	"	18.1	6.50	-	-	-	-	-
23.50	"	17.9	6.50	5.00	-	-	-	-
24.00	19.7	18.4	8.80	4.80	-	-	-	5.8
24.50	"	18.5	-	4.60	-	-	-	-
26.00	"	-	9.50	4.50	66	65	6.7	5.8

t	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
h-s	cm	kg/h	mm	mm	mm	mm	mm	mm
1.00	5.20	12.44	4.96	2.91	1.63	5.67	2.65	1.35
2.00	5.70	-	-	-	-	-	-	-
3.00	-	4.13	4.22	2.53	1.54	6.87	2.99	1.17
7.00	5.70	1.33	3.94	2.43	1.48	7.00	3.48	1.12
10.00	-	0.60	3.92	2.31	1.40	7.20	3.63	1.11
22.50	-	0.31	4.35	2.53	1.47	7.63	3.76	1.10
23.50	-	0.28	3.81	2.26	1.44	7.65	3.78	1.10
24.00	7.20	15.88	5.24	2.67	1.37	-	-	-
24.50	-	13.24	4.95	2.69	1.49	-	-	-
26.00	7.70	6.74	4.72	2.58	1.46	-	-	-

Table 6 - sediment 1

RUN 5

$D_{84}=7.73$mm $D_{50}=3.68$mm $D_{16}=1.00$mm $S_b=0.01$

t	Q	T	h	Sf	Up	Uv	v*
hrs	l/s	°C	cm	×1000	cm/s	cm/s	cm/s
1.00	6.4	17.4	4.0	10.30	44	45	6.3
3.00	"	"	-	-	-	-	-
7.00	"	17.5	4.0	9.80	43	43	6.2
10.00	"	"	"	"	-	-	-
22.00	"	-	-	-	-	-	-
24.00	"	17.5	4.0	9.20	44	45	6.0
24.50	11.3	"	5.2	8.20	68	69	6.3
25.00	"	"	"	"	-	-	-
26.00	"	"	-	-	-	-	-

t	Qs	d84e	d50e	d16e	d84a	d50a	d16a
hrs	kg/h	mm	mm	mm	mm	mm	mm
1.00	10.82	4.86	2.93	1.71	5.94	2.61	1.33
3.00	4.77	4.35	2.60	1.57	7.04	2.91	1.11
7.00	1.57	4.16	2.48	1.45	7.19	3.42	1.05
10.00	0.65	4.08	2.31	1.41	7.32	3.55	1.05
22.00	0.50	4.22	2.34	1.39	7.84	3.84	1.07
24.00	0.39	4.08	2.38	1.43	7.86	3.86	1.07
24.50	22.26	5.31	2.96	1.61	-	-	-
25.00	21.72	4.72	2.54	1.41	-	-	-
26.00	12.17	4.82	2.54	1.40	-	-	-

Table 7-sediment 2

RUN 6

$D_{84}=7.73$ mm $D_{50}=3.68$ mm $D_{16}=1.00$ mm $S_b=0.01$

t	Q	T	h	Sf	Up	Uv	v*
hrs	l/s	°C	cm	× 1000	cm/s	cm/s	cm/s
0.50	6.4	16.7	3.8	-	-	-	-
1.50	"	"	"	-	-	48	-
2.50	"	16.7	"	10.50	52	-	6.0
3.50	"	-	-	-	-	-	-
7.00	"	16.8	3.8	10.50	50	-	6.0
7.50	11.2	17.0	5.2	-	-	-	6.6
8.00	"	17.0	4.8	-	-	-	-
9.00	"	-	"	9.00	68	69	6.1
10.50	"	-	"	8.50	72	-	-
11.00	"	17.2	-	-	-	-	-
14.50	"	17.4	4.9	8.50	76	-	5.9
15.00	16.9	"	5.2	-	-	-	6.3
15.50	"	"	6.0	-	-	-	-
16.50	"	"	6.3	-	76	77	-
17.00	"	"	6.4	7.50	-	-	6.2
18.50	"	"	6.5	-	-	-	-
21.00	"	"	6.7	7.00	76	-	-
21.50	"	"	-	-	-	-	6.2

t	Qs	d84e	d50e	d16e	d84a	d50a	d16a
hrs	kg/h	mm	mm	mm	mm	mm	mm
00.50	11.72	5.78	1.57	1.03	7.98	4.85	0.99
01.50	01.78	1.88	1.23	0.85	8.18	5.58	1.01
03.50	00.47	1.68	1.17	0.80	8.28	5.72	1.03
07.50	00.19	1.65	1.16	0.79	8.36	5.81	1.05
08.00	30.28	6.80	1.82	0.96	9.08	6.33	1.27
09.00	10.13	5.93	1.38	0.85	9.44	6.81	1.27
11.00	01.41	5.10	1.26	0.79	9.52	6.94	1.28
15.00	00.42	4.48	1.22	0.77	9.58	7.01	1.28
15.50	35.56	7.34	4.62	1.01	9.65	7.05	1.17
16.50	19.30	7.05	1.84	0.90	9.74	7.12	1.23
18.50	02.26	6.03	1.29	0.77	9.79	7.26	1.33
21.50	00.47	5.98	1.27	0.78	9.84	7.28	1.37

Table 8 - sediment 2

RUN 7

$D_{84} = 7.73$ mm $D_{50} = 3.68$ mm $D_{16} = 1.00$ mm $S_b = 0.005$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
h:s	l/s	°C	cm	× 1000	cm/s	cm/s	cm/s	cm/s
1.50	12.6	17.5	6.50	6.70	56	54	6.5	-
2.00	"	-	6.50	6.70	54	-	-	6.2
3.50	"	17.6	-	-	-	-	-	-
7.50	23.7	17.8	10.10	6.70	57	-	-	6.0
8.00	"	17.9	10.20	-	55	63	8.2	-
9.00	"	18.0	10.40	4.50	-	-	-	-
10.00	"	"	-	-	-	-	-	-

t	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
h:s	cm	kg/h	mm	mm	mm	mm	mm	mm
0.50	5.70	7.75	6.81	1.93	1.08	7.87	2.66	0.98
1.50	5.70	2.37	5.68	1.34	0.92	7.97	4.41	0.99
3.50	-	0.76	1.99	1.26	0.87	8.13	4.93	1.00
7.50	-	0.27	1.88	1.21	0.82	8.24	5.33	1.02
8.00	5.70	55.63	7.22	2.79	0.99	9.14	5.77	1.01
9.00	8.00	12.65	6.54	1.53	0.89	9.32	5.95	1.08
10.00	-	1.44	5.85	1.32	0.89	9.43	6.45	1.09

Table 9 - sediment 2

RUN 8

$D_{84} = 7.73$ mm $\qquad$ $D_{50} = 3.68$ mm $\qquad$ $D_{16} = 1.00$ mm $\qquad$ $S_b = 0.002$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
h zs	l/s	°C	cm	× 1000	cm/s	cm/s	cm/s	cm/s
1.00	47.4	18.3	22.3	-	56	-	-	-
2.00	52.9	"	24.8	-	-	-	-	-
3.00	"	"	"	2.04	-	64	7.0	5.6
4.00	"	18.7	"	-	65	-	-	-
7.50	"	-	-	2.00	65	-	-	-
8.00	"	19.1	24.8	-	-	-	-	-
11.50	"	-	"	2.00	65	-	-	5.6
15.00	"	19.0	-	-	-	-	-	-
28.50	"	-	24.9	2.00	65	-	-	5.6
29.00	67.5	19.5	26.2	-	-	-	-	-
29.50	"	18.8	-	-	-	-	-	-
30.50	"	19.0	26.8	2.50	72	78	8.5	6.3
31.00	"	19.2	-	-	-	-	-	-

t	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
h zs	cm	kg/h	mm	mm	mm	mm	mm	mm
1.00	15.5	1.34	4.89	1.44	1.03	7.39	3.69	1.00
2.00	"	0.97	4.10	1.34	0.98	7.79	4.24	1.00
3.00	"	-	-	-	-	-	-	-
4.00	-	0.47	2.34	1.30	0.94	7.87	4.55	1.00
7.50	15.5	-	-	-	-	-	-	-
8.00	-	0.23	1.91	1.25	0.88	7.79	4.38	1.01
15.00	-	0.11	6.43	1.50	0.95	7.85	5.00	1.01
29.00	16.0	0.06	6.63	1.38	0.91	7.91	5.15	1.02
29.50	16.3	16.40	6.44	1.50	0.87	8.56	5.89	1.08
30.50	16.3	15.93	6.78	1.69	0.90	9.02	6.34	1.13
31.00	-	5.36	7.01	1.73	0.93	9.12	6.38	1.14

Table 10 - sediment 3

RUN 9

$D_{84} = 6.54$ mm $D_{50} = 2.36$ mm $D_{16} = 1.10$ mm $S_b = 0.002$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
h₁₅	l/s	℃	cm	× 1000	cm/s	cm/s	cm/s	cm/s
1.00	47.3	17.9	20.5	-	65	67	-	-
3.00	"	18.3	"	2.10	64	-	7.1	5.2
16.00	"	18.5	20.7	2.30	64	-	-	-
18.00	"	19.2	"	-	-	-	-	-
25.00	"	19.4	-	-	-	-	-	-
27.00	"	19.7	20.8	-	-	-	6.9	-
40.50	"	-	"	2.40	65	-	-	5.4
41.00	67.3	20.1	26.5	-	-	79	7.7	5.8
42.00	"	-	26.9	3.00	-	-	-	-
43.00	"	-	27.7	-	67	-	-	-

t	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
h₁₅	cm	kg/h	mm	mm	mm	mm	mm	mm
0.50	-	2.63	5.67	2.76	1.36	6.85	2.38	1.11
1.50	-	1.30	4.66	2.43	1.43	7.07	2.37	1.10
3.00	13.1	-	-	-	-	-	-	-
3.50	-	2.01	5.39	2.32	1.27	7.67	2.39	1.07
16.00	13.1	0.62	5.36	2.07	1.73	8.49	2.78	0.94
18.00	-	0.21	4.38	2.05	1.29	8.54	2.83	0.93
25.00	-	0.24	3.71	1.78	1.23	8.73	3.48	0.92
27.00	-	0.18	3.82	1.74	1.07	8.77	3.56	0.92
40.50	13.2	-	-	-	-	-	-	-
41.00	-	0.10	4.42	1.84	1.08	8.88	3.94	0.93
41.50	16.7	37.25	6.58	2.52	1.08	9.42	2.45	1.06
42.50	-	29.09	6.10	2.48	1.20	9.51	2.70	0.97

Table 11 - sediment 3

RUN 10

$D_{84} = 6.54$ mm $D_{50} = 2.36$ mm $D_{16} = 1.10$ mm $S_b = 0.005$

t	Q	T	h	Sf	Up	Uv	v_{*pv}	v_{*wc}
hrs	l/s	℃	cm	× 1000	cm/s	cm/s	cm/s	cm/s
1.50	12.2	19.2	6.10	-	-	-	-	-
2.00	"	19.4	"	5.30	58	62	6.8	5.2
5.50	"	-	"	5.50	57	-	"	-
7.50	"	19.6	"	-	-	-	-	-
9.50	"	19.7	"	-	-	-	-	-
22.50	"	-	6.20	5.10	59	-	-	-
23.50	21.1	20.0	8.90	5.10	-	64	8.8	-
24.00	"	-	"	-	-	-	-	-
24.50	"	18.5	-	4.60	-	-	-	-

t	Rb	Qs	d84e	d50e	d16e	d84a	d50a	d16a
hrs	cm	kg/h	mm	mm	mm	mm	mm	mm
0.50	-	12.06	5.52	2.34	1.27	7.33	2.31	1.05
1.50	-	2.82	4.78	1.86	1.08	7.59	2.51	1.06
2.00	5.1	-	-	-	-	-	-	-
3.50	-	0.96	4.96	1.98	1.14	7.74	2.61	1.05
5.50	5.1	-	-	-	-	-	-	-
7.50	-	0.46	5.00	1.89	1.14	7.85	2.75	1.04
9.50	-	0.20	3.73	1.75	1.08	7.88	2.80	1.05
22.50	5.1	0.12	4.48	1.87	1.10	7.99	3.06	1.05
23.50	-	0.12	3.78	1.81	1.12	8.00	3.08	1.05
24.00	-	25.57	6.73	3.05	1.26	8.49	2.25	0.97
24.50	7.0	-	-	-	-	-	-	-
25.00	-	13.40	6.61	2.90	1.22	8.97	2.26	0.94

Table 12- sediment 3

RUN 11

D_{84}= 6.54 mm D_{50}= 2.36 mm D_{16}= 1.10 mm S_b= 0.01

t	Q	T	h	Sf	Up	Uv	v*
hrs	l/s	°C	cm	x 1000	cm/s	cm/s	cm/s
0.50	6.5	18.7	3.5	-	-	-	-
1.00	"	-	"	-	-	58	-
1.50	"	-	"	10.00	56	-	5.9
3.50	"	19.2	-	-	-	-	-
7.50	"	19.4	-	-	-	-	-
9.50	"	-	3.4	10.50	57	-	5.9
10.50	"	19.4	-	-	-	-	-
22.50	"	19.5	"	-	58	-	-
23.50	13.0	"	4.9	10.00	-	-	6.9
24.50	"	-	6.7	8.50	76	-	7.4

t	Qs	d84e	d50e	d16e	d84a	d50a	d16a
hrs	kg/h	mm	mm	mm	mm	mm	mm
00.50	20.93	5.72	2.46	1.26	7.31	2.65	1.05
01.50	5.90	4.93	1.99	1.15	7.82	3.22	1.03
03.50	1.15	4.19	1.73	1.04	7.99	3.64	1.05
07.50	0.41	4.08	1.82	1.12	8.17	3.93	1.05
10.50	0.25	3.73	1.68	1.03	8.25	4.07	1.05
22.50	0.10	3.84	1.74	1.07	8.37	4.28	1.06
23.50	0.07	3.06	1.63	1.00	8.38	4.30	1.06
24.00	66.00	7.23	4.14	1.45	8.54	1.67	0.92
25.00	22.38	6.23	2.34	1.12	8.95	1.68	0.93

3 FRAMEWORK OF ANALYSIS

Let us assume the simple Bayazit scheme to describe the exchange of grains between the surface and the subsurface layer of bed sediment; if the approaching flow is sediment depleted, grains will be picked up from the mixing layer and replaced in the same quantity by the parent bed material entering the moving control volume from the layer below the mixing layer (Fig. 2).

We denote with: m_i, the fraction of grains in class interval i for the mixing layer; p_i, the fraction of grains in class interval i in the parent bed; t_i, the fraction of grains in class interval i for the transported material; M and P, the mass of sediment per unit area, in the mixing layer and in the parent bed, respectively; assuming constant the apparent density on bed material, P and M may also be interpreted as the elevation of the top of the parent bed and as the thickness of the mixing layer, respectively; T, the mass sediment transport rate per unit width. The equation for the conservation of mass applied to the moving control volume for each size fraction yields:
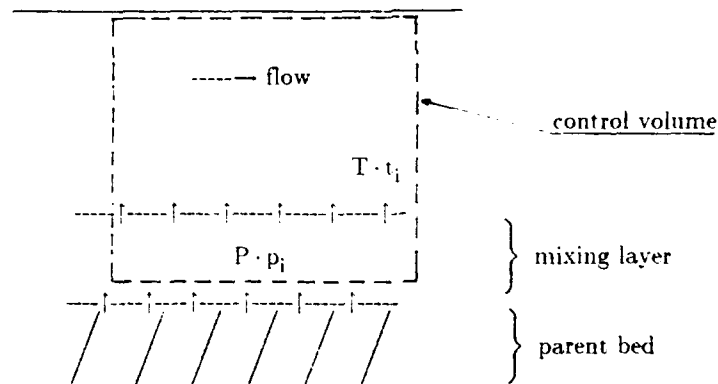
Fig. 2 - Definition sketch of the of the mixing layer.

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1)
$$\frac{\partial(M\ m_i)}{\partial t} + p_i(P)\ \frac{\partial(P)}{\partial t} + \frac{\partial(T\ t_i)}{\partial x} = 0$$

(the mass of the sediment in motion is neglected).

Summing over all the size classes and considering that:

$$\sum_i m_i = \sum_i p_i = \sum_i t_i = 1$$

we obtain:

2)
$$\frac{\partial(M + P)}{\partial t} + \frac{\partial T}{\partial x} = 0$$

Denoting with E the eroded mass rate:

3)
$$E = -\frac{\partial(M + P)}{\partial t} = \frac{\partial T}{\partial x}$$

Under the assumptions that during the armouring process the thickness of mixing layer is constant with time, and the composition of eroded material is approximately constant along the channel length, that is:

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$$\frac{\partial M}{M \partial t} << \frac{\partial m_i}{m_i \partial t} \qquad \frac{\partial T}{T \partial x} << \frac{\partial t_i}{t_i \partial x}$$

the equation of mass conservation gives:

4)
$$M \frac{\partial m_i}{\partial t} + p_i \frac{\partial P}{\partial t} + t_i \frac{\partial T}{\partial x} = 0$$

In the same order of approximation, eq. 4 can be written as:

5)
$$M \frac{\partial m_i}{\partial t} + E (t_i - p_i) = 0$$

4 ANALYSIS OF EXPERIMENTAL DATA
4.1 TEMPORAL VARIATION OF BED SURFACE GRAIN SIZES

Changes in grain size distribution occurring at the bed surface during the armouring process can be evaluated through the continuity equation for each size fraction. eq. 5. applied to an appropriate control volume. In particular. by integrating eq. 5 over the finite interval Δt. we obtain:

7)
$$M \Delta m_i = \Delta E (p_i - t_i)$$

where Δm_i is the variation of the fraction i in the mixing layer during the time interval Δt. and ΔE is the mass of sediment which is removed from the bed during the same interval Δt. The value of the thickness of the mixing layer has been found by comparing the results obtained by eq. 7 with those obtained by the analysis of photographs. The value which provides the best

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fitting with experimental data has been found to be equal to the D_{50} of the armoured layer during its formation. while for the calculation of the bed surface grain size evolution during the destruction process the best results were obtained with the thickness of the mixing layer equal to $2 \cdot D_{50}$.

In figures 3 and 4 are plotted the sediment diameters of bed surface obtained from photographs against the corresponding values calculated through eq. 7. for armoured conditions and after destruction of the armoured layer, respectively.
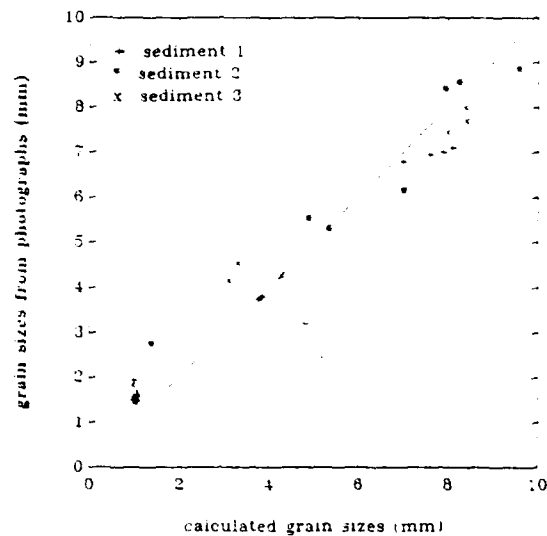
Fig. 3 - Calculated and observed sediment sizes of the surface layer for armour conditions.

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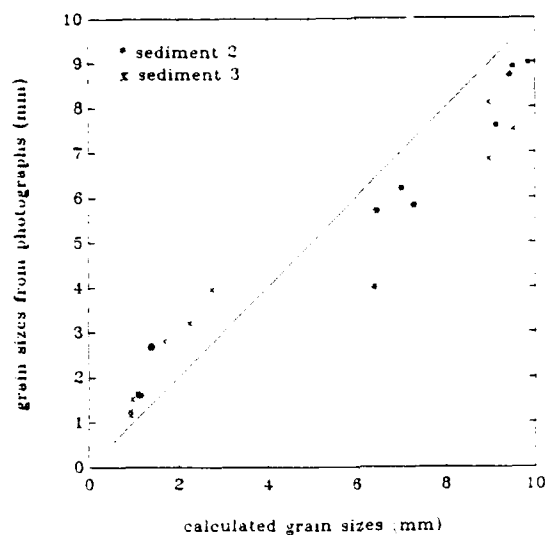
Fig. 4 - Calculated and observed surface sediment sizes after destruction of the armour layer.

4.2 FORMATION OF THE ARMOUR LAYER

Let v_i be any variable associated to the size class i (e.g., the central diameter of size fraction i, D_i, or the square D_i^2), and v_m the mean value of v_i according to the frequency distribution curve m_i (i.e., $\sum_i v_i m_i$); hence, v_m represents the mean value of a generic variables for the mixing layer:

Similarly, extending the same definition to the parent bed and the eroded material, we obtain the mean values v_p and v_t respectively.

Multiplying equation 5 by v_i and summing over all classes, we obtain:

6) $$M \frac{\partial v_m}{\partial t} + E (v_t - v_p) = 0$$

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The choice of characterization of the distribution and of the associated form of the mass balance, eqs. 5 or 6, depends on the shape of the size distribution: equation 5 is more appropriate when the bed material is obtained by mixing few classes of monogranular sediments, while equation 6 appears more convenient when, for instance, the size distribution is lognormal.

Analysis of eq. 5 and 6 leads to the conclusions that, in a erosive process of armour layer formation, as in the present case ($E > 0$), the progressive coarsening of the mixing layer (i.e., $\dfrac{\partial D_m}{\partial t} > 0$) is associated to the condition:

$$D_t \; < \; D_p$$

and viceversa for the destruction process.

In the initial stage of armour process (i.e., when $D_m = D_p$), the development of armouring is automatically controlled by the selectivity of transport; consequently, $D_t < D_m$; as the armour condition develops, D_m becomes coarser than D_p until new equilibrium conditions are reached.

The following equilibrium conditions can be identified:

- STATIC ARMOUR, if $T = E = 0$;

- DYNAMIC ARMOUR, if $t_i = p_i$ and $E > 0$ (pavement according to Parker);

- SEMI-STATIC ARMOUR, if $E = 0$ and $T > 0$, that is, the coarser fractions are not mobile and the finer fractions, coming from the upstream reaches, are in motion; in other words, static armour for the coarser fractions coexisting with a dynamic armour of the finer fractions.

The case $E = 0$ with all classes in motion ($T \neq 0$; $t_i = p_i$) eliminates any dynamical connection between the surface processes (sediment transport and mixing layer dynamics) and the subsurface layer. It can be considered a particular condition of a dynamic armour (pavement) without any interaction with the sub-layer.

Following these considerations, the analysis of the formation of the armour layer has been made on the basis of the experimental data. Eq. 6 allows the analysis by assuming just one representative parameter, c, of the grain size distribution; for sediment 1, which is

approximately lognormally distributed, the mean diameter has been assumed as significative parameter; for sediment 2 (bimodal frequency distribution), the percentage of finer material (less than 2.8 mm) was considered while for sediment 3, showing a quite flat frequency distribution, the mean diameter plus the percentage of fine has been assumed.

In figures 5, 7, and 9 the temporal evolution of the representative parameters is shown for sediment 1,2 and 3 respectively.

Figures 6, 8 and 10 show the behaviour of sediment transport rate, referred to the maximum value, qsmax, obtained in each run, as function of time referred to the armouring time, td. run 3 has not been represented because it was partially fed with sediment material from upstream during the experiment.

Sediment 1: from figures 5, 6 it can be seen that the condition for a static armour, i.e., $E \to 0$, $T \to 0$, and $v_t < v_p$ is practically satisfied for all runs. It is worth to note that, at the very initial stage of armour formation, where $E > 0$, figure 3 shows that $v_t > v_p$; according to eq. 6 this should imply $\dfrac{\partial v_m}{\partial t} < 0$: results obtained from eq. 7 for run 1, where this phenomenon is more pronounced, seem to confirm the initial trend of the surface layer to become finer and this is probably due to the "artificial disposition" of bed material surface when it is initially placed in the flume at the beginning of each run.

Sediment 2:   with reference to figures 7,8 . a trend toward a regime of static armour can be observed in runs 6 ,7,and 8; particularly. run 6 was prolonged after the first armour destruction occurred 8 hours later the beginning of run: the process of armour formation was activated again generating a new static equilibrium: after 8 hours, a second destruction was imposed to the bed: again, the armour layer start to develop, until the end of the run.

Sediment 3: for this type of material, the pattern shown in figure 9, 10 is similar to the one obtained for sediment 1. In this case, the initial artificial sorting of the bed surface was less marked than sediment 1, while the temporal evolution appeared practically similar.
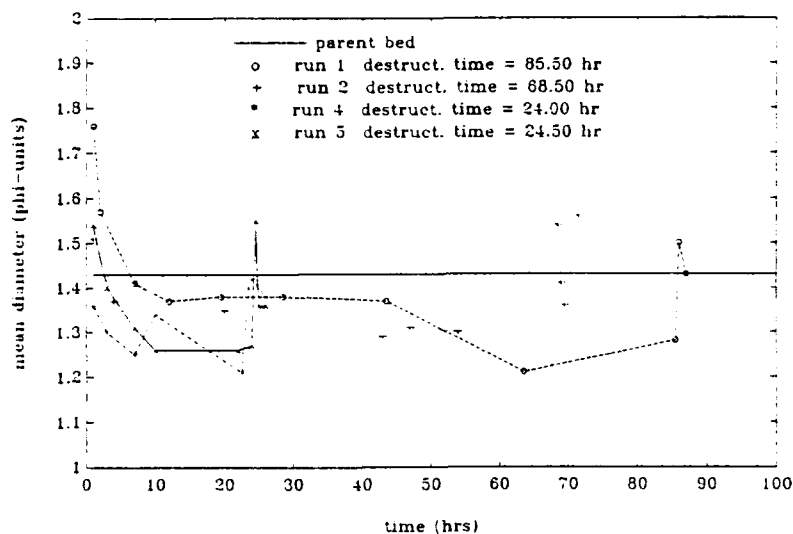
Fig 5 - Temporal evolution of the mean diameter for runs 1, 2, 4, 5.
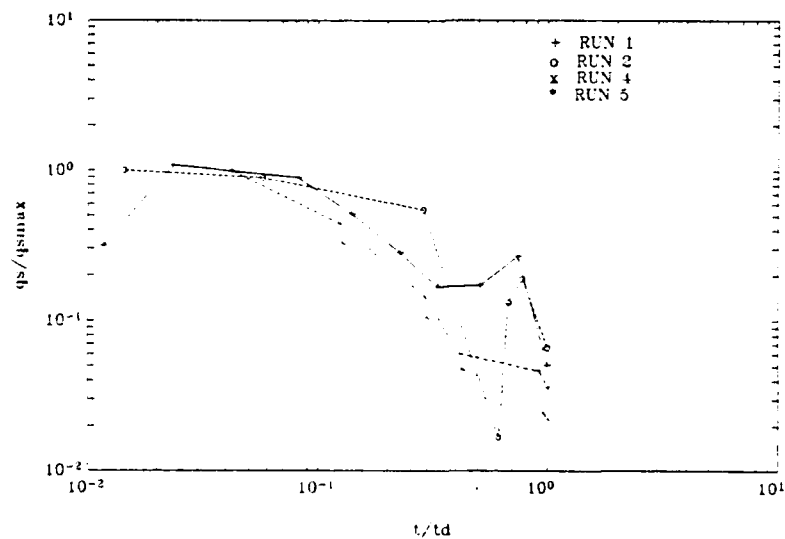
Fig. 6 - Sediment transport rate as a function of time for runs 1, 2, 4, 5 (sediment 1).
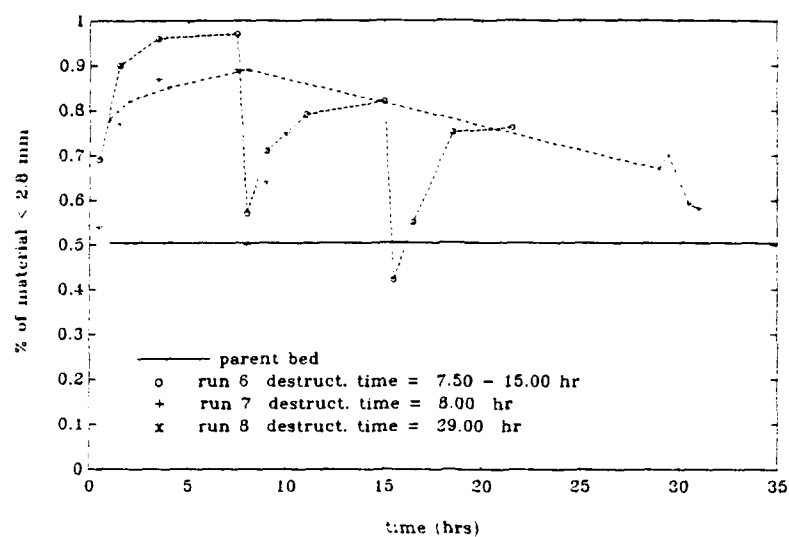
Fig. 7 - Percentage of fine material as a function of time for sediment 2.
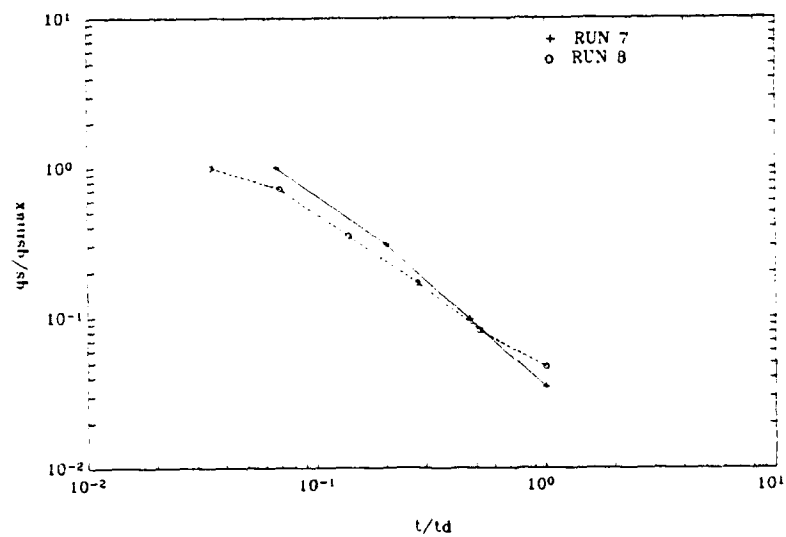
Fig. 8 - Sediment transport rate as a function of time for runs 7, 8 (sediment 2).

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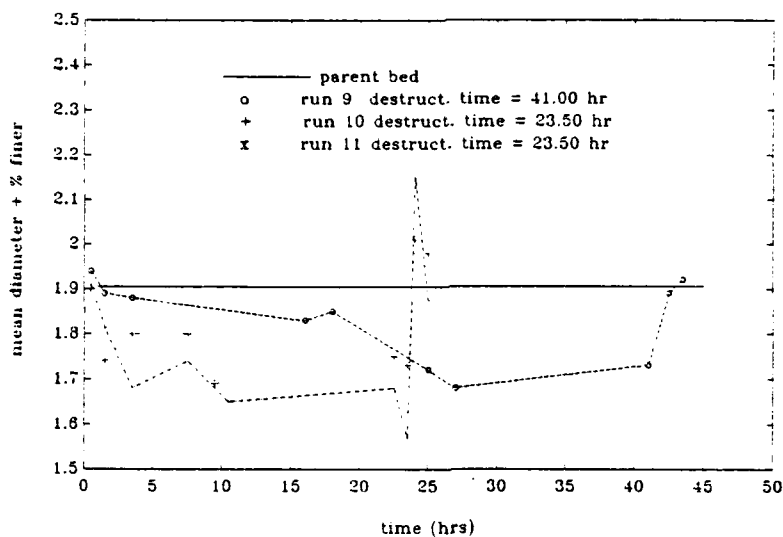
Fig. 9 - Temporal evolution of the representative parameter for sediment 3.
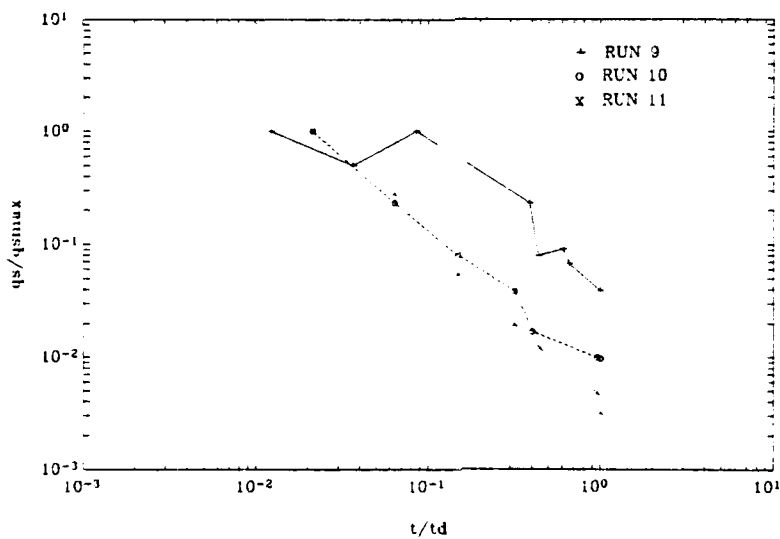
Fig. 10 - Sediment transport rate as a function of time for runs 9, 10, 11 (sediment 3).

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5 CONCLUSIONS

Experimental results obtained by a laboratory investigation on armour processes have been presented . The following conclusions can be drawn:

- analysis of the continuity equation applied to the mixing layer shows that the coarsening of the surface layer may be consequent either to a progressive reduction of the sediment transport rate reaching practically zero (static armour) or to a progressive equalization of the size fractions transported with fractions present in the parent bed material (dynamic armour); intermediate levels of the armour process may occur when some coarse fractions are immobile and the finer are in motion (semi-static armour). According to these definitions, the present experimental investigation has been carried out to analyze static armour conditions.

- the experimental results seem to confirm that the equilibrium condition of the armour layer is essentially controlled by the mechanism of sediment sorting which practically vanishes as the sediment transport tends to zero. However. an increase of the bed shear stress may reactivate the sediment transport and its selectivity and the process of the armour layer formation starts again to generate a new equilibrium condition. On the other hand. if the increase of bed shear stress is able to reduce considerably the selectivity of the transport. all the fractions would be in motion. leading to the general destruction of the armour layer.

- in the case of a dynamic armour. an increase of bed shear stress inevitably induces a reduction of the grain size distribution difference between parent bed material and surface layer leading at least to a partial destruction of the armour coat. as a consequence of the more marked reduction of the flow selective capacity with increasing shear stress. The selective capacity is here assumed to be the ratio of the transport capacity of the finer fraction to the coarser ones.

- the experimental data have confirmed the validity of assuming a constant thickness of mixing layer during the armouring process equal to the D_{50} of the armour layer in representing the temporal evolution of bed surface grain sizes.

- photograph technique has been proved to be a reliable method to sample bed surface composition without altering bed structure during the experimental run.

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SPATIAL VARIATION IN ARMOURING IN A CHANNEL
WITH HIGH SEDIMENT SUPPLY

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ABSTRACT: Recent advances in the understanding of the origin and function of armouring of gravel-bed rivers have not addressed the role of nonuniformity and unsteadiness of flow. These leave a strong imprint on both surface and subsurface bed material sizes observed at low flow, from which we draw interpretations of bedload transport at high flow.

Mean bed armouring, measured as the difference in grain size of surface and subsurface bed material, is low in a channel in which recent and ongoing bed aggradation indicates that sediment supply has exceeded transport capacity. Alternate bar topography induces strong variations in boundary shear stress especially at low flow. Resulting winnowing of fine sediment from zones of high stress and deposition in zones of low stress produce wide variations in armouring. Bed areas with coarse surfaces are strongly armoured; fine areas are unarmoured and commonly consist of a thin layer of fine sediment overlying coarser material.

Spatial variation in armouring and boundary shear stress can lead to apparent size-selectivity in bedload transport in subsequent rising stages. The relatively fine bedload transported during early rising stages can originate from unarmoured deposits of fines, where all particle sizes present on the bed surface may be nearly equally mobile. These surfaces are devoid of larger particles, however, that appear in bedload only when coarser armoured surfaces become more widely entrained.

INTRODUCTION

In recent years there have been considerable progress and some controversy over the origin and function of armouring of gravel riverbeds. Parker and Klingeman (1982) and Andrews (1983) stimulated new attention on the problem by advancing the hypothesis of approximate equal mobility of bed particles represented on the surface of a heterogeneous gravel bed. Accordingly, a coarse surface layer is a necessary condition of equilibrium transport of heterogeneous sediment under moderate excess boundary shear stress. Large particles, which inherently resist entrainment as uniform bed material more than small particles, are overrepresented on the bed surface, so that the mobility of all grain sizes tends to be equal. Debate continues on whether or not the degree of size selection observed in sediments entrained from mixed-size bed material justifies an approximation of equal mobility (Komar and Li, 1988). Nevertheless, experiments (Kuhnle and Southard, 1988), field data (Milhous, 1973; Andrews and Erman, 1986) and theory (Wiberg and Smith, 1987) indicate that the mobility of particles of different size on a coarse surface layer are much closer in mobility than they would be if each size composed a uniform bed, and that the equal mobility approximation is reasonably accurate in many cases.

The bed surface layer ceases to be coarser than the subsurface when the transport of all sizes present is intensive. Parker and Klingeman (1982) predicted that surface coarsening would disappear at boundary shear stresses

greatly exceeding the entrainment threshold of the median particle size, or specifically, at a value of $tau* = tau/g(ps-p)D50 = 0.26$. Wilcock and Southard (1989) found it to occur at $tau* = 0.07$ in a flume experiment employing a recirculating sediment system (transport rate dependent on flow).

Dietrich et al (1989) demonstrate that intensive bedload transport can also diminish surface coarsening under moderate stresses when a high sediment load is imposed. In their experiment, mixed-size sediment was fed at a high rate into a flume containing the same size mixture as the feed. At a value of $tau*$ of only 0.087, which is well exceeded at bankfull stage in many gravel-bed rivers, bed armouring disappeared. But as the feed rate was reduced in two steps to one-tenth the original while boundary shear stress was held approximately constant, the surface coarsened until nearly the entire bed was armoured. At equilibrium, the size mixture of bedload necessarily equalled that of the feed at all feed rates. Dietrich et al (1989) quantify the adjustment of surface coarseness to load as $q*$, the ratio of bedload transport computed from a predictive equation using bed surface particle size to that computed using bedload or subsurface particle size. Qualitative correlations of $q*$ with sediment load in six stream channels in California (Kinerson and Dietrich, 1989) support this hypothesis.

The interpretation of the role of the coarse surface layer in sediment transport of Dietrich et al (1989) is somewhat different than that of Parker and his associates. The coarseness of the surface of a streambed offers resistance to the entrainment of bed material. If transport capacity exceeds the sediment load supplied from upstream, the bed surface coarsens, thereby increasing resistance to sediment transport. This interpretation is not in conflict witn the equal mobility approximation.

The interpretation of armouring processes observed in flumes to that occurring in rivers is limited by necessarily choosing between sediment systems that impose either complete independence or dependence of transport rates. If transport rate is considered a dependent variable (mimicking sediment recirculation), then degree of surface coarsening in nature becomes a function of boundary shear stress and so varies little between gravel-bed rivers; if transport rate is an independent variable (mimicking sediment feed), coarsening is independent of stresses exceeding the entrainment threshold of the bed.

Bedload transport in natural channels can be either more nearly independent or dependent on hydraulic forces, depending on scales of time and space considered (Wilcock and Southard. 1989). Over a period of decades or centuries, a river conveys the sediment load imposed by its basin and adjusts accordingly. Most bedload is typically stored in the bed itself, however, which thus constitutes tne supply, and so sediment transport through a reach of river over a period of hours or days depends more closely on hydraulic forces than on supply. That sediment supply still affects armouring at this scale is suggested by Kinerson and Dietrich's correlation of armouring with supply in rivers. At still shorter spatial scales, sediment supply to a local area of bed may be augmented or depleted by changes in sediment transport direction or by bed scour or fill occurring upstream under different hydraulic or sedimentological conditions, and therefore appear independent of local conditions.

Before controversies over the applications of various models for the causes and functions of bed surface armouring can be resolved, the effects of nonuniform and varied flow, which are definitive aspects of natural channels, must be determined. We usually observe the texture of channel beds when much of the surface is out of water or at least visible, and do not understand very well how waning flows have modified the bed since significant transport took place. As the flow varies over a channel with nonuniform topography, such as that typically produced by bars, changes occur locally in sediment transport direction and the magnitude of boundary shear stress relative to that over other areas of the channel. Some parts of the channel continue to supply sediment to transport while others become sites of deposition. All of these changes must bear on the process of armouring and add complexity to bed surface texture. The degree of armouring, for example, has been shown to vary widely between local areas of a streambed (Mosley and Tindale, 1985· Maloy, 1988; Church et al, 1989).

The purpose of the study described below was to investigate spatial variations in degree of armouring and their relation to the distribution of boundary shear stress. We selected a channel with a large in-channel supply of bedload because we expected variations in armouring to be high and most of the bed surface to have been active during recent high flows.

REDWOOD CREEK

Redwood Creek drains a 720 km^2 basin in north coastal California, USA (Fig. 1). Streamflow, sediment supply and transport rate, channel change, and land use history are well documented (Janda, 1978). The basin receives an average of 2000 mm of precipitation annually, most of which falls as rain between October and March. Total basin relief is 1615 m; average hillslope gradient is 26 percent. For much of its 108-km length Redwood Creek flows along the trace of the Grogan Fault, which juxtaposes two distinct bedrock types. The east side of the basin is generally underlain by unmetamorphosed sandstones and siltsones of Mesozoic Franciscan Assemblage, whereas the western side is predominantly underlain by a quartz-mica schist.
Both rock units are deformed by numerous fractures and shear zones. The rock incompetence due to this deformation coupled with high rainfall and steep terrain contribute to the high erodibility of the watershed.

Water discharge records (since 1955) and sediment discharge records (since 1970) are available for five U.S. Geological Survey (USGS) gaging stations in Redwood Creek (Fig. 1). Sediment discharge is strongly flow-dependent; a large proportion of the annual sediment load is transported in infrequent, high magnitude flows. The estimated long-term sediment yield (suspended sediment and bedload) from 1954-1980 was 2480 t/km2/yr at South Park Boundary and 2100 t/km2/yr at Orick (USGS, written communication, 1981). Bedload constitutes 10-30% of the clastic load. Since 1980 sediment loads have decreased to 1190 and 920 t/km2/yr at the two stations respectively, primarily due to mild winter storm seasons during the last decade. These sediment yield rates are still higher than those from other rivers in the U.S that do not drain active volcanoes or glaciers.

Early aerial photographs taken in 1936 and 1947 show that the basin was covered with old-growth redwood and Douglas fir forests and a few areas of grassland. Redwood Creek was narrow and sinuous in most reaches, and bordered

by a thick canopy of trees over much of its length. Very little logging or
road construction had occurred by 1947.

Timber harvest in the Redwood Creek basin began in earnest in the
early-1950's. and by 1978 81 percent of the old-growth coniferous forest had
been logged (Best, 1984), and thousands of kilometers of logging roads had
been built. Erosion rates during the per'-d of accelerated timber harvest
(Janda (1978) were about 7.5 times greater than the natural rate estimated by
Anderson (1979).

Channel-bed aggradation and degradation

Widespread channel changes have occurred since 1964 in response to large
floods and a destabilized landscape. Large floods occurred in the Redwood
Creek basin in 1861, 1890, 1953, 1955, 1964, 1972, and 1975, but accelerated
erosion and channel response was not substantial before the flood of 1964.
From 1954 to 1980, a total of 30.5×10^6 tonnes entered Redwood Creek mostly
from streamside landslides (debris slides, streambank failures, forested block
slides, and earthflows), and fluvial erosion originating on unpaved logging
roads (gullies, stream diversions, failed stream crossings) (Kelsey et al,
1981; Weaver et al, in press). Most of the sediment was input during the
floods of 1964, 1972, and 1975. The 1964 flood was especially damaging and
caused streamside landsliding, widespread aggradation (up to 7 m) and channel
widening in Redwood Creek (Madej, in press), even though the peak flow was not
unusually high (recurrence interval of 50 years; Coghlan, 1984). Floods of
1972 and 1975 resulted in additional aggradation in the downstream third of
Redwood Creek.

Of the total sediment input during this period, 31% was deposited on
floodplains and in the Redwood Creek channel; little sediment was stored on
the steep hillslopes of the basin. Channel-stored sediment has become an
on-going source of sediment to downstream reaches. For example, from 1964 to
1980, 1.8×10^6 t of sediment was eroded from the channel bed in the upstream
two-thirds of Redwood Creek, and much of that sediment was redeposited in the
downstream lower gradient reaches of Redwood Creek (Madej, in press). Annual
surveys since 1973 of 60 channel cross profiles snow that the 1975 flood
(recurrence interval = 25 years) caused the upstream third of the channel to
degrade as much as 1.3 m while the downstream-most 25 km aggraded as much as
1.5 m (Varnum and Ozaki. 1986).   Since 1980, surveys show that in Upper
Redwood Creek the stream bed has stabilized at its pre-1964 level, the channel
bed in the middle reach continues to erode, and the downstream 16 km of
Redwood Creek either continues to aggrade or remains at an elevated level.
Thus effects of high sediment input upstream will have been felt downstream
for decades.

Study reaches

We selected two study reaches, one showing recent degradation (Fig. 2a), and
the other, 12 km farther downstream, showing recent aggradation (Fig. 2b).
Although the upstream reach is actively degrading, it still carries a high
sediment load and may not yet have scoured down to its pre-1964 level. Both
study sites are straight alluvial reaches at least 15 channel-widths in length
and have similar channel characteristics (Table 1). Ea . study reach begins

and ends at riffle crests, and has well developed alternate bars and pool-riffle sequences.

Table 1. Characteristics of Study Reaches

Reach:	Degraded	Aggraded
Drainage area (km^2)	523	605
Bankfull discharge (cms)	370	430
Bankfull width	60	110
Bankfull depth (m)	2.2	1.9
Channel gradient (m/m)	0.0026	0.0014
Length (m)	1284	1670
Sinuosity	1.09	1.03
D50-bed surface (mm)	22	15

Gaging stations lie in or near each study reach. The Wier Creek gaging station in located in the middle of the upstream study reach. The Elam Creek (aggraded) reach lies between two gaging stations (Miller Creek station 5 km upstream and the Orick station 4 km downstream). The Miller Creek gaging station is at a transition point between degrading reaches upstream and aggrading reaches downstream. From 1973-1983, the U. S. Geological Survey periodically collected bedload samples at the Miller and Wier Creek gaging stations with a Helley-Smith bedload sampler. Flows up to a recurrence interval of five years were sampled.

Field work commenced in 1988 in the aggraded reach and in 1989 in the degraded reach. Peak flows measured in Redwood Creek at Orick for 1988 and 1989 were 431 and 606 cms, respectively (recurrence intervals of 1.3 and 1.7 years). These moderate flows caused minor channel modifications, but were still capable of transporting considerable bedload. Bedload measured at the Orick gaging station was similar in the two years, 56,500 t in 1988 and 55,300 t in 1989.

METHODS

Mapping of surface grain size and channel topography

The channel bed in both study reaches showed distinct spatial variation in bed surface texture. In order to map the spatial distribution of bed surface grain size and to efficiently sample bed material, we stratified the bed into recognizable areas whose bed-surface grain size composition fell into certain predetermined grain-size ranges, which were arbitrarily chosen to represent common grain size ranges observed on the bed surface of each reach. Individual mapped areas are referred to as facies; each facies represents one of three or four facies types defined by a range of grain size. We delineated facies boundaries based on a visual estimate of the D75 of the surface particles, and then stratified individual facies by facies type for sampling purposes. In the aggraded reach the definitions of facies types were:

D75 < 22mm = fine pebble facies
22 mm < D75 < 64 mm = coarse pebble facies
D75 > 64 mm = cobble facies

In the degraded reach, the same definitions were used with the addition of a fourth facies type -- bimodal sand -- whose bed surface was covered with >25% sand intermixed with coarse pebbles or cobbles.

Facies were mapped over each reach by staking out and surveying boundaries between facies. In most cases the boundaries between facies were distinct, and mappers independently delineated the same boundaries. Topographic maps were constructed by using elevation data from facies boundaries and surveying miscellaneous elevations (aggraded reach) or by surveying a series of cross sections spaced at intervals of one- to two-thirds channel width (degraded reach). Surveys encompasssed the channel banks to at least bankfull height, and covered all significant topographic features of the channel bed.

We selected individual facies to sample by probability proportional to size (in this case, facies area) without replacement. The number of facies sampled was proportional to the total area of the facies type relative to the total channel area (Table 2). We gave slightly more weight to sampling the cobble facies type because its size distribution was more variable than the other types. Total areas of each facies type were similar between the two reaches. The coarse pebble facies, which differed most in area, would have been more similar if bimodal facies in the aggraded reach had been classified separately instead of, in most cases, included in coarse pebble facies. Facies types were well represented in both reaches. The bimodal facies type in the degraded reach had the smallest proportion of bed area (9.3%); the coarse pebble type in the aggraded reach had the largest (52.9%).

Table 2. Number of sampling units in study reaches

Reach:	Degraded		Aggraded	
Facies type	# of units	% of total area	# of units	% of total area
Fine pebble	6	27.5	16	30.7
Coarse pebble	13	43.4	19	52.9
Cobble	8	19.8	15	16.5
Bimodal sand	5	9.3	-	-
Total	32		50	

Measurements of particle size

Surface and subsurface bed material in each selected facies were sampled systematically. Over each facies, we selected 100 to 150 particles from paced grids (Wolman, 1954), and determined the sieve size range of each particle by passing it through a template with square holes whose sizes ranged at half-phi intervals down to 4 mm. The size distribution of material <4mm on the bed surface material was estimated using the size distribution <4mm of sampled subsurface material (described below).

Samples of subsurface material from each sample facies were compiled from nine to twelve subsamples taken from points on rectangular grids set up using random starts. Subsamples were taken instead of single large samples because a pilot study had shown that the median grain size of one large sample had a higher variance than that of several smaller samples of equal total weight

taken from the same facies (Lisle, Madej, Hilton in preparation). Each
compiled subsurface sample totalled about 100 kg. This sample volume was
deemed adequate to give reproducible results based on the criteria of Church
et al (1987) that the largest particle, which usually did not exceed 90 mm in
this case, would comprise about 1% of the sample weight.

At each subsample location, we removed the surface layer with a flat-bottomed
shovel to a depth equal to the D90 size of the surface, although for fine
pebble facies we removed about 2 cm regardless of surface grain size. We then
collected a subsample of about 10 kg of subsurface material from a layer about
twice as thick as that removed from the bed surface. Subsamples from areas
under water were collected from inside a 30-cm-diameter cylinder that had been
worked vertically into the bed (McNeil and Ahnell, 1964). Approximately 20%
of all subsamples were taken under water.

Samples were dried on tarps and field sieved into size classes at half-phi
intervals down to 11 mm, and weighed. The fraction <11 mm was split, and a
6-8 kg subsample was brought back to a laboratory to be sieved at half-phi
size intervals down to 1 mm. The fraction <1 mm was disregarded because
suspended sediment samples (U.S. Geological Survey, 1970-88) indicated that 1
mm was the approximate upper size limit of suspended sediment, which we
assumed did not play a role in armouring processes. Samples taken under water
were wet sieved in the field down to 11 mm, the finer fraction was split and
retained for dry sieving in the laboratory, and wash water containing
suspended material was discarded.

Data Analysis

Particle size distributions of surface and subsurface material, expressed as
the percentage by weight falling within each size class, were computed for
each sampled facies. Sampling of surface and subsurface material by different
methods may lead to non-equivalent grain size distributions. Using the formal
conversion factors of Kellerhals and Bray (1971), we assumed that grain size
distributions from grid-by-number measurements (pebble counts) are equivalent
to those from sieve-by-weight measurements (subsurface material). This is
supported by an empirical test by Church et al (1987) using a sample of 200
stones.

From these distributions, values of D50 were computed for each sampled facies,
then each facies type, and finally for the reach as a whole, weighted by area
in each step. Similarly, the distributions for each facies were weighted by
area to compute mean percentages of each size class for each type and
similarly for each study reach, using a method suggested by Baldwin (U.S.
Forest Service, personal communication). Values of D50 for surface and
subsurface material and D50(surface)/D50(subsurface), hereafter referred to as
the D50 ratio, were computed for each facies type and a mean weighted average
determined for the study reach.

RESULTS

Spatial distribution of facies types

The coarseness of the bed surface correlated commonly with the qualitative
magnitude of boundary shear stress at moderate flow when the bed surface was

last active. This was most apparent in the part of the channel submerged at
low flow. Cobble facies commonly occurred in zones of high or
downstream-increasing boundary shear stress at low flow, such as over riffles
or upstream portions of bar surfaces where streamlines converged toward pools
(fig. 3). Fine pebble facies occurred in zones of low or
downstream-decreasing shear stress at low flow, such as pools, downstream
portions of bars, and along streambanks.

Size distributions of facies types

Mean surface and subsurface grain size distributions for the reaches as a
whole and for each facies type showed distinct patterns (fig. 4). All grain
size distributions (except surface sizes of bimodal facies) were unimodal,
spanning sizes from sand to large cobbles. All sizes present in surface
material of each facies type were also present in their respective subsurface
material. Subsurface grain size distributions were similar between the
different facies, particularly near the coarse end of the size spectrum.

Coarser facies were more strongly armoured than finer facies. With increasing
coarseness of facies types, subsurface grain size distributions became
slightly coarser, but not as much so as the surface grain size distributions.
Similar results in another channel are reported by Maloy (1988). As
quantified by D50 ratios (Table 3), cobble facies in Redwood Creek were most
strongly armoured; coarse pebble facies exhibited an intermediate degree of
armouring approximately equal to that of the channel as a whole. Fine pebble
facies in both reaches were unarmored as were bimodal facies in the aggraded
reach.

Table 3. D50 of surface and subsurface material and D50 ratios and their
respective standard errors (s.e.) for study reaches of Redwood Creek.

Reach:	Degraded			Aggraded		
	D50(s.e.)		D50ratio(s.e.)	D50(s.e.)		D50ratio(s.e.)
Facies type	Surface	Subsur.		Surface	Subsur.	
Fine pebble	5.7(1.0)	10.5(1.6)	0.68(0.14)	6.1(0.7)	6.2(0.5)	1.05(0.14)
Coarse pebble	26.2(1.5)	19.2(1.1)	1.44(0.06)	14.9(0.6)	9.7(0.4)	1.60(0.09)
Cobble	38.2(2.6)	24.4(2.6)	1.66(0.19)	30.0(1.3)	13.0(0.6)	2.45(0.16)
Bimodal Sand	18.4(4.1)	22.0(3.1)	0.96(0.19)	--	--	--
Weighted Ave.	22.2(0.9)	18.1(0.9)	1.23(0.06)	14.7(0.4)	9.1(0.3)	1.57(0.07)

Fine pebble facies commonly appeared to be, in fact, "anti-armoured", that is,
having a surface layer finer than the subsurface layer. For example, the D50
ratio averaged for the fine pebble facies of the degraded reach (0.68) was
well less than unity. In this reach, the thickest and most extensive fine
pebble facies consisted of sheets of sand and small pebbles less than 0.5 m
thick that mantled the beds of pools (fig. 5). Deep water in much of these
areas prevented taking subsurface samples. Downstream, as the water became
shallow enough to obtain samples, the fine deposits thinned and graded into
bimodal facies, which graded into coarse pebble or cobble facies as large
particles underlying fine sediment were increasingly exposed. Subsurface

samples were thus biased toward thin areas of fine pebble facies, which were likely to be anti-armoured.

Fine pebble facies in the aggraded reach were more widely distributed and were commonly associated with unit bars, which will be described later. Fine facies in pools contained abundant pebbles. The thickness of these deposits could not be probed because they were more difficult to penetrate than the sandy deposits in pools of the degraded reach, and any underlying coarser layer could not be detected. Comparison of topographic surveys done one year apart showed that the new deposits were locally as much as 1 m thick.

Anti-armouring is somewhat an artifact of our method to sample the subsurface. In scraping away the surface of fine pebble facies that were thin, we essentially removed those facies -- that is, the bedload material that was last transported in that location -- and sampled coarser, underlying material. Considering difficulties in obtaining subsurface samples unbiased by water depth and uncertainties in interpreting anti-armouring, it is perhaps prudent to assume that fine facies have D50 ratios no less than unity.

Comparison between reaches

An unexpected result was that the bed of the aggraded reach, whose transport capacity was demonstrably exceeded, was apparently more strongly armoured on average than that of the degraded reach. A possible explanation is that a greater instability of the aggraded reach apparently led to stronger apparent armouring of its coarsest facies. The greatest difference in D50 ratios of facies types between the two reaches was the greater armouring of the cobble facies of the aggraded reach (D50 ratio of 2.45 vs. 1.66). The difference in D50 ratios arose primarily from the superior coarseness of subsurface material of the cobble facies in the degraded reach (Table 3, fig. 4). Although both surface and subsurface material of cobble facies of the degraded reach were coarser than those of the aggraded reach, subsurface materials showed the greatest contrast, and that of the aggraded reach had a pronounced mode at proximately 64mm.

A possible origin for the difference in D50 ratios is that the subsurface layer of cobble facies in the aggraded reach may have been locally thinner (in particle diameters) than in the degraded reach. If so, there would be a greater likelihood in the aggraded reach that subsurface samples of cobble facies would incorporate underlying material from finer facies. In contrast, cobbles in the degraded reach perhaps accumulated in the subsurface as the buried frameworks of armour layers created under consistent sedimentologic conditions that favored accumulation of the coarsest particles in transport. Coarsening of subsurface bed material in a degrading channel with alternate bars was observed in a laboratory experiment (Iseya et al, 1989), in which a constant, poorly graded size distribution of transported bedload was maintained. Sequential aerial photograph of the degraded reach show that its large-scale alternate bars have been essentially stationary over the last decade and smaller-scale unit bars have not been observed in the field.

In contrast, aggradation and unit bar migration in the aggraded reach apparently often destabilized local sediment transport conditions under which coarse particles could accumulate to any thickness. Resurveying of bed topography of the aggraded reach after one rainy season revealed that

migrating unit bars with average heights of 0.2 m had significantly altered channel morphology (fig. 6). These bars, similar to those described by Bluck (1982) in a braided channel, covered approximately one-fourth of the channel and had accreted onto the larger scale alternate bars. Deposition of unit bars accounted for most of the aggradation at least over that season. Particularly in the portion of the channel submerged at low flow, the new deposits had become so modified after becoming annealed to the major alternate bar topography as to become indistinct as unit bars.

DISCUSSION AND CONCLUSIONS

Comparison of armouring with that of other channels

In comparison with other single-thread, gravel-bed rivers, the bed of Redwood Creek is weakly armoured (fig. 7), that is, the bed surface is only slightly coarser than the subsurface. Although data on sediment loads of the other channels is not available, we assume that, per basin area, most are smaller than that of Redwood Creek. Redwood Creek's sediment transport capacity has clearly been equalled or exceeded by the supply of bedload delivered to its channel, as indicated by its aggraded channel. Comparison of armouring must be made with caution because bed samples for the other data sets were fewer and were located nonrandomly. The D50 ratio of most of the other channels exceeds a value of 2; the reach-averaged values for the aggraded and degraded reaches of Redwood Creek are 1.57 and 1.23, respectively. Material less than 1mm in diameter was not included in calculating D50 ratios for Redwood Creek. If all sizes are included in order to more fairly compare with the other channels, D50 ratios for Redwood Creek equal 1.77 and 1.39 for the aggraded and degraded reaches respectively. Thus the conclusion that the bed of Redwood Creek is weakly armoured remains valid.

The fact that the bed is weakly armoured provides further evidence that armouring can adjust to sediment supply and thereby alter bed resistance to sediment transport (Dietrich et al, 1989). This evidence is given additional weight by the intensity of measurement of armouring, which to the best of our knowledge has not been exceeded elsewhere.

We speculate that mean bed armouring in both reaches of Redwood Creek has decreased to a minimum, and that deposition of bedload sediment in the form of unit bars in the aggraded reach manifests a further adjustment to excess sediment load not represented in the degraded reach. The aggraded reach, whose transport capacity was clearly exceeded, was significantly armoured (D50 ratio = 1.6, s.e. = 0.07) and more so than the degraded reach (D50 ratio = 1.2; s.e. = 0.06), which apparently had a smaller relative sediment supply. Mean bed armouring, therefore, is not a simple function of sediment supply. As described earlier, superior channel stability of the degraded reach has allowed coarse particles to accumulate to a greater thickness in cobble facies of that reach than in those of the aggraded reach, resulting in a smaller difference between surface and subsurface material in the cobble facies of the degraded reach and in the bed of the reach overall.

Moreover, the difference in mean bed armouring between study reaches is minor compared to that between Redwood Creek and other channels that presumably have a much smaller relative supply of sediment. Although trends in bed elevation of the aggraded and degraded reaches show categorically different responses to

changing sediment supply, the contrast in supply relative to transport capacity between the two reaches is not strong. The degraded reach remains aggraded after floods of the past decades and is supplied with large volumes of sediment from channel erosion upstream (fig. 8).

Importance of winnowing

Winnowing and selective accumulation of coarse particles on the bed surface can both be responsible for armour formation. Grain size distributions of facies types in Redwood Creek give evidence that winnowing leaves a strong imprint on the bed surface when sediment transport has ceased. Differences in bed material indicate some selectivity in sizes deposited. All facies types, however, exhibit similarities between the coarse limb of the frequency distribution of surface and subsurface layers. This suggests that the surface layer of the armoured facies was depleted in fines, not enriched in coarse particles. In contrast, the surface of the bed classified as the finest facies type was enriched in fines and commonly consisted of a veneer of fine material overlying a coarser bed. Together, these suggest that winnowing of fines from coarse facies and deposition downstream to form fine-grained facies was largely responsible for spatial variations in armouring observed at low flow. The coincidence of high boundary shear stress at low flow with coarse (winnowed) facies and low shear stress with fine-enriched facies is consistent with this interpretation.

It is important to distinguish between vertical and downstream winnowing (Parker and Klingeman, 1982). In vertical winnowing, fines are lost to the subsurface layer; in downstream winnowing fines are carried downstream. As Gomez (1984) points out, vertical winnowing requires that the coarse particles in the surface layer be mobile; downstream winnowing requires that they be temporarily immobile. The juxtapositon of coarse and fine facies indicates that downstream winnowing of coarse armour layers in Redwood Creek was prevalent, at least during waning stages.

The size composition and location of different facies in Redwood Creek indicate that a strong imprint on the texture of the surface, including mean bed armouring, has been left by wide spatial variation of sediment sorting processes in response to varying flow. The bed at low flow, when it is practical to observe armouring of the entire channel, is to a large degree the product of waning flows. Although low flows are too feeble to transport important volumes of bedload, spatial variations in shear stress near entrainment thresholds enhance heterogeneity in bed texture through size-selective transport and deposition. As general transport ceases, fine sediment is eroded from vulnerable hiding places among the large particles on the bed surface, especially where shear stress remains relatively high. Fine sediment is thereby winnowed from zones of high shear stress and deposited in zones of low shear stress. Thus some areas of the bed of any channel in which armouring is uniformly weak at important bedload-transporting stages can be expected to exhibit strong armouring at low flow. Coupled winnowing and deposition driven by bar topography may affect observed mean bed armouring. One must therefore use caution in interpreting bedload transport processes at high stages from armouring and apparent bed mobility observed at low stages.

Equal mobility vs size-selective entrainment

Spatial distribution of surface particle sizes and boundary shear stress at low flow also has bearing on the relative mobility of different grain sizes represented in the bed of Redwood Creek. At relatively low discharges (no greater than 17 cumecs) we have observed sand and fine gravel transported, commonly as migrating dunes over an armour layer. Bedload samples at high flow (USGS, 1970-88) contain the largest particles that can fit into a Helley-Smith bedload sampler with a 3-inch orifice. Comparisons of the overall grain size distribution of bed material with bedload samples over a range of flows, therefore, would lead one to conclude that there is strong stage-dependent, selective transport by grain size.

Extrapolation of such results to the channe as a whole is erroneous, however. The most mobile sources of bedload at discharges when only sand and granules are transported are fine-grained winnowed sediment deposited in pools during the previous waning stages. These areas are the last to be deposited upon and the first to be entrained. All particle sizes on these areas may be nearly equally mobile; gravel and cobbles are simply not represented on the bed surface. This fine bedload is imposed on intervening coarser areas and overpasses a stable armour layer as streaks of moving material that disappear as stage drops and the bed is winnowed. Thus the degree of size selection in entrainment and transport depends on the part of the bed observed. At rising stages bedload may be first entrained from fine areas that exhibit equal mobility. The apparent size selectivity of entrainment on the bed taken as a whole, in this case, arises not so much from selective entrainment of particles from a mobile armour layer but from areal sorting that is induced by topographically driven variations in boundary shear stress and sediment transport direction.

ACKNOWLEDGMENTS

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Figure Captions

1. Location map showing Redwood Creek, aggraded and degraded study reaches, and five gaging stations.

2a. Cross section 20 of Redwood Creek showing a general lowering of the gravel bar surface and thalweg from 1973 to 1989. This scour is typical of the degraded reach.

2b. Cross section 6 of Redwood Creek showing both the gravel bar surface and thalweg at higher elevations in 1988 than in 1973. This pattern of elevation channel beds is typical of the aggraded reach.

3. Channel bed topography and facies of the degraded and aggraded study reaches, Redwood Creek. Contour interval is 0.5 m.

4. Grain size distributions for surface and subsurface bed material averaged for facies types and for study reaches, Redwood Creek.

5. Longitudinal profile of the water surface, bed surface, and armour layer buried under fine sediment in a pool of the degraded reach, Redwood Creek. The survey was conducted during low flow (31/8/89).

6. Map of the upstream half of aggraded reach showing channel areas that had aggraded more than 0.2 m from 1988 to 1989. Most of these areas were associated with discernible unit bars superimposed on the existing channel morphology.

7. Median grain size of subsurface bed material vs median grain size of surface bed material for gravel-bed rivers including Redwood Creek. The line demarks a ratio of 2.

8. Cross section 23, located 2 km upstream of the degraded reach, showing extensive erosion of left bank flood deposits from 1973 to 1989. As Redwood Creek erodes channel-stored sediment, the channel bed and banks become an important source of sediment to downstream reaches.

Figure 1
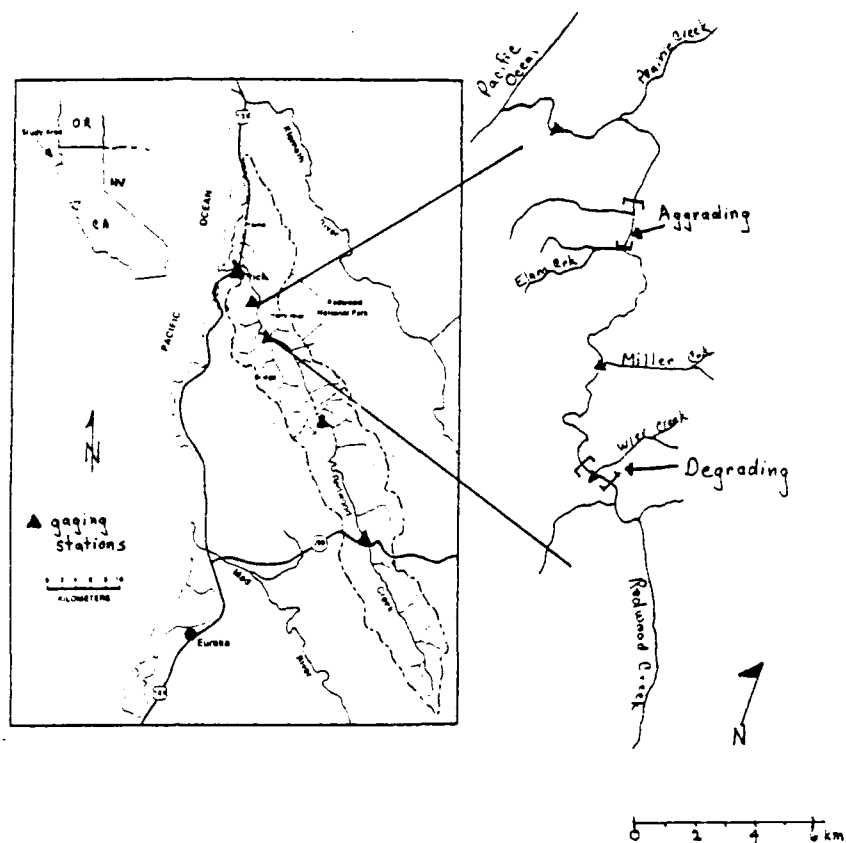
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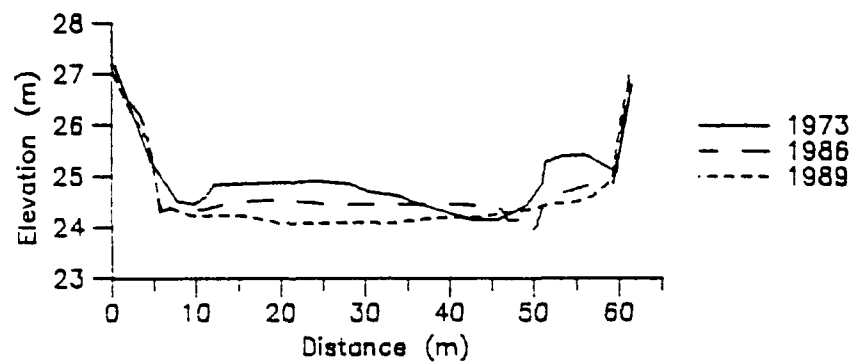
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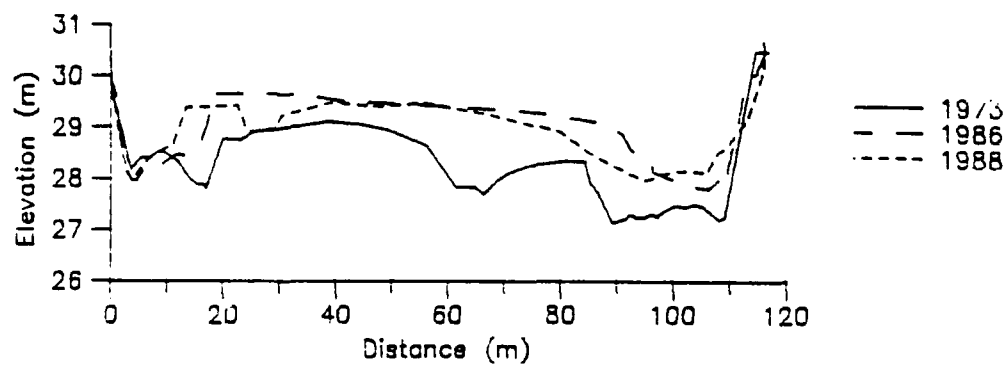
Aggraded Reach
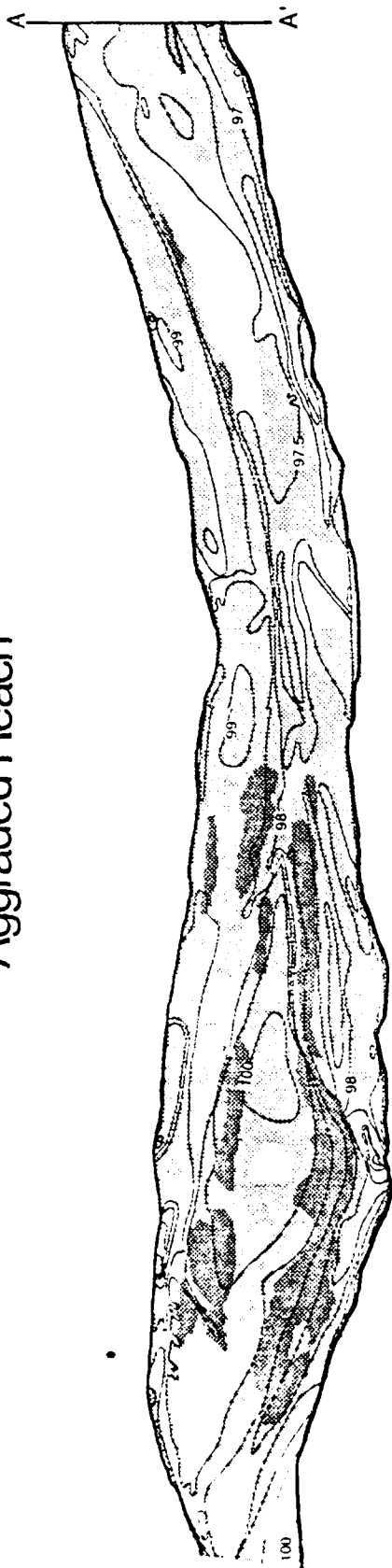
fine pebble

coarse pebble

cobble

bimodal sand

woody debris

unclassified

N

0 50 100 150 200 250

Meters
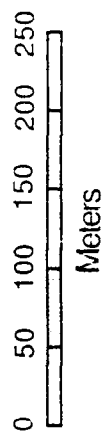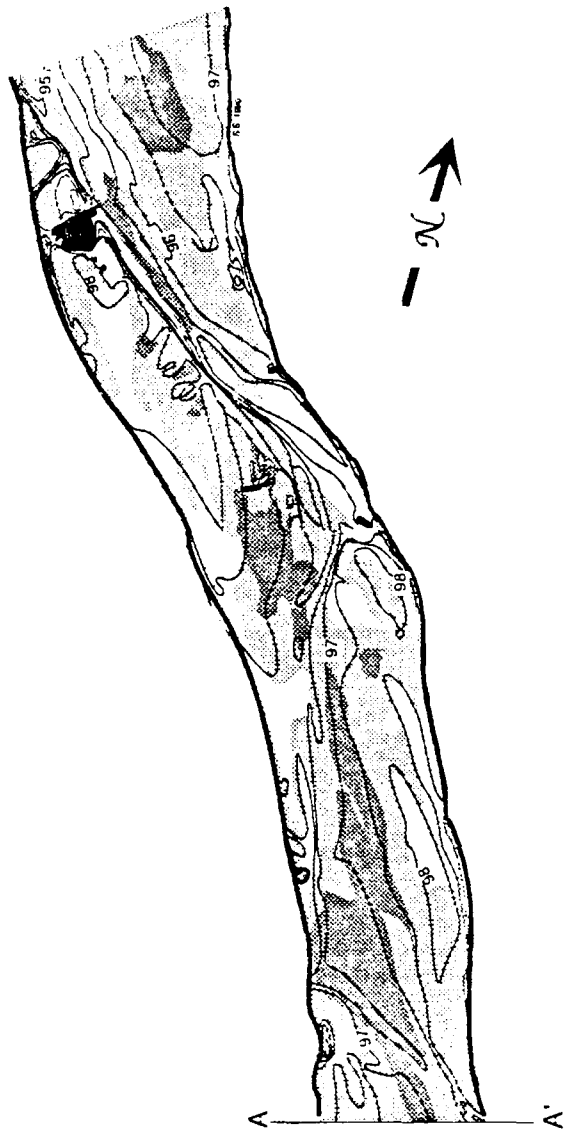
Figure 3

Degraded Reach
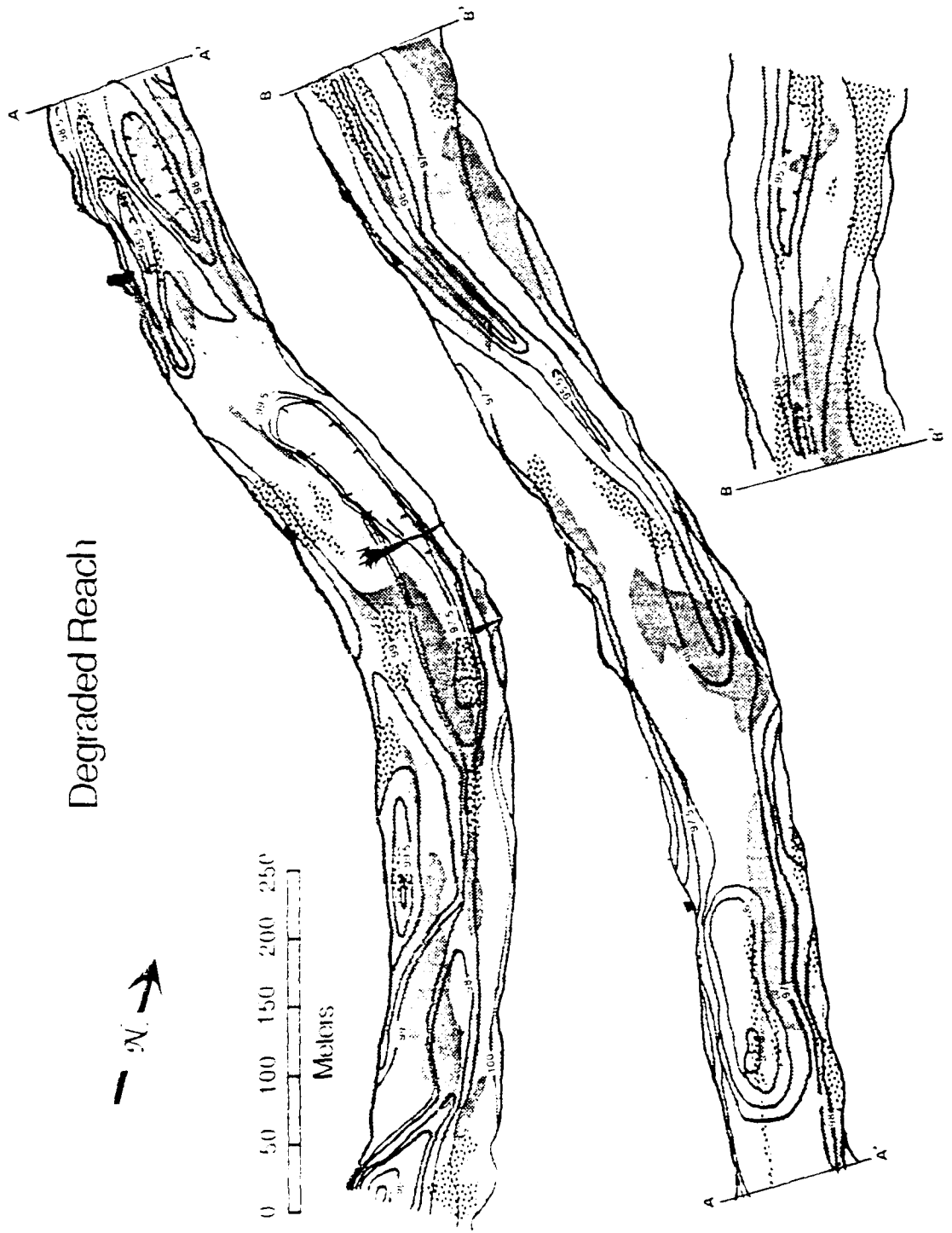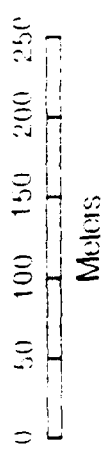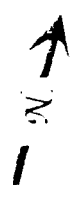
Aggrading reach   Degrading reach

All Facies   All Facies

Fines   Fines

Coarse Pebble   Coarse Pebble

Cobbles   Cobble

                         Bimodal

Percent finer than

Sieve size (mm)
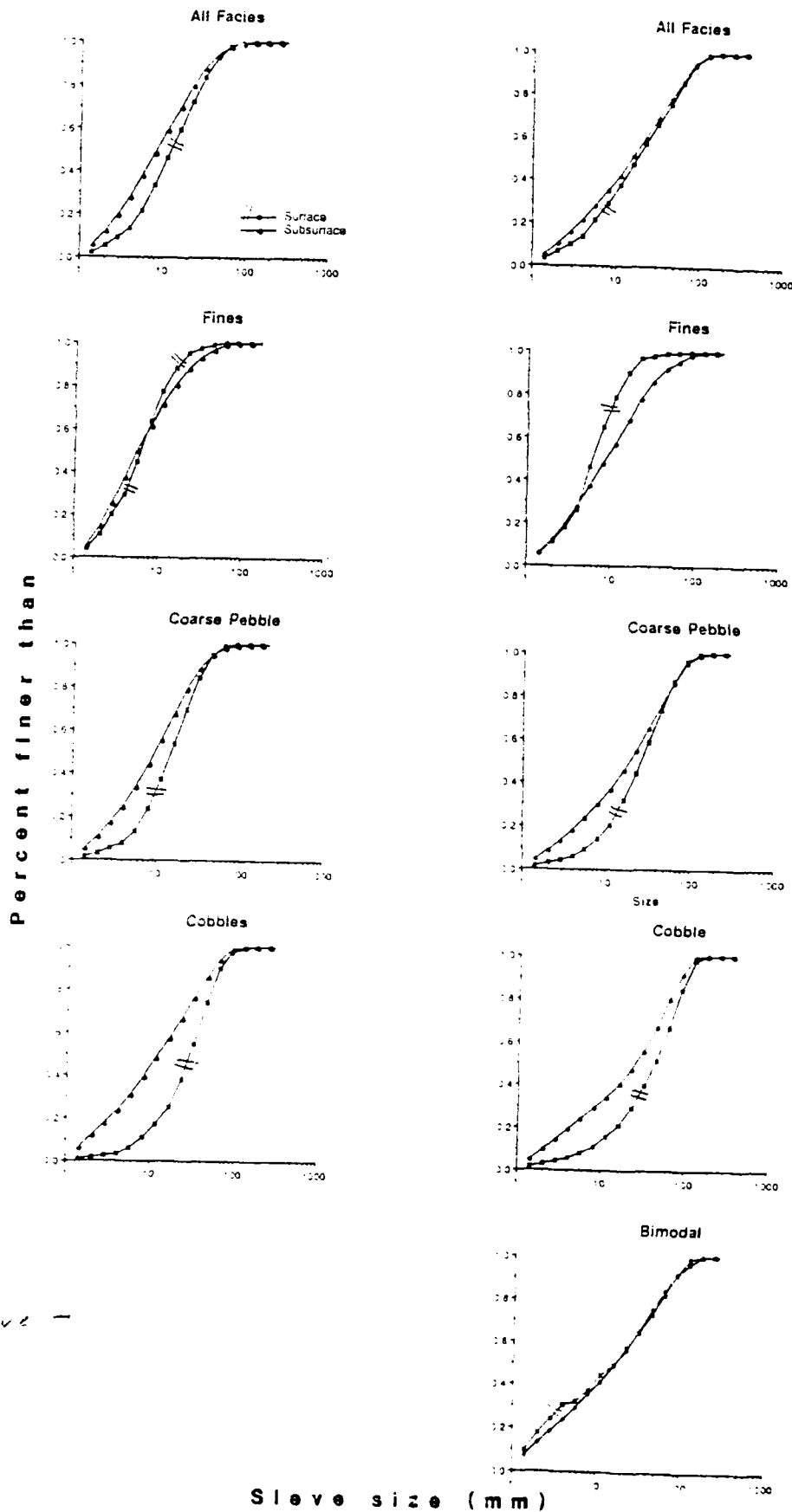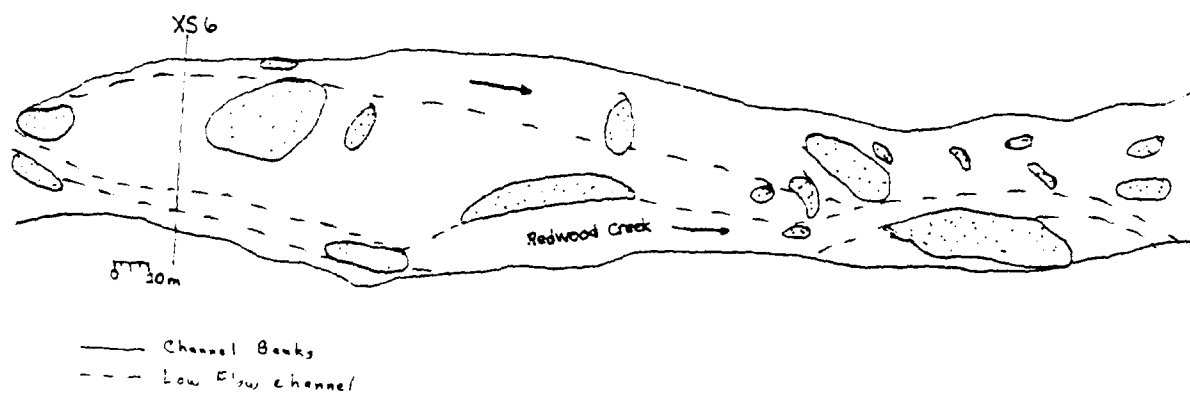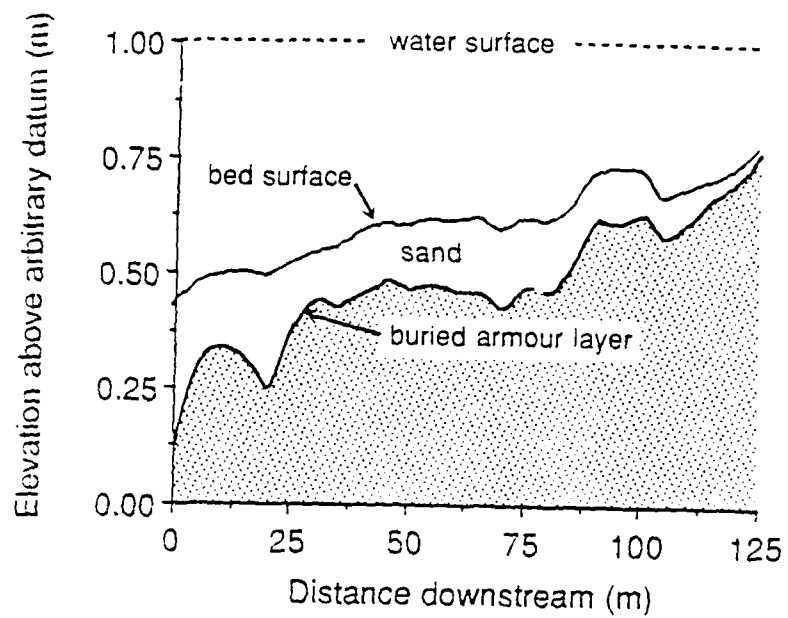
Figure 5

Figure 6

Figure 7
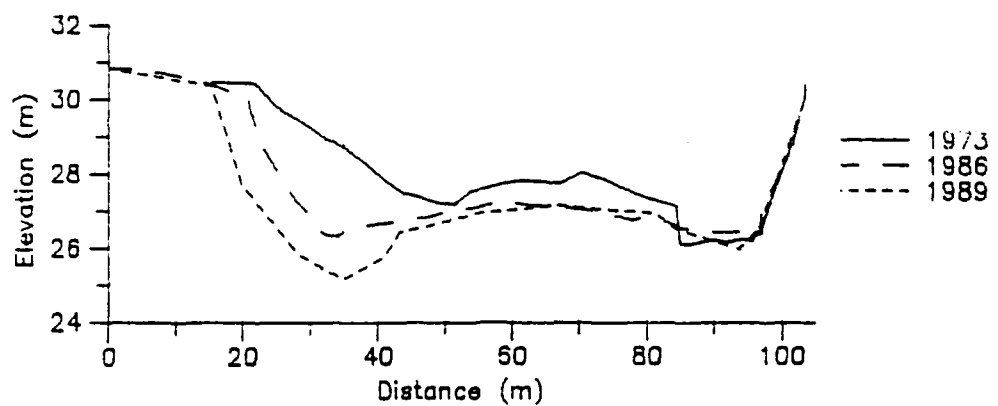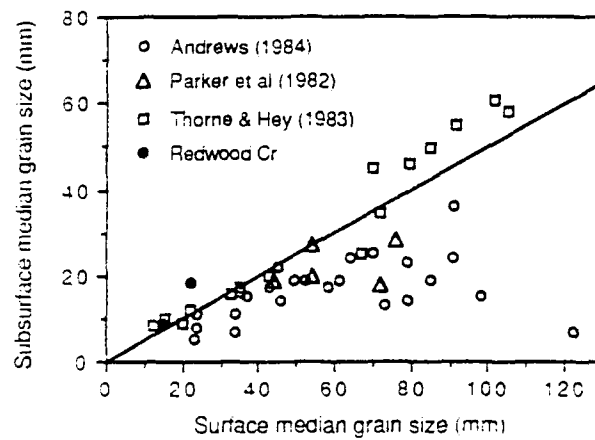
Figure 8

Equal Grain Mobility Versus Changing Bedload Grain Sizes in Gravel-Bed Streams

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ABSTRACT

Parker et al. (1982) made an assumption of equal-grain mobility as a first-order approximation in developing relationships for evaluating bedload transport rates of grain-size fractions in Oak Creek, a gravel-bed stream. That assumption yielded reasonable results in determinations of gravel transport rates, but did not account for observed variations in bedload grain-size distributions at different flow stages. As discharges and bed stresses increase in Oak Creek, the bedload grain sizes become significantly coarser and their distributions are increasingly skewed as they approach the distribution of the bed material (pavement plus subpavement). These changing bedload grain sizes demonstrate that there is a marked departure from equal grain entrainment and transport in Oak Creek. Higher-order solutions for predicting transport rates, those which do not assume equal mobility, have been developed by Diplas (1987) and Shih and Komar (1990). These advanced analyses provide predictions of changing bedload grain sizes as well as yielding improved calculations of transport rates.

Various lines of evidence have been offered in support of the equal-mobility hypothesis. A series of publications have analyzed the relationship between the largest particles found in bedload samples and flow discharges or bed stresses. Rather than demonstrating that the bedload grain-size distributions are nearly constant at all flow stages, which should prevail during equal mobility, those studies show that there are rapid changes in grain sizes as reflected by the largest particles in the bedload samples. It can be argued that these data represent a transitional stage during which the grain-size distributions of bedload samples are approaching the size distributions of the bed material, and that the faster this transition, the closer the conformity with the equal-mobility hypothesis. This interpretation constitutes a broader view of equal mobility in a stream than the specific conditions to which the bedload transport analyses of Parker et al. (1982) apply. There are problems with this broader interpretation in that comparisons between data from Oak Creek (well-developed pavement) and Great Eggleshope Beck (absence of a pavement) imply that the latter stream comes closer to equal mobility, in spite of the expectation that the pavement in Oak Creek should tend to equalize grain mobility. It is clear from this example that factors other than the presence of a pavement layer are important to sorting processes leading to variations in bedload grain sizes and the relative transport rates of different size fractions.

INTRODUCTION

Gravel-bed streams characteristically have wide ranges of grain sizes and display size-sorting patterns within their bed deposits. Particularly important is that the surface of the bed layer, its pavement, may be considerably coarser than the subpavement gravel [We are using these terms, as opposed to armor versus subarmor, to denote the frequent mobility of the gravels, corresponding to the use by Parker et al. (1982)]. This vertical sorting into pavement and

subpavement plays an important role in sediment transport in gravel-bed streams. Parker et al. (1982) have hypothesized that the existence of a bed pavement regulates the entrainment of particles by the stream, resulting in their being approximately equal in mobility, that is, all grain sizes are entrained at about the same flow discharge and are transported at rates in proportion to their presence in the bed material. This has come to be known as the equal-mobility hypothesis.

Subsequent studies have supported this hypothesis. Andrews (1983) and Andrews and Erman (1986), utilizing data from several streams, examined variations in the largest particle sizes found in bedload samples at different flow stages. They concluded that the maximum particle size undergoes almost no change over a range of flow discharges or bed stresses. Wilcock and Southard (1988) have undertaken flume experiments involving bed sediments of mixed sizes, and concluded that "all fractions in a size mixture begin moving at close to the same value of bed shear stress during steady-state transport conditions." A theoretical foundation for the hypothesis has been provided by the analysis of Wiberg and Smith (1987). The collective impact of these studies has been to "prove" the equal mobility or near-equal mobility of gravel entrainment and transport in streams.

This strong support for equal mobility appears to go well beyond the original intention of Parker et al. (1982) when they formulated the hypothesis. The objective of their work was to develop a method for calculating transport rates of different size fractions, and an assumption of equal mobility served as a first-order approximation. However, a corollary of perfect equal mobility is that with varying flow discharges, there should be no change in the grain-size distributions of the bedload samples, including no shifts in the maximum particle sizes. Parker et al. based their analysis on the data of Milhous (1973) from Oak Creek, Oregon, and noted that there actually are significant variations in bedload grain sizes with changing flow stage. Parker et al. recognized that such variations in bedload grain sizes represent a departure from their assumption of perfect equal mobility. With that recognition, they were the first to attempt the development of a higher-order analysis that would account for varying bedload grain-size distributions. Their approach was later modified by Diplas (1987).

Our research initially focused on flow-competence techniques, the evaluation of flood hydraulics from the sizes of particles transported (Komar, 1987a, 1987b, 1989; Komar and Li, 1988). Competence evaluations have traditionally depended on observations of the largest particle sizes moved, and assume that the greater the flood discharge the larger the particles in the resulting deposits. This use of variable maximum particle sizes has been interpreted by some as being in conflict with the equal-mobility hypothesis. It does disagree with the results of Andrews (1983) and Andrews and Erman (1986) who, based on analyses on maximum particle sizes in bedload samples, concluded that all sizes are entrained at effectively the same flow stress (a conclusion that would make flow-competence evaluations impossible). On the other hand, Ashworth and Ferguson (1989) found significant variations in bedload medians and maximum particle sizes in three streams, and concluded that there is substantial size-selective transport representing a departure from equal mobility. Our recent work has further established the nature of the shifting size distributions of bedload gravels, and has examined alternatives to maximum particle sizes for flow-competence evaluations (Shih and Komar, 1990a, 1990b; Komar and Carling, in review).

The objective of this paper is to undertake a reexamination of the equal-mobility hypothesis and the various types of data and analyses that have been offered to test the concept. This will begin with a review of the theory itself, as developed by Parker et al. (1982) and modified by Diplas (1987). Particular attention will be given to how the reference transport relates to grain threshold and to reformulations in terms of maximum particle sizes, and thus to variations in

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bedload grain-size distributions. The analyses presented here will be based on the Oak Creek measurements of Milhous (1973), a data set of particular relevance to questions of equal mobility in view of its use by Parker et al. (1982) to establish that concept. A general review will be presented of the studies reporting on the movement of maximum particle sizes and changing grain-size distributions in an attempt to discern whether there is a tendency toward equal mobility resulting from the presence of a bed-material pavement.

THE EQUAL-MOBILITY HYPOTHESIS

Basic derivation by Parker et al. (1982)

The study of Parker et al. (1982) was based primarily on a reanalysis of the bedload samples collected by Milhous (1973) in Oak Creek, a flume-like gravel-bed stream in western Oregon. Oak Creek has a well-developed pavement, consisting of a narrow range of large grain sizes with a median diameter of 6 cm; the grain-size distribution of the subpavement is broader, with a median of 2 cm. Bedload samples were collected in a vortex trap which insured a high sampling efficiency for all grain sizes within the bedload range. Milhous provides sieve analyses of the bedload samples, which permit evaluations of their individual transport rates.

The approach of Parker et al. (1982) was to analyze ten grain-size ranges, governed by the sieve intervals, and attempt to correlate the bedload transport rates of each range with the flow stresses. They used only those measurements of Milhous (1973) obtained during conditions of broken pavement, that is, when most of the grain sizes in the pavement are represented in the bedload. This limited the analysis to 22 measurement sets collected at discharges in excess of approximately 1 m^3/sec. The analysis was placed in dimensionless form. If q_{si} denotes the volumetric bedload transport rate per unit channel width, then the dimensionless Einstein bedload parameter for the ith grain-size range, represented by a mean D_i for that sieve fraction, is

$$q_{si}^* = \frac{q_{si}}{f_i[RgD_i]^{1/2}D_i} \tag{1}$$

where $R = (\rho_s - \rho)/\rho$ and ρ_s and ρ are respectively the gravel and water densities [Table 1 defines the symbols used here, together with those employed by Parker et al. (1982)]. In equation (1), f_i is the frequency of the sieve fraction as found in the size distribution of the subpavement; its inclusion has the effect of normalizing the transport rate of that size fraction to its availability in the stream bed [Parker and Klingeman (1982) provide a similar analysis in terms of the pavement grain-size distribution]. The flow stress τ is made dimensionless by its inclusion in the Shields entrainment function

$$\theta_i = \frac{\tau}{(\rho_s - \rho)gD_i} \tag{2}$$

expressed in terms of D_i for the size fractions. Correlations were developed between θ_i and

$$W_i^* = \frac{q_{si}^*}{(\theta_i)^{1.5}} \tag{3}$$

As noted by Parker et al. (1982), there are two advantages in using W_i^* rather than q_{si}^* as a

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dimensionless sediment transport rate. Plots of q_{si}^* versus θ_i are very steep, whereas the slopes of W_i^* versus θ_i are lower, and this makes the job of determining empirical relationships from the data somewhat easier. In addition, W_i^* is independent of D_i (the dependencies of q_{si}^* and θ_i on D_i cancel in the W_i^* ratio); in the correlation between W_i^* and θ_i, only the Shields function contains D_i.

The correlations of Parker et al. (1982) involved log-log plots and least-square regressions for each grain-size range to establish relationships of the form

$$W_i^* = \alpha_i \, (\theta_i)^{m_i} \tag{4}$$

Each grain-size range formed an individual correlation having different α_i and m_i coefficients. These were collapsed into a single relationship by the establishment of a reference bedload transport level. This was set at $W_r^* = 0.002$, and the corresponding θ_{ir} reference values were obtained from the regressions of equation (4) for the individual size fractions. These θ_{ir} values were plotted as a function of D_i/D_B, where D_B is the median diameter of the subpavement bed material. The correlation yielded

$$\theta_{ir} = 0.0876(D_i/D_B)^{-0.982} \tag{5}$$

This definition of a W_{ir}^* reference transport rate, and accompanying reference θ_{ir}, enabled Parker et al. to collapse the ten regressions of equation (4) for the individual size fractions into a single curve of W_i^* versus θ_i/θ_{ir}, given by

$$W_i^* = 0.002 \left(\frac{\theta_i}{\theta_{ir}} \right)^{m_i} \tag{6}$$

Parker et al. selected a weighted-mean value of 13.38 for the m_i exponent. In doing so, they noted that m_i actually increases with increasing D_i size fraction, and that this represents a departure from perfect equal mobility.

The above derivation is the first-order analysis of Parker et al. (1982), that which is based on an assumption of perfect equal mobility of different size fractions within the bedload. As seen in Figure 1A, the analysis provides a reasonable comparison between predicted and measured transport rates of gravel sieve-size fractions in Oak Creek. However, it should be recognized that in sediment transport predictions, an acceptable result is one where the predicted and measured values are within a factor of 5 or less. From that standpoint, this first-order solution would be acceptable in most applications. Therefore, the analysis has been a success in terms of predicting gravel-transport rates.

The reference transport and grain threshold

Arguments by subsequent investigators in favor of the equal mobility hypothesis have focused mainly on the reference condition of equation (5) for θ_{ir}. Most noteworthy is that the exponent is close to -1. From the definition of the Shields entrainment function, equation (2), the relationship of equation (5) is equivalent to

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$$\tau_{ir} \propto D_i^{0.018} \tag{7a}$$

which indicates that there is only a very small dependence of the reference stress on D_i. With an exponent of -1 in equation (5), this becomes

$$\tau_{ir} \propto D_i^0 = \text{constant} \tag{7b}$$

indicating that there is no dependence on D_i. If the reference transport, W_{ir}^*, and its derivative θ_{ir}, are thought of as a grain-threshold condition, then equation (7b) implies that all grain-size fractions are entrained at the same flow stress — ie., there is an equal mobility of the several size fractions with respect to their entrainment.

The origin of equation (5) is somewhat circuitous. It begins with a comparison of W_i^* and θ_i, equation (4), which actually involves an *a priori* assumption that the transport rate (W_i^*) is inversely proportional to D_i which is found in the denominator of θ_i. That assumption turns out to be incorrect, and this results in the horizontal displacements of the regressions of equation (4) parallel to the θ_i axis. The analysis leading to equation (5) has the effect of undoing this initial assumption that $W_i^* \propto 1/D_i$. This circularity accounts for the near-perfect correlation found in establishing equation (5), with $r^2 = 0.9997$ (Parker et al., 1982).

This does not detract from the usefulness of a reference transport rate W_r^* and equation (5) as employed by Parker et al. (1982) to collapse the regressions for the individual size fractions into a single relationship, equation (6). The reference transport can also be viewed as something of a threshold criterion for the several size fractions. However, equation (5) again obscures this threshold condition which should be thought of as some level of bedload transport which is the same for each size fraction. If we assume that the exponent in equation (5) is -1, then combining that relationship with the $W_r^* = 0.002$ reference transport level and the W_i^* definition of equation (3), leads to

$$q_{sir} = 0.0000519 \, f_i \, (Rg)^{1/2} D_B^{3/2} \tag{8}$$

for the actual reference transport rate, having eliminated its dimensionless form. It is seen that q_{sir} lacks any direct dependence on individual D_i sizes, but is proportional to f_i of the size fractions as found in the subpavement, and also depends on the D_B median of the subpavement. Both of these dependencies are reasonable for a threshold condition. The use of a small level of transport as a threshold criterion has been used in previous studies [see review by Miller et al. (1977)]. Those earlier studies involved the entire bedload transport, rather than individual size fractions. In formulating such a threshold criterion for a size fraction within the bedload, it is reasonable to have it proportional to the availability of that size in the stream bed (ie., to f_i), and to the overall grain size as represented by the D_B median of the bed material. This is in keeping with the quantitative definitions of grain threshold as developed by Neill and Yalin (1969) and extended by Wilcock (1988) in the context of applications to grain-size fractions.

It is of interest to explore just how much bedload transport is involved in the reference level established in the analysis of Parker et al. (1982). This can be calculated directly with equation (8) — the resulting q_{sir} for the 10 size fractions are given in Figure 2A. As expected from the formulation of equation (8) and its dependence on f_i of the subpavement, the resulting histogram of the transport rates of size fractions mirrors the grain-size distribution of the subpavement. The

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histogram of Figure 2B gives the results expressed in terms of the numbers of particles transported in the size fractions for the reference transport rate. As expected, the reference transport involves thousands of grains in the small to medium-size fractions. There is a rapid decrease in numbers of grains in the coarser fractions. Even with several hours of transport at this low rate, no particles would be captured in the coarsest sieve fractions. The largest particles of the bedload samples, those employed in flow-competence evaluations (Komar, 1987a), are mainly found in still coarser sieve sizes than diagramed in Figure 2 and included in the analysis of Parker et al. (1982). Additional analyses show that the bedload transport contributed by those largest particles in almost all cases falls below the $W_r^* = 0.002$ reference level. It is apparent that the reference transport rate defined by Parker et al., which can be viewed as a type of threshold criterion, differs from the initial movement of the largest particles found in the bedload samples. There is no straight-forward way to relate the two approaches, which are conceptually different.

Higher-order solutions

The first-order solution of Parker et al. (1982), that which assumes equal mobility, does not account for changing grain sizes of the bedload. It always yields the same distribution as the subpavement, just as seen in Figure 2A for the reference transport. As flow discharges or bed stresses increase, the quantities of transport of the size fractions are greater, but remain in the same proportions so that the resulting grain-size distributions are invariant. As noted in the Introduction, Parker et al. were aware that size-distributions of bedload samples from Oak Creek do undergo marked changes, and this induced them to develop a higher-order solution. Their initial attempt has been extended by Diplas (1987), who concentrated on variations in the m_i exponent of equation (4). Diplas noted that rather than being constant as assumed by Parker et al., m_i ranges from 5.49 to 21.41 with increasing grain size, a 290% change. Diplas correlated m_i and D_i to obtain

$$m_i = 13.71(D_i/D_B)^{0.3214} \qquad (9)$$

Bedload grain-size distributions calculated with this modified approach for a series of flow discharges are shown in Figure 3A. These were obtained by evaluating transport rates of the 10 size fractions, calculating the total bedload for a 1-hour interval, and then converting it to a grain-size distribution. The original 10 size fractions represent irregular sieve intervals, so some have been combined to yield the 8 size fractions given in Figure 3A, representing even 1-ϕ intervals. It is seen that this higher-order solution of Diplas (1987) yields a coarsening of the bedload grain sizes as the flow discharge increases, and even displays the development of asymmetry as found in actual bedload samples from Oak Creek (Shih and Komar, 1990a). This represents an obvious improvement over the equal-mobility approximation of Parker et al. (1982). As seen in Figure 1B, there is also a considerable improvement in the calculated transport rates of the size fractions when the Diplas (1987) solution is compared with the equal-mobility results (Figure 1A).

It is of interest that the modification by Diplas (1987) involved changes in the m_i exponents. It is apparent from the basic dependence of equation (4), that an increase in m_i for the coarser size fractions will enhance their transport rates relative to the finer fractions; this accounts for the coarsening of the bedload samples at the higher flow stages. In modifying the bedload computations, Diplas obtained nearly the same coefficients as given in equation (5) for the reference transport

condition, those obtained by Parker et al. (1982). The exponent was changed by Diplas to -0.943, and in spite of its still being close to -1, the results now yield realistic variations in grain-size distributions and, therefore, significant departures from equal grain mobility. It is apparent that variations in the m_i exponents are more important to arguments of equal mobility, or are at least represent a more sensitive test, than examinations of the reference transport and grain threshold conditions.

ANALYSIS BASED ON GRAIN-SIZE DISTRIBUTIONS

A distinctly different approach has been taken by Shih and Komar (1990a, 1990b) to evaluate transport rates of the grain-size fractions in Oak Creek. It was first established that there is a systematic evolution of the bedload grain-size distributions, demonstrating that the distributions progressively become more skewed with increasing flow stage. Goodness-of-fit comparisons established agreement with the theoretical Rosin distribution at discharges in excess of 1 m^3/sec, and empirical relationships were established between the peak mode grain size (k) and spreading coefficient (s) of the Rosin distribution with the flow stress and discharge (Figure 4A). The total gravel transport rate, also a function of the flow stress or discharge, was then apportioned within the frequency distribution of grain sizes to determine the distribution of fractional transport rates (Figures 4B and 4C). Among the advantages of this approach are that it yields smooth frequency curves of both the bedload grain-size distributions and transport rates, rather than being confined to histograms based on the sieve-size intervals that happened to be used in the original data analyses. If desired, the frequency curves can be "degraded" back into sieve-size evaluations of bedload grain sizes and transport rates. A series of calculated bedload grain-size distributions is shown in Figure 3B, both as the original smooth frequency curves and as the derivative histograms. These provide a closer simulation of the changing bedload distributions in Oak Creek than does the Diplas series in Figure 3A. The analysis procedure yields the predicted versus measured transport comparison shown in Figure 1C – it is apparent from the respective diagrams in Figure 1 that the approach of Shih and Komar provides better predictions than either the equal-mobility approximation of Parker et al. (1982) or the modified analysis of Diplas (1987).

Note: We are presently using the procedures developed by Shih and Komar to simulate evolving bedload grain-size distributions and fractional transport rates. This simply involves a change in the flow-stress dependencies of k and s of the Rosin distributions, and the dependence of the total bedload transport rate on the flow stage. This generates "data" for fractional transport rates of the series of sieve fractions, having manipulated the degree of overall changes in the bedload grain-size distributions and how they might approach the distribution of the bed material. The simulated "data" are being analyzed by the procedures of Parker et al. and Diplas to further establish how the reference equation (5) and the m_i exponents of equations (4) and (6) depend on varying degrees of evolution of bedload grain-size distributions. We had hoped to be further along so as to provide some discussion of the results here. We would like to add it later.

EQUAL MOBILITY AND FLOW COMPETENCE
Correlations involving maximum-particle diameters

Subsequent investigations to test the relative mobility of gravels in streams have focused primarily on equation (5), rather than on examinations of the actual transport rates of the various size fractions or on the m_i exponents. As noted above, this is unfortunate in that

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variations in m_i and the accompanying changes in bedload grain-size distributions provide a more sensitive test. The analyses have involved the largest particle sizes (D_m) found in bedload samples, so that equation (5) was modified to

$$\theta_m = a(D_m/D_B)^b \tag{10}$$

where D_B is the median diameter of the bed material. This seemingly straight-forward change has led to confusion with respect to what actually constitutes equal grain mobility. Equation (10) is equivalent to the basic proportionality

$$\tau \propto D_m^{b+1} \tag{11}$$

since the other parameters (ρ_s, ρ and D_B) are constants for a given stream reach. It has been assumed by most investigators that a $b = -1$ exponent provides evidence for equal mobility, just as it did in equation (5). Equation (11) then reduces to

$$\tau \propto D_m^0 = \text{constant} \tag{12}$$

Equations (11) and (12) are basically flow-competence relationships between the flow stresses and the maximum particle diameters entrained from a deposit of mixed sizes. For nearly all data sets of gravel movement in streams, the exponent of equation (11) is greater than 0 (i.e., $b > -1$), so that τ increases with increasing D_m, the typical flow-competence correlation (Komar, 1987a, 1987b). The zero exponent of equation (12) implies the absence of such a relationship, making flow-competence calculations impossible. However, a more precise meaning of the condition inherent in equation (12) is unclear. It could be obtained, for example, by the collection of repeated samples at a fixed flow stage, so that τ is essentially constant while there is a range of D_m values from the series of samples. With respect to flow-competence considerations, equation (12) denotes an extremely rapid change in D_m with little or no change in τ. However, this is contrary to the basic premise inherent in the equal-mobility assumption for computations of gravel transport rates, that is, the grain-size distributions of the bedload samples (presumably including D_m) do not change with the flow stress.

   Rather than viewing the variations of maximum particle sizes as an aspect of grain threshold, it is more enlightening to consider the D_m values as part of the bedload grain-size distributions, and to examine how the total spectrum of sizes varies with the flow stage. This reversal makes the flow stress the independent variable, and D_m the dependent variable, a more natural consideration of the actual processes. Equation (10) then reduces to

$$D_m \propto \tau^{1/(b+1)} \tag{13}$$

However, now we can generalize to consider any grain-size parameter of the bedload distributions, so that the relationships become

$$\theta_x = a(D_x/D_B)^b \tag{14}$$

and

$$D_x \propto \tau^{1/(b+1)} \tag{15}$$

where D_x could variously be the medians (D_{50}) of the bedload distributions, other percentiles (D_{60}, D_{70}, etc.), as well as D_m. The condition of equal mobility in terms of there being no changes in bedload grain-size distributions now becomes

$$D_x \propto \tau^0 = \text{constant} \tag{16}$$

that is, all percentiles including D_m remain invariant, independent of τ. From equations (15) and (16) we have $1/(b+1) = 0$ and therefore $b = \infty$. Hence, viewing D_m as one of several parameters representing the bedload grain sizes yields a much different interpretation of equation (10). Furthermore, it is apparent in equation (15) that as b approaches -1, the value generally assumed to represent equal mobility, the $1/(b+1)$ exponent rapidly increases, signifying that there are progressively greater changes in D_x for a given change in τ. When $b = -1$, the relationship becomes $D_x \propto \tau^\infty$; in effect, an instantaneous jump in D_x for an infinitesimal increase in flow stress τ.

The measured D_m maximum diameters, D_{50} bedload median diameters and D_{90} percentiles for Oak Creek are plotted in Figure 5 as a function of the flow stress, the dependence indicated by equation (15). This is the data collected by Milhous (1973) at discharges in excess of the critical 1 m^3/sec when the pavement layer is mobilized, the same subset utilized by Parker et al. (1982) to establish the equal-mobility model. The line for D_{50} versus τ is significantly steeper than that for D_m, establishing that for a give change in τ, there is a larger variation in D_{50} than for D_m. The respective proportionalities are:

$$D_{50} \propto \tau^{8.9} \tag{17a}$$

$$D_{90} \propto \tau^{3.8} \tag{17b}$$

$$D_m \propto \tau^{2.8} \tag{17c}$$

Comparable relationships have been obtained for the other grain-size percentiles of the bedload distributions (D_{60}, D_{70}, ... D_{95}). The exponents are intermediate between those for D_{50} and D_m, forming a continuous trend of decreasing exponents as $D_x = D_{50}$, D_{60} ... D_m shifts to coarser sizes (Komar and Carling, in review). This pattern reflects the development of increasing skewness in the grain-size distributions at higher flow stresses, as shown by Shih and Komar (1990a), with the D_{50} medians shifting most rapidly and D_m showing the least response to varying flow stresses. With respect to the equal mobility assumption that there are no variations in bedload grain sizes, D_m comes closest to that condition while the D_{50} medians represent the greatest departure. The proportionalities of equation (17) are equivalent to

$$\theta_{50} = \tau/(\rho_s - \rho)g\, D_{50} \propto (D_{50}/D_B)^{-0.89} \tag{18a}$$

$$\theta_{90} = \tau/(\rho_s - \rho)g\, D_{90} \propto (D_{90}/D_B)^{-0.74} \tag{18b}$$

$$\theta_m = \tau/(\rho_s - \rho)g\, D_m \propto (D_m/D_B)^{-0.64} \tag{18c}$$

The exponent of the θ_{50} relationship comes closest to the $b = -1$ value, in spite of the fact that D_{50} represents the greatest deviation from a condition of equal mobility as an approximation for computations of gravel-transport rates.

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Rather than viewing the above results as the actual conditions required for application of the equal-mobility analysis to compute gravel transport rates, as developed Parker et al. (1982), one can consider them as the transitional phase leading to that condition of equal mobility. This is evident for the Oak Creek data in Figure 5. By this interpretation, the shifts in $D_{50} \ldots D_m$ with increasing flow stresses represent an evolution in bedload grain-size distributions as they progressively approach the distributions of the bed material. The medians of the pavement and subpavement distributions are marked by arrows in Figure 5, and the line for the D_{50} medians of the bedload samples has been extrapolated (dashed portion) to illustrate how the transitional stage of increasing D_{50} might reach the D_{50} = constant condition necessary for application of the equal-mobility computations of gravel-transport rates. In our comparisons of the complete grain-size distributions (Shih and Komar, 1990b), we concluded that the bedload samples collected at the highest flow stages were approaching an approximate mixture of 70% subpavement and 30% pavement. That mixture yields the D_{50} = 3.5 cm bedload median for the level "equal mobility" portion of the dashed curve in Figure 5. All of the percentiles, and D_m, should follow a similar evolution and at some high flow stage reach nearly-constant values determined by the bed material grain-size distribution.

This interpretation that the Milhous (1973) data from Oak Creek represent a transitional phase of rapidly evolving bedload grain-size distributions is certainly valid. It is unfortunate that Milhous did not obtain additional measurements at still higher flow stages in that Figure 5 implies that the change over to the D_x = constant equal-mobility condition must take place at only marginally greater flows (discharges on the order of 5 m^3/sec). Recognition that the Milhous measurements represent a transitional phase explains the inadequacies of the equal-mobility approximation for calculating bedload transport rates. It is apparent that the assumptions made by that analysis, specifically that the bedload grain-size distributions are the same as the bed material and do not vary with flow stage, are not achieved in Oak Creek until discharges exceed about 5 m^3/sec.

A similar interpretation has been made by Ashworth and Ferguson (1989) for their bedload samples collected in three streams. They established that the medians and maximum particle sizes increase with flow stage, and concluded that the data represent selective size-sorting. Their measurements suggest the proportionalities $D_{50} \propto \tau^{1.1}$ and $D_m \propto \tau^{3.8}$ (obtained respectively from an approximate fit to their plotted data, and algebraic modification of their θ_m versus D_m/D_B correlation). Their results differ from those obtained here for Oak Creek in that D_m changes more rapidly than D_{50}. The implication is that the bedload samples collected by Ashworth and Ferguson shift toward coarser sizes with increasing flow stage, but the distributions do not become progressively skewed as do those in Oak Creek. However, their trends for D_{50} variations are based mainly on data collected in the Lyngsdalselva, while variations in D_m are supported mainly by data from the Dubhaig and Feshie. Therefore, conclusions about evolving bedload grain-size distributions (skewness, etc.) are not possible without detailed analyses of the original samples. Their results do clearly establish, however, that the data represent a transitional stage of changing grain sizes, and that if true equal mobility is achieved, it will be at a higher flow stage than represented by their measurements.

Comparisons of field data for D_m variations

It is probable that most, if not all, of the studies that have examined D_m variations in streams are for data in the transitional phase with rapidly evolving grain-size distributions. In that case,

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the closer the b exponent of equation (10) is to -1, the more rapid the transition of bedload grain sizes toward that of the bed material. Accordingly, this can be incorporated into a boarder view of what constitutes equal mobility in a stream, one that includes the transitional phase of evolving bedload distributions as well as the final achievement of the true equal-mobility condition when D_x = constant. Then, the more rapid the transition (i.e., the closer b is to -1), the sooner the equal-mobility approximation of Parker et al. (1982) could be used in transport calculations.

The results of several studies of D_m variations are compiled in Table 2, ordered with respect to magnitudes of the b exponents in equation (10). Oak Creek forms one extreme with b = -0.43 if the total winter-1971 data of Milhous (1973) are used and b = -0.64 if only the subset for Q > 1 m^3/sec are employed (Komar, 1987a; Komar and Carling, in review). With the assumption that b = -1 constitutes equal mobility, Oak Creek gravel entrainment and transport would represent the greatest departure, in spite of its well-developed pavement which should act to equalize mobility. In contrast, Great Eggleshope Beck yields b = -0.82 , even though it has a minimal development of bed pavement (Carling, 1983; Komar and Carling, in review). This argues against the broad interpretation that for measurements in the transitional phase of changing bedload grain sizes, the stream that comes closest to b = -1 due to regulated entrainment from a pavement layer represents equal mobility.

The listing in Table 2 shows that the studies of Andrews (1983) and Andrews and Erman (1986) obtained exponents closest to b = -1 in equation (10). Andrews (1983) based his analysis on data from the East Fork River, Wyoming, obtained with a total bedload trap (Leopold and Emmett 1976, 1977), and from the Snake and Clearwater Rivers of Idaho, collected with a Helley-Smith portable sampler (Jones and Seitz, 1980). Andrews and Erman (1986) sampled gravel transport in Sagehen Creek, California, also employing a Helley-Smith sampler. The -0.87 exponent obtained by Andrews (1983) is equivalent to the proportionality $D_m \propto \tau^{7.7}$ so that a 3-fold increase in the flow stress, modest for most streams, results in a 4,700-factor change in D_m.

In analyzing the East Fork data, Andrews (1983) had to rely on the geometric mean of the coarsest sieve fractions of transported bedload, since the data collections did not include direct measurements of the maximum-size clasts. In order to test whether this procedure can still reflect the changing flow competence, we analysed the Oak Creek data in the same way, that is, by using the coarsest sieve fractions reported by Milhous (1973). The analysis yielded essentially the same empirical exponents as obtained utilizing the largest grain sizes transported. Therefore, it can be concluded that the reliance of Andrews on sieve fractions probably yielded reasonable results and does not account for his obtaining a b closer to -1 than found in other studies.

The investigations that have obtained exponents closest to -1 in equation (10) relied almost exclusively on portable bedload samplers, whereas the Oak Creek data (Milhous, 1973) and Great Eggleshope measurements (Carling, 1983) were both obtained with total bedload traps. This suggests that the spectrum of b exponents seen in Table 2 may have resulted in part from differences in sampling techniques. The potential sampling problems in using the Helley-Smith to determine the movement of the coarsest fractions is illustrated by the study of Andrews and Erman (1986) in Sagehen Creek, California. Their results are graphed in Figure 6 in terms of θ_m versus D_m/D_B so as to provide a comparison with equation (10). It is seen that the data do yield an exponent b = -1. Andrews and Erman report that runoff during the two weeks of their sampling was the longest period of sustained high flows that had occurred in the previous 30 years. The largest particles captured in the Helley-Smith ranged up to a maximum 8.6 cm. However, their experiments at other times with tagged clasts showed that material of 10-cm diameter readily moves, even during flow stages that are significantly lower than when the Helley-Smith

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measurements were made. It appears that in this particular study the Helley-Smith was unsuccessful in sampling the larger clast sizes required to evaluate flow competence and to provide an adequate test of equation (10). However, any systematic under-representation of D_m obtained by Andrews and Erman in their gravel sampling probably does not account for the plotting pattern of their data in Figure 6 and the resulting b = -1 exponent. It would be reasonable to expect that as the flow stress increases in the stream, there would be greater problems in sampling the increasingly larger D_m. This would have the effect of increasing the slope of a τ versus D_m regression, and a counter-clockwise rotation of the line fit to θ_m versus D_m/D_B, that is, away from the b = -1 condition. Instead, the b = -1 result in Figure 6 for Sagehen Creek probably results from the measurements of D_m being random rather than representing systematic sampling errors. Further evidence for this is that the data collected by Andrews and Erman were obtained under a small range of flow stresses, the median diameters of the bedload samples showed little change (1.1 to 1.9 cm), while the maximum diameters varied more widely (5.2 to 8.6 cm). Using a random-number generator on a computer, constrained by similar ranges of stresses and D_m values, we obtained "data" that look very similar and yield b = -1 in equation (10). If the flow stress were absolutely constant, then a plot of stream-derived data as θ_m versus D_m/D_B simply constitues a graph of $1/D_m$ versus D_m, and the scattered D_m values yield a b = -1 exponent. Thus, the closer the data is to random, the better the fit to equation (10) with b = -1 and $r^2 = 1$.

Caution is, therefore, required in interpreting θ_m versus D_m/D_B diagrams, particularly if the exponent is close to b = -1. This is more generally illustrated in Figure 7, employing the complete set of winter-1971 data from Oak Creek (Milhous, 1973). In Figure 7A the data are plotted as τ versus D_m, a flow-competence comparison, while in Figure 7B the same measurements are presented as θ_m versus D_m/D_B. The solid curve in Figure 7A is based on a fit of τ versus D_m, and yields

$$\tau = 108 \, D_m^{0.569} \tag{19a}$$

Algebraic manimulation of this, using the constant values $\rho_s = 2.85$ and $\rho = 1$ g/cm^3 and $D_B = 2$ cm for the subpavement, yields

$$\theta_m = \tau/(\rho_s - \rho)g \, D_m = 0.044(D_m/D_B)^{-0.43} \tag{19b}$$

which has been plotted in Figure 7B. The dashed curves have been derived in reverse order, based first on a comparison of θ_m versus D_m/D_B, and then simplified to τ versus D_m. The corresponding relationships are

$$\tau = 125 \, D_m^{0.35} \tag{20a}$$

and

$$\theta_m = 0.045(D_m/D_B)^{-0.65} \tag{20b}$$

Equation (20b) was obtained by Komar (1987a) for the combined data of Milhous (1973), Carling (1983) and Hammond et al. (1984). It shows best agreement with the latter data sets, but from the θ_m versus D_m/D_B comparsion in Figure 7B, it would appear to represent the Oak Creek data as well. However, when the equivalent equation (20a) is plotted as τ versus D_m in Figure 7A, it becomes apparent that the agreement is poor. Three of the data points that appear to substantiate the relationship in Figure 7B are now seen to be outliers. The data collected at low flow stages, particle sizes less than 5 cm, are seen in the τ versus D_m comparison to have little or no trend of their own. However, these data are important in establishing equation (20b) in the θ_m versus D_m/D_B comparsion since they are strung out over a D_m/D_B range 0.4 to 2.3. That subset of

data from low-flow stages is essentially random, and taken alone yields an exponent $b = -1$. Therefore, the $b = -0.65$ exponent in equation (20b) is a combination of the -1 exponent from the random subset of data and the -0.43 exponent which is based mainly on data collected at discharges greater than 1 m^3/sec.

The influence of a subset of random data is still more apparent in analyses of the D_{50} median diameters from the same Oak Creek bedload samples. The data are presented in Figure 8, again as τ versus D_{50} and θ_{50} versus D_{50}/D_B. There is a very tight correlation at stress levels above the critical stage for breakup of the armor, Figure 8A, which yields the relationship

$$\tau = 339 \, D_{50}^{0.152} \tag{21a}$$

Algebraic conversion to yield $\theta_{50} = \tau/(\rho_s - \rho)gD_{50}$ versus D_{50}/D_B gives

$$\theta_{50} = 0.104(D_{50}/D_B)^{-0.848} \tag{21b}$$

which is seen in Figure 8B to agree with the principal data trend. The dashed line in Figure 8B is fit to the subset of data obtained at low flow stages which are again basically random. It is seen in Figure 8A that for this subset, there are large fluctuations in the measured medians of the bedload samples while the flow stresses have only a very small range of values. When plotted as θ_{50} versus D_{50}/D_B in Figure 8B, this subset taken alone yields a $b = -1$ exponent.

Some portion of the differences in degrees of sorting found by the various studies, represented by the b exponents tabulated in Table 2, is undoubtedly a product of contrasting environmental conditions. In the case of the Oak Creek versus Great Eggleshope Beck data, we have concluded that the differences in the b exponents result mainly from contrasting degrees of skewness of the bed materials in these two streams and how this regulates the evolution of the bedload grain-size distributions (Komar and Carling, in review). It was also shown in that study that the exponents in the competence relationships can be significantly affected by sampling intervals, whether the samples are collected over a relatively short period of constant stage (Oak Creek) or represent integrated samples over a complete flood cycle (Great Eggleshope). It is certain that there will also be systematic differences resulting from the use of total bedload traps, as employed by Milhous (1973) and Carling (1983), versus the use of portable samplers. It might be expected that this would affect D_m comparisons more than the D_{50} medians of the bedload distributions. Finally, it has been seen here that the resulting empirical coefficients can be greatly affected by the data analysis, specifically if the comparison is θ_x versus D_x/D_B or τ versus D_x. It is probable that the range of empirical coefficients compiled in Table 2 result mainly from these systematic differences in experimental and analysis techniques, which unfortunately obscure potential environmental factors.

SUMMARY AND DISCUSSION

An assumption of equal grain mobility by Parker et al. (1982) proved to be a reasonably successful first-order approximation for calculations of bedload transport rates of grain-size fractions in gravel-bed streams. The essence of the equal-mobility analysis is that at some low-flow stage all grain sizes are entrained in their relative proportions found in the bed material, and as the flow increases the transport rates of the size fractions increase but remain in those initial proportions. Therefore, according to the equal-mobility model, the bedload

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grain-size distributions are always the same as found in the bed material and do not change with flow stage.

This assumption of constant bedload grain-size distributions constitutes the most obvious failure of the equal-mobility hypothesis. In Oak Creek the bedload distributions become progressively coarser with increasing flow, and become systematically more skewed as they approach the distribution of the bed material. Estimates here indicate that agreement between bedload and bed-material distributions is not achieved in Oak Creek until discharges exceed about 5 m^3/sec. The data of Milhous, therefore, represent a transitional phase of evolving bedload transport and size distributions rather than a condition of true equal mobility. This requires higher-order transport calculations than provided by the equal-mobility assumption, approaches that have been developed by Diplas (1987) and Shih and Komar (1990b). Diplas modified the basic analysis of Parker et al. (1982), while Shih and Komar developed an approach that more closely follows the evolving bedload grain-size distributions. Both analyses predict changing bedload grain sizes (Fig. 3) as well as yielding improved estimates of transport rates (Fig. 1).

Most subsequent studies designed to test the equal mobility of gravel transport in streams have focused on the reference transport condition, and the derivative equation (5) which has been viewed as a grain-threshold criterion. This is unfortunate in that the analyses by Diplas (1987) demonstrated that variations in the m_i exponents of equation (4) are more critical to changing grain-size distributions and the departure from equal mobility. The focus, instead, has been mainly on variations in the D_m maximum particle sizes in bedload samples, fitted to the empirical equation (10) as θ_m versus D_m/D_B. The results have been interpreted in terms of grain threshold, an extension of equation (5) for the series of size fractions. It has been argued here that equation (10) is better viewed as part of the changing bedload grain-size distributions, D_m being one of several grain-size parameters (D_{50}, D_{60}. . ., D_{95}, D_m) that would reflect variations in distributions with flow stage. By this interpretation, the closer the exponent is to b = -1 the more rapid the change in bedload grain sizes as they approach the distribution of the bed material. This could constitute a broader view of equal mobility in a stream where the presence of a pavement layer enhances the rate of this evolution (making b closer to -1). However, this interpretation is not borne out by comparisons of data from various streams. Oak Creek, with a classic development of pavement and subpavement sorting, yields a b exponent that departs most from -1 of the streams studied. It was concluded that this b exponent can reflect the development of skewness in the bedload samples as they approach the distribution of the bed material, as well as any influence of a bed pavement in regulating entrainment. In addition to these environmental dependencies, b can also be affected by sampling techniques and data analysis procedures. Unfortunately, the critical b = -1 condition is most apt to represent random samples, data where the flow stress has a small range of values (typical for most streams) while fluctuations in D_m are large. This was illustrated with the Oak Creek data (Figs. 7 and 8), where it was seen that even when a small subset of the data is random or where there are outliers, a plot of θ_m versus D_m/D_B tends to yield a steeper slope, one that is closer to b = -1. It was apparent in the analysis that it is preferable to first directly compare τ with D_m, and then to algebraically derive θ_m versus D_m/D_B.

The existence or degree of departure from equal mobility in a stream is best reflected in grain-size distributions of bedload samples and how they compare with bed-material distributions. There should be greater emphasis on documenting the complete distributions of bedload grain sizes and how they respond to varying flow stresses and discharges, rather than relying solely on the largest transported particles which tend to be affected by sampling techniques. Detailed analyses of bedload grain-size distributions are required if we are to

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understand grain-sorting patterns and how they are controlled by environmental factors such as the presence of a bed pavement.

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FIGURE CAPTIONS

Figure 1: Measured versus predicted transport rates of sieve-size fractions in Oak Creek: A. The equal-mobility analysis of Parker et al. (1982). B. The modified higher-order solution of Diplas (1987). C. The grain-size distribution approach of Shih and Komar (1990b). The individual sieve-size fractions are not identified here, but can be seen in the plots as presented in Shih and Komar (1990b).

Figure 2: A. The distribution of the q_{sir} bedload transport rates for the individual size fractions, based on equation (3) which is equivalent to combining the $W_{ir}^* = 0.002$ reference transport of Parker et al. (1982) with the derivative relationship, equation (5). B. The reference transport rate in terms of the numbers of grains found in each size fraction.

Figure 3: A. Grain-size distributions derived from calculated bedload transport rates obtained with the Diplas (1987) analysis approach. B. Frequency curves and derived histograms for bedload grain-size distributions calculated according to the approach of Shih and Komar (1990).

Figure 4: A schematic summary of the procedure of Shih and Komar (1990a, 1990b) for calculating bedload grain-size distributions and fractional transport rates. A. The Rosin distribution, with its mode size k and spreading coefficient dependent of τ for the bedload size distributions in Oak Creek. B. The total bedload transport rate as a function of τ. C. Partitioning the total transport into the distribution of sizes, to obtain frequency curves for the fractional transport rate as a function of τ.

Figure 5: Changes in the maximum particle sizes (D_m), the D_{90} coarse percentile and the medians (D_{50}) of the Oak Creek bedload samples with varying discharges and bed stresses.

Figure 6: The data of Andrews and Erman (1987) from Sagehen Creek, yielding b = -1 in eq.(10).

Figure 7: A. The Oak Creek flow-competence relationship for the flow stresses τ versus the maximum particle sizes, D_m. B. Plotting of the same data in terms of θ_m versus D_m/D_B where D_B is the median diameter of the subpavement.

Figure 8: A. The Oak Creek flow-competence relationship for the stresses τ versus the D_{50} median diameters of the bedload samples. B. Plotting of the same data in terms of θ_{50} versus D_{50}/D_B.

Table 1 — List of the symbols used here, compared with those employed by Parker et al. (1982).

Meaning of Symbol	This Paper	Parker et al.
Grain diameter representative of sieve-size fraction:	D_i	D_i
Median diameter of subpavement bed material:	D_B	D_{50}
Median diameter of bedload sample:	D_{50}	--
Percentile sizes of bedload samples:	D_{60}, etc.	
Largest particle diameter in bedload sample:	D_m	--
Frequency of D_i sieve fraction in subpavement:	f_i	f_i
Exponents in transport relationships [eqs. 4 and 6]	m_i	m_i
Volumetric bedload transport rate of size fraction:	q_{si}	q_{Bi}
Dimensionless volumetric transport rate:	q_{si}^*	q_{Bi}^*
Normalized, dimensionless transport rate:	W_i^*	W_i^*
Grain density:	ρ_s	ρ_s
Water density:	ρ	ρ
Stream flow mean bed stress:	τ	τ
Shields entrainment function:	θ	τ^*

Table 2 — Exponents in the $\theta_m = a(D_m/D_B)^b$ empirical relationship obtained by various investigations.

	b	D_m/D_B
Oak Creek		
total data set	-0.43	0.4 - 6
Q > 1 m³/sec	-0.64	2 - 6
English Channel (Hammond et al.)	-0.58	0.4 - 5
Lyngsdalselva, Dubhaig and Feshie (Ashworth & Ferguson)	-0.74	0.1 - 2
Great Eggleshope Beck (Carling)	-0.82	1 - 25
East Fork, Snake & Clearwater (Andrews)	-0.87	0.3 - 6
Sagehen Creek (Andrews & Erman)	-1.	1 - 4

Figure 1:

Figure 2:

A. REFERENCE TRANSPORT
(Parker et al., 1982)

B. NUMBER OF GRAINS
IN 1 HOUR
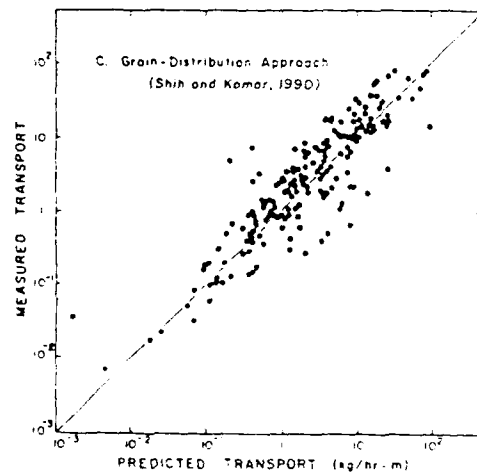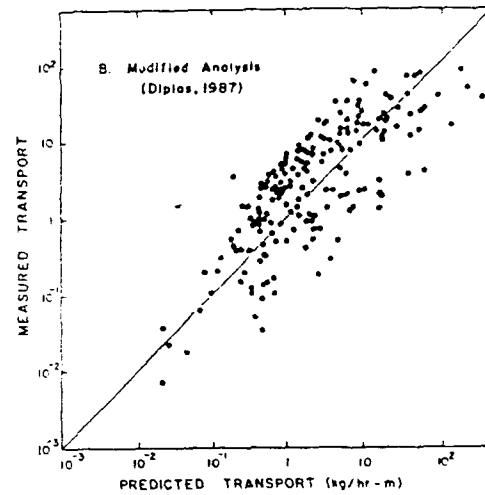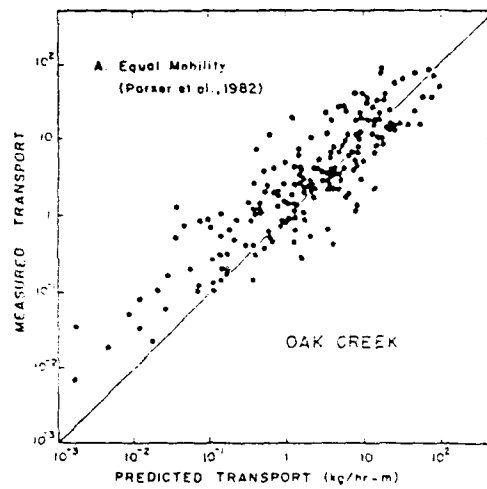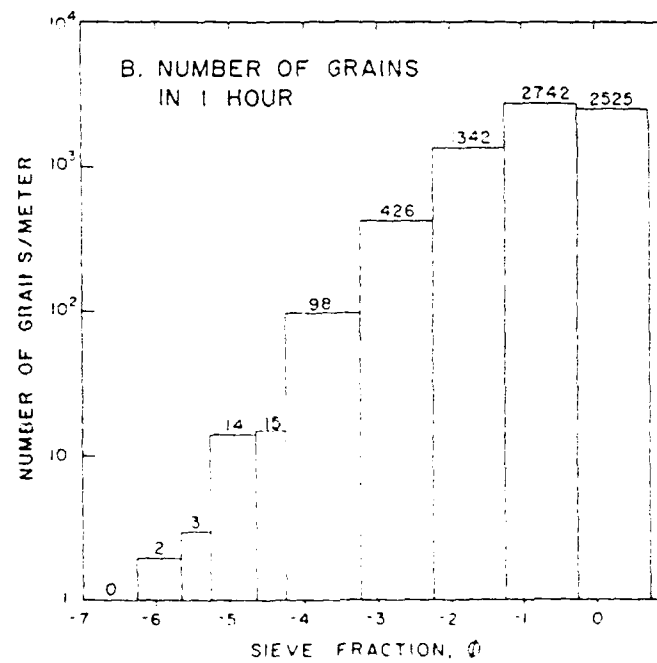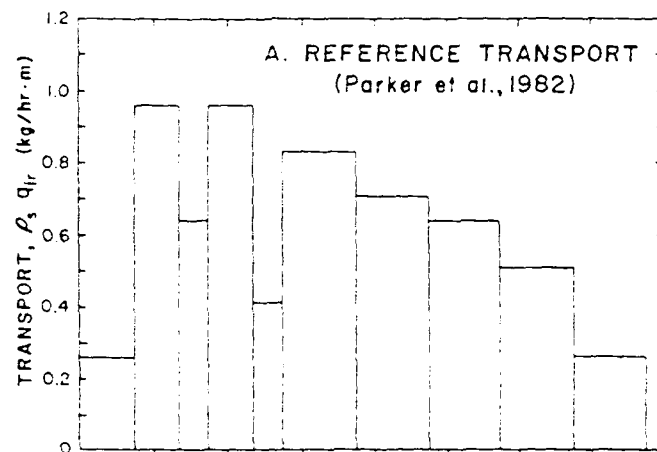
Predictions of Bedload Grain-Size Distributions

Figure 3:

A. Diplas (1987)
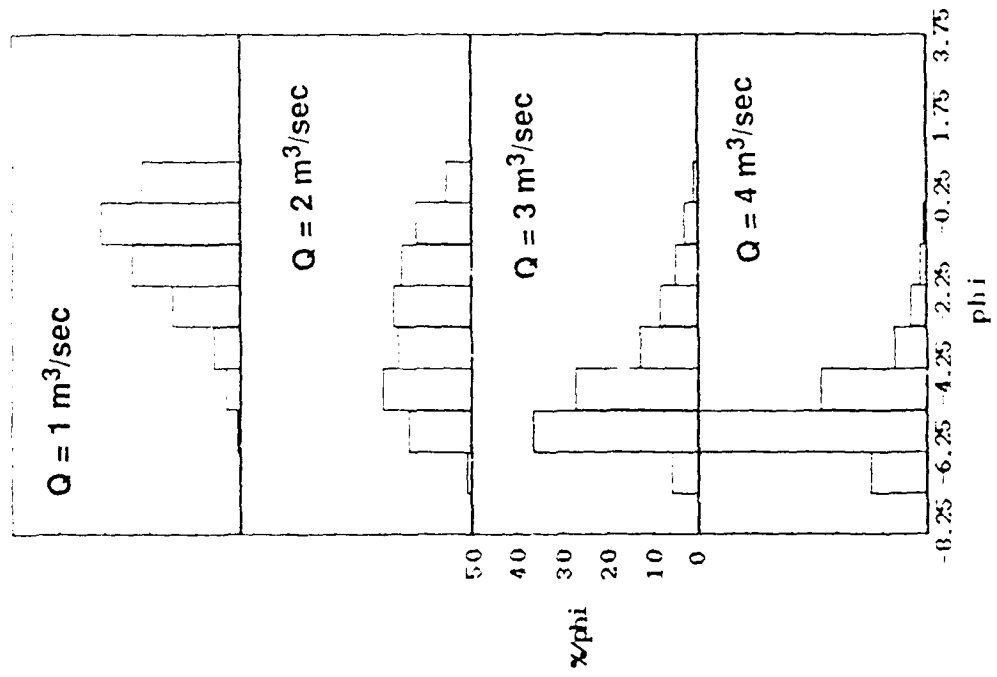
B. Shih and Komar (1990)
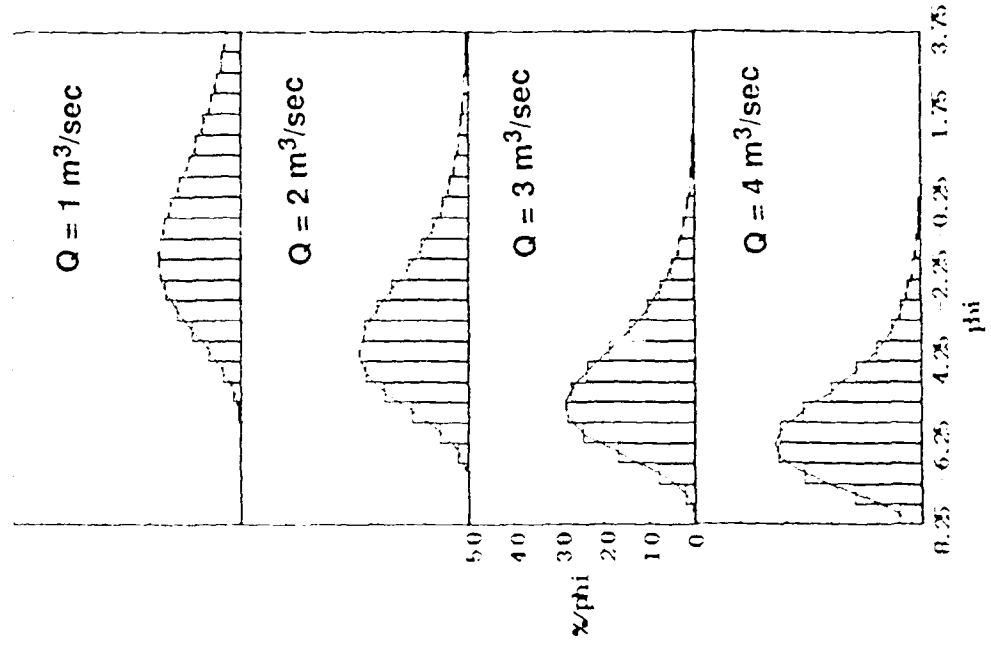
BEDLOAD GRAIN-SIZE AND TRANSPORT DISTRIBUTIONS

(Shih and Komar, 1990)

A. Rosin Grain-Size Distributions

$$f(D_i) = 100\,(s/k)(D_i/k)^{s-1}\exp[-(D_i/k)^s]$$
where   k = grain size at peak frequency
          s = spreading coefficient
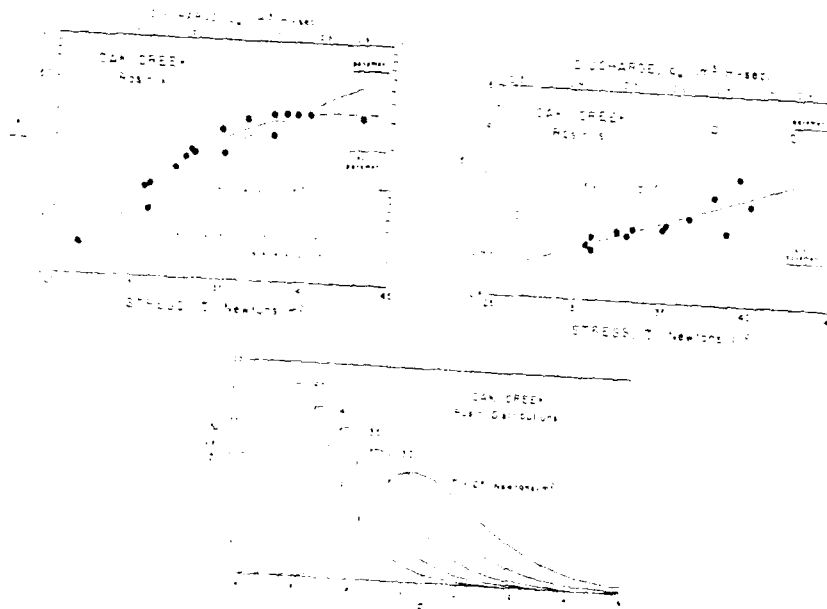
B. Total Bedload Transport Rate, q_s
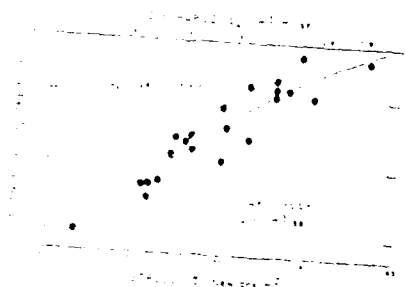
Note:

This figure will be redrafted to simplify the graphs and to make the summary more thematic.

C. Fractional Transport Distribution

$$q_s(D_i,\tau) = q_s(\tau)\,f_i(D_i,\tau)$$
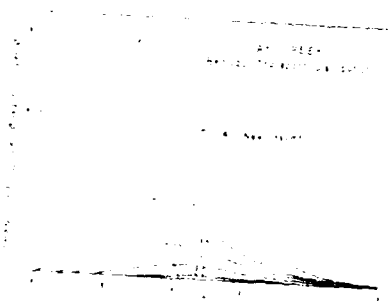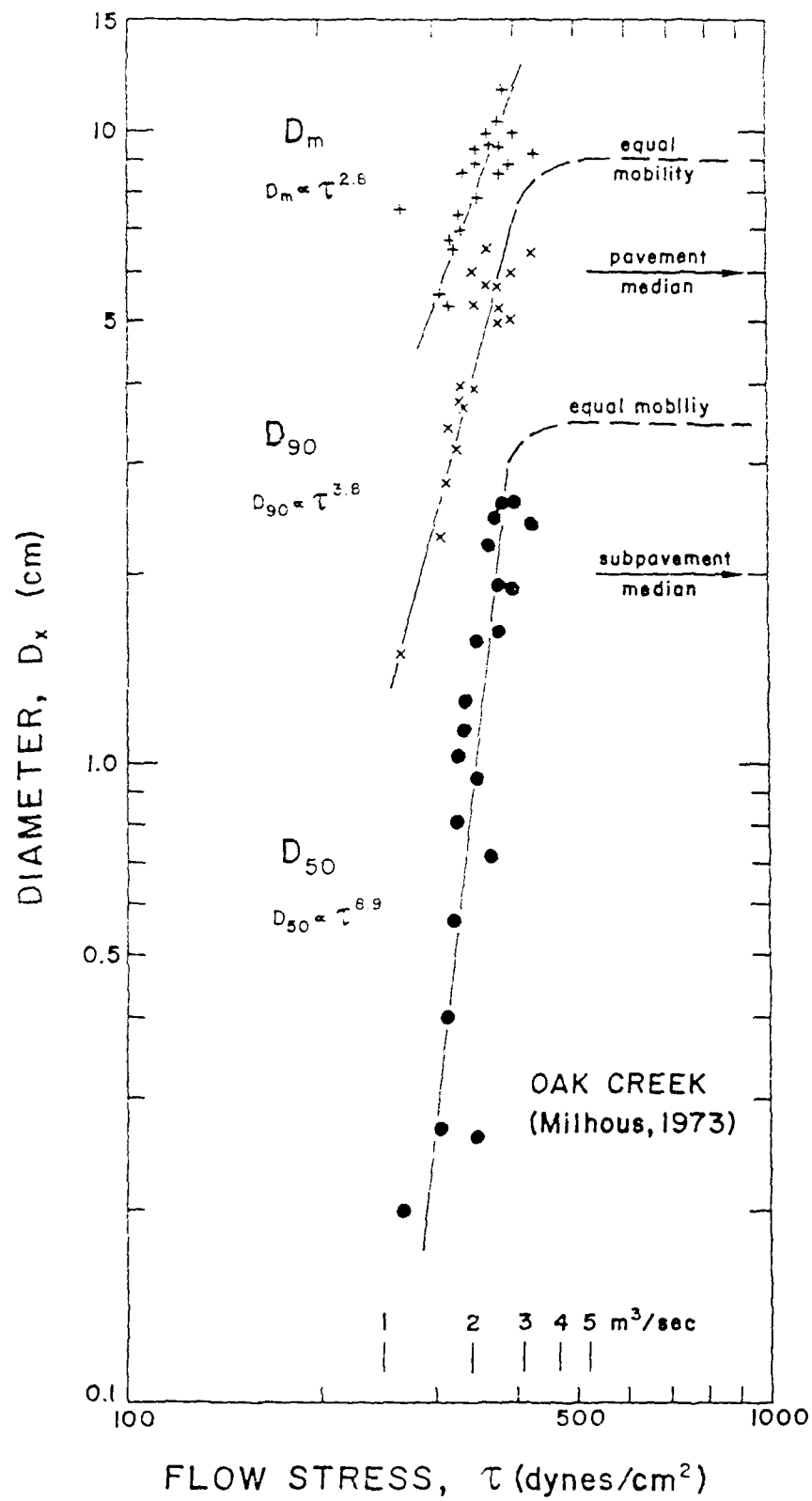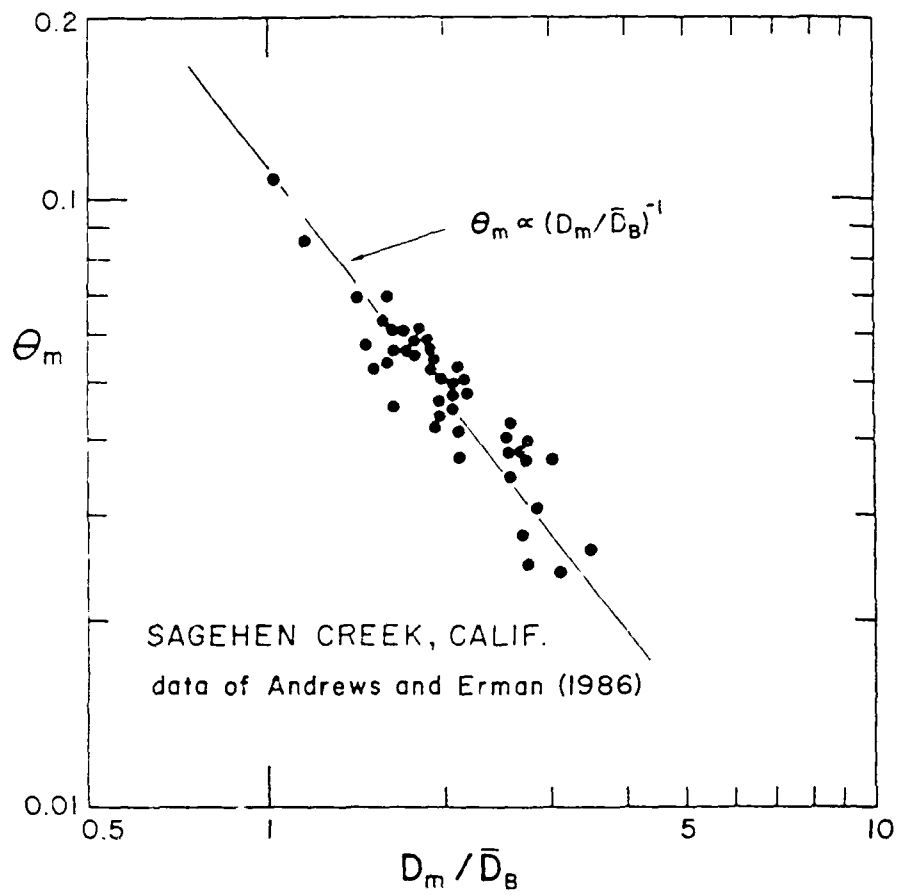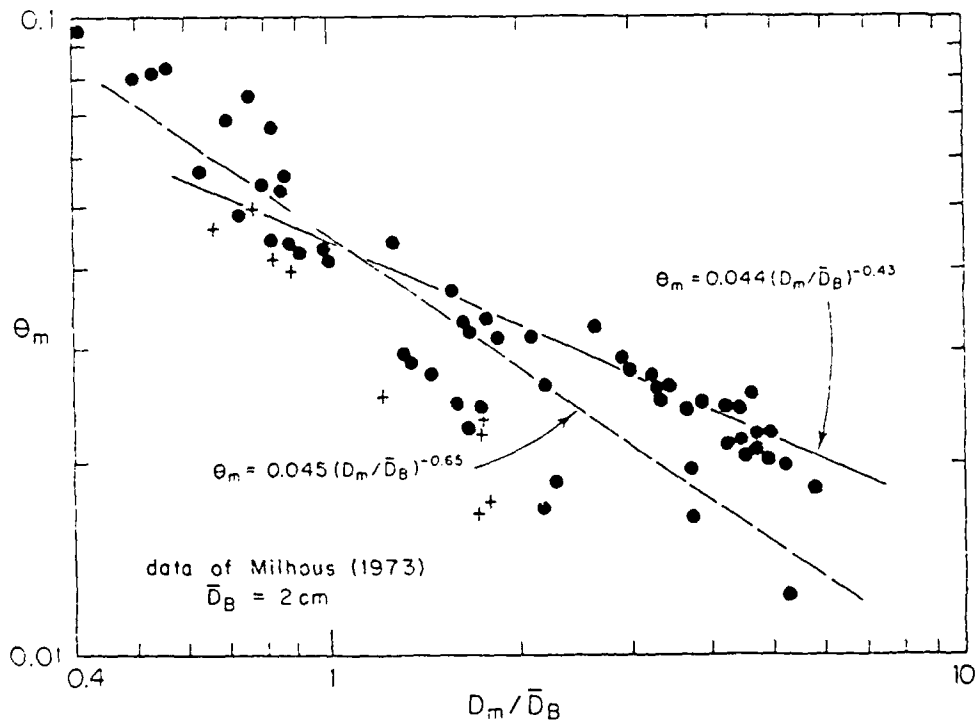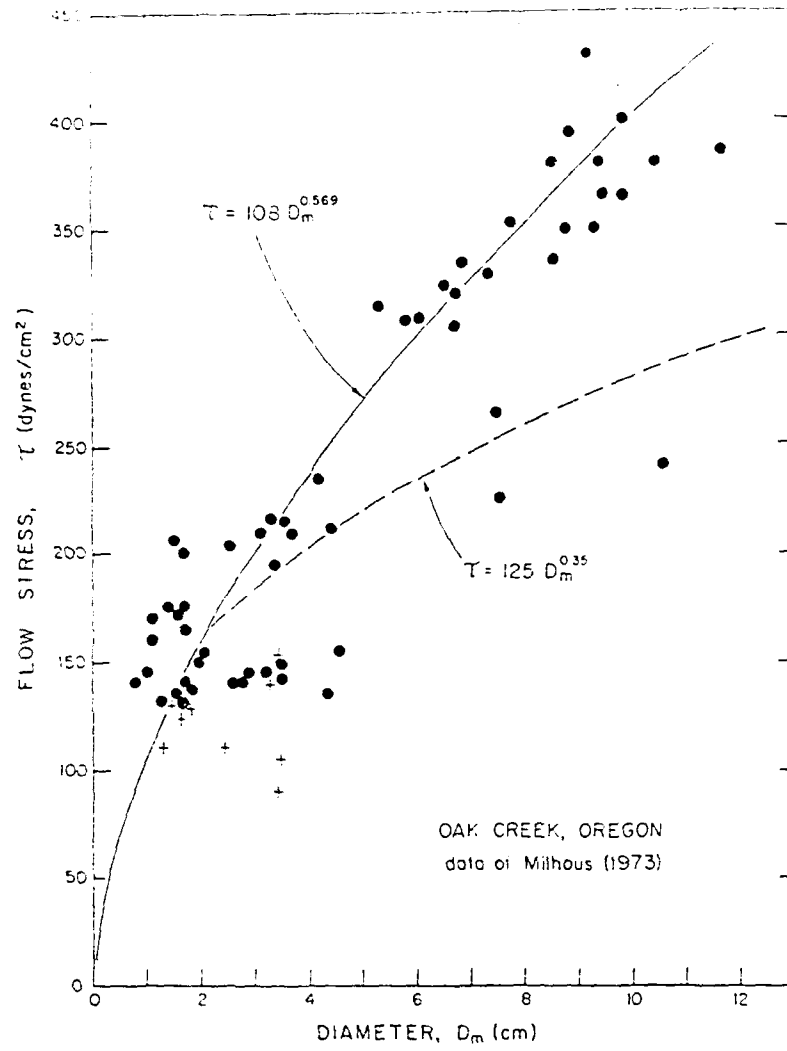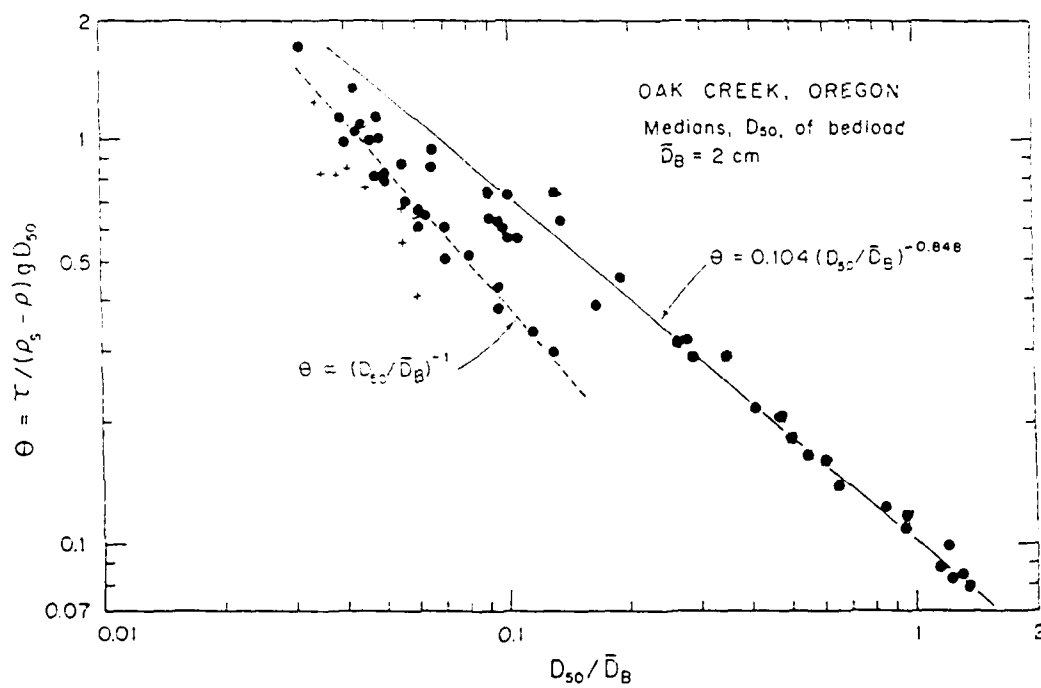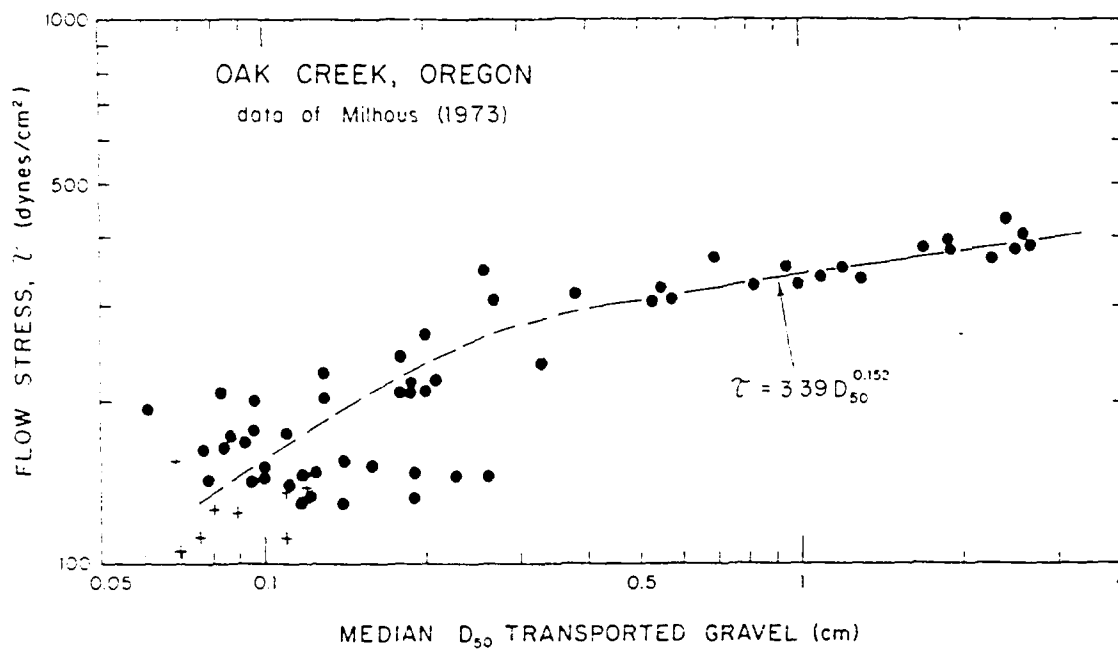
Figure 5:

Figure 6:

SAGEHEN CREEK, CALIF.

data of Andrews and Erman (1986)

$\theta_m \propto (D_m/\bar{D}_B)^{-1}$

Figure 7.

OAK CREEK, OREGON
data of Milhous (1973)

$\tau = 108\, D_m^{0.569}$

$\tau = 125\, D_m^{0.35}$

FLOW STRESS, τ (dynes/cm²)

DIAMETER, D_m (cm)

$\theta_m = 0.044\,(D_m/\bar{D}_B)^{-0.43}$

$\theta_m = 0.045\,(D_m/\bar{D}_B)^{-0.65}$

data of Milhous (1973)
$\bar{D}_B = 2$ cm

θ_m

$D_m/\bar{D}_B$

Figure 8:

BED-LOAD TRANSPORT OF MIXED-SIZE SEDIMENT

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INTRODUCTION

In order to predict the response of mixed-size sediment to flow, it is necessary to predict the transport rates of the individual size fractions in the mixture. It has been well demonstrated that fractional transport rates depend strongly on relative grain size, which is most commonly expressed as D_i/D_m, where D_m is a central value of the bed grain-size distribution, commonly D_{50}. But D_i/D_{50} does not uniquely specify the relative grain size of individual fractions, so it is necessary to ask whether relative grain size effects are well described by D_i/D_{50} alone. Using a lognormal size distribution as an example, it is not immediately clear that the fractional transport rates for the fraction $D_i/D_{50} = 2$ will be the same when it is on the coarse fringe of a mixture (98% finer than for $\sigma_\phi = 0.5$) or closer to the center of the mixture (68% finer than for $\sigma_\phi = 2.0$). Can a hiding function based on a single relative grain size parameter (e.g. D_i/D_{50}) work for all sediment mixtures? Do fractional transport rates depend on the local bed shear stress τ_o and D_i/D_{50} in the same fashion for all size mixtures?

The goal of this paper is to investigate the effect on fractional transport rates of different sediment mixtures. The results are reported from experimental work in which transport and flow observations were made over a range of flows for eight sediments, including five with a mixture of sizes. In three mixtures, only the mixture sorting was varied while the grain-size distribution shape and mean size of the sediment were held constant. In three other mixtures, the effect of bimodality was examined by using three bimodal sediments in which the proportion in each mode and the spread between modes was varied. Observations of the mixture effect on transport are also extended to other mixtures using field and flume data collected by others.

As in most things, it is true in sediment transport work that one sees only what one looks for. Beyond the basic experimental plan leading to the choice of sediment mixtures used in the experiments, the results reported here depend on the suite of variables measured in the transport experiments. The experiments and measurement program were designed to investigate the mixture effect on:

(1) Incipient motion of individual size fractions (by clustering flows near initial-motion conditions).

(2) Established motion of individual size fractions (by varying the flow by a factor of 2.5 or more in τ_o)

(3) The bed surface texture (using magnetic spray paint samples of the bed surface)

(4) The bed configuration (using longitudinal profiles and bed descriptions)

The last three observations provide the opportunity to investigate the interaction among bed surface texture, bed configuration, and fractional transport rates, particularly as it effects our ability to predict mixed-size sediment transport.

EXPERIMENTAL MATERIALS, APPARATUS, AND METHODS

We report here on 58 experimental runs made with eight different sediments. General hydraulic and transport parameters for each run are given in Table 2. The experiments were conducted so that flow depth was held within a small range about 11cm while τ_o varied over a wide range.

Sediments

The heart of the experimental program is the range of different sediment mixtures used. General properties of these mixtures are shown in Figure 1 and Table 1. Table 1 also includes properties of sediments used by others whose transport results will be discussed in this paper. The effect of mixture sorting alone was examined using three sediments with the same mean size (1.83mm) and grain-size distribution (lognormal), but different values of mixture standard deviation $\sigma_\phi = 0.2\phi$, 0.5ϕ, 0.99ϕ for the mixtures MUNI, 1/2ϕ and 1ϕ respectively, where σ_ϕ is the standard deviation of the size distribution when expressed in ϕ units ($\sigma_g = 2^{(\sigma\phi)}$, where σ_g is the geometric standard deviation of the size distribution expressed in mm). To provide comparison with the better understood unisize case, two additional, nearly unisize, sediments ($\sigma_\phi \approx 0.2\phi$) were used. These are named FUNI and CUNI, and fall toward the fine and coarse ends of the 1ϕ mix.

The next set of sediments reported on here were bimodal and incorporated parts of the nearly unisize FUNI MUNI, and CUNI sediments. In considering additional mixture properties whose effect on transport we wished to examine, bimodal sediments were chosen for several reasons. For the lognormal sediments, we found that mixture sorting by itself had little effect on fractional transport, once D_{50} and D_i/D_{50} were accounted for. Therefore, it seemed unlikely that higher orders of the bed sediment grain-size distribution would produce important mixture effects. This conclusion was further supported by the fact that we observed similar and consistent initial-motion conditions for a variety of different sediment mixtures used by others, including rectangular, fine-skewed, and weakly bimodal grain-size distributions. A more likely mixture property that would cause observed mixture effects to break down would be bimodality. How far apart can the modes of a bimodal sediment be separated until the relatively consistent behavior observed for unimodal sediments no longer occurs? Does the proportion in each mode make a difference? Because many gravel-bed rivers have bimodal grain-size distributions (e.g., Shaw and Kellerhals, 1982), an investigation of the effect of bimodality on mixed-size sediment transport is of interest directly, as well as part of a general study of mixture effects on transport. Three bimodal sediments were used, one with equal amounts of MUNI and CUNI (MC50), one with equal amounts of FUNI and CUNI (FC50), and one with 67% FUNI and 33% CUNI (FC70).

Apparatus

The experiments were conducted in two flumes, one at MIT and the other at the Johns Hopkins University (JHU). Both flumes have clear acrylic channels 0.6m wide and 0.3m deep. Some runs in the JHU flume were conducted with a 0.15m flow width. The MIT flume is considerably longer, with a working length of 22m compared to 11m at JHU. The operation of both flumes is identical. Water and sediment are recirculated separately. The water passes into a tailbox with no overfall and the mean flow depth is set by the volume of water in the flume. Uniform flow is obtained by adjusting the flume slope to match the water surface slope. All of the sediment is caught in a trap at the downstream end of the flume and recirculated with a small amount of water using an air-driven double-diaphragm pump. Sediment transport is sampled by intercepting the sediment/water mixture leaving the trap and passing it through a sieve that removes all of the sediment and allows the water to be recirculated. An open-air, gravity-driven sampling system was chosen to provide maximum flexibility in sampling methods and periods (sample intervals from 10s to 4 hours were used) and to allow convenient removal of samples for size analysis. The capability of sampling the entire transport load and removing samples of this load for size analysis are requirements for accurate mixed-size sediment transport work.

An upstream portion of each flume bed was artificially roughened to allow the boundary layer to become developed before encountering the loose sediment bed. Water discharge was determined using calibrated head-loss

sections in the water return pipe. Mean flow depth and water surface slope were determined using a point gage mounted on a cart that traversed rails parallel to the flume floor.

Samples of bed surface texture were taken by spray-painting an area of the bed 12.7 cm in diameter with a mixture of paint and fine magnetite. Surface grains (defined as those having any paint on them) were removed from the sample with a strong hand magnet. Some additional separation was necessary to remove unpainted grains that had adhered to painted grains during removal of the sample from the bed. This technique (Wilcock and Stull, 1989) was developed because manual removal of painted grains finer than 1 or 2mm was found to be impossibly tedious. We prefer the magnetic-paint technique because it allows direct observation of the sample and thereby provides an accurate check that all grains on the surface have been sampled, and that the sample contains only those grains that were on the bed surface. A considerable controversy exists as to whether areal surface samples, such as those collected by the paint or clay techniques, may be compared with samples taken volumetrically, such as by removing a predetermined volume of sediment from the bed (Church et al., 1987; Diplas and Sutherland, 1988). We have chosen to sidestep this problem by making comparisons only among samples taken with identical techniques. Transport samples are compared only with each other and use the volumetric grain-size distribution of the sediment mixture as a reference. Surface samples are compared only to other surface samples and use as a reference surface samples taken with the same method from the bed at its start-up condition. Although limiting ourselves in this fashion prevents us from making direct comparisons between particular transport and bed-surface grain-size distributions, it does allow us to make unambiguous comparisons of the different transport and bed surface grain-size distributions, which is one of our goals in this paper.

Procedure

The sediment was entirely homogenized and screeded to a plane bed before each run. After filling with water, a run was begun by adjusting the flume slope and discharge to a value close to that needed for uniform flow. Flow and transport measurements were made after long-term steady transport became established. In early experiments (Wilcock and Southard, 1988), equilibrium was judged by the absence of long-term variability in the flow properties (related to the development of a stable bed configuration) and in the size distribution of the transported sediment (related to the development of a stable grain-size distribution on the bed surface). To avoid any significant impact on the sediment bed and transport rates during the run, intermediate samples were not taken and final equilibrium was assumed to have been established when the run duration was greater than or equal to sufficient values in previous runs. While the transport sampling was underway, the water-surface elevation was read and the head loss in the return pipe was measured for later conversion to water-surface slope and discharge. After each run, the flume was drained, the bed was described and photographed, a profile of the bed elevation along the flume centerline was made, and bed-surface samples were taken.

Estimating Bed Shear Stress

Because bed forms were present in many runs, an estimate of the skin-friction portion τ_o' of the total bed shear stress was necessary. Different models were evaluated in an attempt to derive the best estimate of the skin friction. Unfortunately, very limited means exist to evaluate the accuracy of skin-friction estimates. The only sensitive test available to us was to evaluate the degree to which a given skin-friction estimate decreases the observed scatter in the transport rate. In this regard the drag-partition model of Einstein (1950) performed best. This model involves a fictitious flow hydraulic radius R' that corresponds only to the skin friction and is computed iteratively from a Keulegan-type profile

$$\frac{U}{\sqrt{gR'S}} = \frac{1}{\kappa} \ln\left(\frac{R'}{2.718z_0}\right)$$

(1)

where U is mean channel velocity, g is the acceleration of gravity, S is the water slope, κ is von Karman's constant, taken here to be 0.4, and z_0 is roughness length characteristic of the sediment bed. A value of $z_0 = D_{65}/30$ was found to give a good approximation of the roughness for the unimodal sediment beds for flows with plane beds and little or no transport. For the bimodal beds, the mean grain size of the coarse mode (5.66mm) was used in place of D_{65} because the coarse grains clearly protruded from the bed at low flows and presumably controlled the bed roughness. Because the primary use of the bed shear stress data in this paper is to determine a critical shear stress, no corrections were made for an increased roughness attributable to moving grains. To provide a good estimate of bed shear stress and transport rates near incipient motion, several runs with each sediment were clustered about initial-motion conditions. The transport rates of these plane-bed runs fell near the reference value used to estimate the critical shear stress of each fraction. At least two runs with a plane bed were conducted for each sediment. For these runs, $\tau_o' = \tau_b$, the total bed shear stress, and τ_o' can be determined with much better accuracy than for runs with bed forms. For plane-bed runs, values of τ_o' were found to be nearly identical to values of τ_b given by sidewall correction formulas (Vanoni and Brooks, 1957; Shimizu, 1990). For consistency, (1) was used to compute τ_o' for all runs.

INITIAL MOTION

Method

The initial-motion conditions, or critical shear stress τ_{ci}, of individual fractions in a mixed-size sediment have typically been determined using one of two different classes of methods. The first method involves estimating τ_{ci} as a reference bed shear stress τ_{ri} that produces a small transport rate for each fraction. Transport rates of individual fractions are measured for a number of flows and the shear stress that corresponds to the reference transport rate is determined from a fitted relation between dimensionless shear stress and dimensionless transport rate for each fraction (Parker et al., 1982; Day, 1980b). The second method involves determining the largest clast in a sediment mixture that is moved by a given bed shear stress. The largest grain displaced may be measured directly using the largest clast found in a transport sample (Andrews, 1983; Carling, 1983) or measured visually by observing the largest grain moving over an area of the bed (Hammond et al., 1984). The largest mobile grain is assumed to represent initial-motion conditions if coarser grains are available in the bed. Wilcock (1988) describes and compares these methods in some detail. In this paper, the reference-transport method is used to estimate τ_{ci}. This is because the largest-grain method suffers from serious sample-size effects and because it may not be used to properly scale the estimates of τ_{ci} so that they may be compared between fractions (Wilcock, in review). The reference transport method was also used because, for some of the mixtures, all sizes were found to be in transport even at the lowest transport rates. Under these conditions, the largest-grain method provides no information on τ_{ci} for any fraction other than the coarsest and may not be used.

The reference transport method used here to estimate τ_{ci} is based on that used by Parker et al. (1982) wherein the reference transport rate is defined as a constant value of the fractional-transport parameter W^*_i,

$$W^*_i = \frac{(s-1)g\, q_{bi}}{f_i\, (\tau_o/\rho)^{3/2}}$$

(2)

where s is the ratio of sediment and water density ρ_s/ρ, f_i is the proportion of fraction i in the sediment bed, q_{bi} is the fractional transport rate computed as $p_i q_b$, p_i is the proportion of fraction i in the transport, and q_b is mass

transport rate per unit width. A reference transport method based on the Ackers and White (1973) transport model has also been used by Day (Day, 1980b; White and Day, 1982). A comparison of the two methods is given by Wilcock (1988) and Wilcock and Southard (1988). The Parker et al. method is used here because grain size is not present in W^*_i and the τ_{ri} values are, therefore, undistorted with respect to grain size, a considerable convenience when comparing τ_{ri} among different size fractions.

Values of τ_{ri} were determined by fitting a transport relation to plots of $W^*_i - \tau^*_i$ for each fraction, where τ^*_i is the Shields parameter $\tau_o/(s-1)\rho gD_i$ formed using the fraction size. To preserve the grain size independence of the fitting procedure, a single transport curve was fitted to the data for each fraction and this curve was translated only parallel to the τ^* axis. Because the transport data show a concave-downward trend, a curved transport relation was used. The relation chosen is a power approximation of the Einstein (1950) bed-load function at low stresses derived by Parker (1979)

$$W^*_i = 11.2 \left(1 - \frac{0.8531\tau^*_{ri}}{\tau^*_i}\right)^{4.5}$$

(3)

This curve was fitted by eye to the transport data for each fraction. Its position was primarily determined by the position of data points located near the reference transport rate $W^*_{ri} = 0.002$. The accuracy of the fit is dependent on points found near or below W^*_{ri}, so some runs with each mixture were clustered at low transport rates. The fit of (3) to the data at higher transport rates was generally good for our data, although this fit was of minor concern because of the presence of data points near W^*_{ri}. The fit of (3) to the $W^*_{ri} - \tau^*_{ri}$ data of others was more important, because some of these data did not include transport rates near W^*_{ri}. Under these conditions, the use of a curved transport relation such as (3) provided a more reasonable estimate than the straight log line used previously (Parker et al.; 1982). For each fraction, an estimate of the error in τ^*_{ri} was determined by recording values of τ^*_{ri} when the transport curve was moved as far as possible to the right and left while still providing some conceivable fit to the fractional transport data. Because of the subjective nature of the process, these error bounds were made as large as possible. The error bounds were typically on the order of 10% and do not affect the conclusions drawn in this paper. For clarity, the error bounds are not shown on the figures presented here.

Test of Sorting

Figure 2 presents τ_{ri} for the two lognormal mixed-size sediments and the three unisize sediments. Comparison of τ_{ri} between fractions of the same size (and therefore the same D_i/D_{50}) in the $1/2\phi$ and 1ϕ mixtures clearly demonstrates that the relative-size effect on τ_{ri} can be well characterized by only one relative-size parameter, in this case D_i/D_{50}. Even though the sorting differs by a factor of two between the two mixtures and, therefore, the percentile value for a given fraction differs between the two mixtures, τ_{ri} is identical for two fractions with the same D_i (and D_i/D_{50}). Beyond showing that mixture sorting does not affect τ_{ri} for these mixtures, once D_i/D_{50} is specified, Figure 2 also shows that for these sediments τ_{ri} does not vary with D_i (or D_i/D_{50}). This is part of the equal mobility concept, which will be discussed further below. Because only mixture sorting varies between the $1/2\phi$ and 1ϕ mixtures, Figure 2 clearly demonstrates that mixture sorting does not effect τ_{ri}, once the effect of D_i/D_{50} is taken into account. It also shows that the absence of a sorting effect is evident in sediments that range from poorly sorted to unisize.

Figure 3 presents initial-motion relations in terms of τ^*_{ri} against S^* ($=[D_i^3(s-1)g]^{1/2}/v$), a convenient alternative to the grain Reynolds number found in the classic Shields curve. Shown in this plot are our data, along with our estimate of τ^*_{ri} for the other, previous published, data. On this curve, an absence of size dependence in τ_{ri}

plots as a line with slope of -3/2. The slopes of all these curves fall within the narrow range of -1.31 to -1.59, which means that the exponent b of the relation

$$\tau_{ri} = a \, (\frac{D_i}{D_{50}})^b$$

(4)

falls between 0.19 and -0.09. Only one of the values of b differs from 0.0 by more than 0.1. This is the value of 0.19 for sediment A of Day (1980), which was the only sediment for which no values of W^*_i fell close to W^*_r, making it the least certain of the sets of τ_{ri} values. The narrow range in the observed values of b is striking, because the experimental control behind our data in Figure 2 is not maintained in Figure 3. Even though the mixture sorting of the sediment in Figure 3 varies from 0.19ϕ to 2.1ϕ, there is no consistent trend between mixture sorting and the τ_{ri}-D_i/D_{50} slope. This provides more general empirical support for our controlled experimental result. In addition to the broad variation in mean size and sorting evident in Figure 3, other potentially important independent variables, such as relative depth, grain Reynolds number, and mixture grain-size distribution, vary between the mixtures but do not apparently influence the observed τ_{ri}-D_i/D_{50} relations. Also shown on Figure 3 is the Shields curve for incipient motion of unisize sediment. Even though the mixtures examined here are broadly distributed across the plot and extend to varying degrees into the transitionally rough and smooth turbulent ranges of the figure, the form of the τ^*_{ri} plots are remarkably consistent.

Although the relative-size effect is well explained by D_i/D_{50}, the separation between the various curves on Figure 3 is a result of the absolute size of each mixture. Figure 4 presents a plot of τ_{r50} (τ_{ri} for D_{50}) against D_{50} for each mixture, which shows that the Shields value for D_{50} predicts τ_{r50} well (a value 88% of the Shields value provides the best prediction of τ_{r50}).

<u>Test of Bimodality</u>

Figures 3 and 4 suggest that remarkably consistent incipient-motion relations exist for individual fractions in a wide range of mixtures. One of our goals in performing experiments with bimodal mixtures was to see if these results might hold for a broader range of mixtures. Figure 5 present values of τ_{ri} for three bimodal mixtures, along with our unimodal mixtures. For the MC50 mixture, which contains equal parts of two modes separated in size by a factor of 2.8, the results are similar to those for the unimodal sediments. τ_{ri} is essentially the same for all sizes, and the critical shear stress for the entire mixture is well represented by the Shields value for D_{50}. These results are also similar to those we obtained using the fractional transport rates measured with four bimodal mixtures by Misri et al. (1984). In their experiments, the two modes were separated in size by a factor of 2.6. Our estimate of τ_{ri} from these data show that all sizes begin moving at a single shear stress that is well represented by the Shields value for D_{50}.

The mixtures FC50 and FC70, which are composed of the FUNI and CUNI sediments and separated in size by a factor of 8.0, show strikingly different results. The two different modes in each have significantly different values of τ_{ri}. For FC50, which has equal proportions of the two modes, the coarse fractions have τ_{ri} similar to MC50, but τ_{ri} for the fine fractions are smaller by a factor of almost three. The fine fractions begin moving at a much lower shear stress than the coarse fractions. For FC70, in which 67% of the bed sediment is composed of FUNI, values of τ_{ri} for both modes are smaller by a factor of about 1.5 relative to FC50. τ_{ri} values for the coarse fractions in FC70 are the same as those for identical fractions in the 1ϕ mixture, suggesting the existence of a minimum value of τ_{ri} for coarse fractions in FC mixtures. It is important to note that the difference in τ_{ri} between FC50 and FC70 is due to the proportion in each size fraction only, and, therefore, cannot be accounted for by D_i/D_{50} alone. For FC50, the critical shear stress of the entire mixture is well represented by the Shields value for D_{50} (if D_{50} is approximated as the geometric mean of the two modes, even though no sediment of that size is found in the mixture). This is not

the case for FC70, in which D_{50} falls within the fine mode and the Shields value for this fraction is too low. Also, the Shields value for the geometric average of the two modes (which equals D_{50} for FC50, but not FC70) does not provide a good estimate for the critical shear stress of the FC70 mixture. The lower values of τ_{ri} for the FC70 mixture are consistent with the concept that a finer sediment bed will decrease the resistance to motion and increase the flow exposure of any given size fraction, but prediction of τ_{ri} for strongly bimodal sediments is clearly not as simple as for unisize, unimodal, or weakly bimodal sediments.

ESTABLISHED MOTION

The effect of relative grain size (and, hence, sediment mixture properties) is not necessarily limited to incipient motion only. Once a sediment is in motion, the transport rate of each fraction may be a function of D_i/D_{50} and mixture properties, even when size-dependent variation in τ_{ci} is accounted for. The degree to which mixture effects on fractional transport rates are evident depend on the manner in which the data are presented. Most commonly used is a dimensionless plot of transport rate versus the fractional Shields parameter τ^*_i. Such a plot, using W^*_i vs. τ^*_i, is shown in Figure 6a for the 1ϕ mixture. This plot immediately suggests that a simple similarity transformation may be used to collapse the data to a single trend, as has been done by Parker et al. (1982), White and Day (1982), and others. In the case of the 1ϕ mixture, the incipient motion results discussed above suggest that such a collapse may be achieved by simply replacing τ^*_{ri} with τ^*_i/τ^*_{ri}. For $\tau^*_{ri} = \tau^*_{r50}$, this transformation is equivalent to τ_o/τ_{50}. Unfortunately, this view of the fractional transport rates hides important aspects of the problem which can lead to errors of greater than an order of magnitude when predicting fractional transport rates. The chief problem is that the plot in Figure 6a obscures very large differences in q_{bi} from fraction to fraction. This problem is illustrated in Figure 6b, which presents the fractional transport rates for the same sediment and flows as Figure 6a, but in this case using a dimensional form of the fractional transport q_{bi}/f_i, as a function of grain size expressed as D_i/D_{50}. It is immediately apparent from Figure 6b that q_{bi} varies by as much as two orders of magnitude over all sizes for a given run.

Figure 6b may be thought of as a plot of the transport grain-size distribution expressed as p_i/f_i and scaled by q_b. This figure is well suited for examining the existence of equal mobility in the fractional transport rates. The complete definition of equal mobility is simply $p_i = f_i$, that is, the transport grain-size distribution is equivalent to that in the bed. In Figure 6b, equal mobility would appear as a straight line parallel to the abscissa.

A similarity collapse based only on adjusting τ^*_i for size-dependent differences in τ^*_{ri} cannot incorporate size-dependent differences in q_{bi}. Although Parker et al. (1982) showed that three different fractional transport functions were necessary to fit the fine, middle, and coarse portions of the Oak Creek transport data, the general approach to modeling fractional transport rates has been to assume that relative size effects on fractional transport rates are included only in their effect on τ_{ci} (Ashida and Michue, 1972; White and Day, 1982; Parker and Klingeman, 1982; Andrews and Parker, 1987). The goal of this section is to examine relative size and mixture effects on fractional transport rates. To do this, we will use plots like Figure 6b.

Test of Sorting

Figure 6b and 7 present the fractional transport rates for the 1ϕ and $1/2\phi$ mixtures. The legends in these figures use a value of τ_c for the entire mixture, which is straightforward in these and other cases where all sizes begin moving at nearly the same value of τ_o. At the lowest flow with the 1ϕ mixture ($\tau/\tau_c = 1.14$), the finer fractions show roughly equal mobility, but the coarse fractions do not. At higher flows, but still less that $\tau/\tau_c \approx 2$, both fine and coarse fractions are substantially underrepresented in the transport. Once τ/τ_c exceeds a value somewhere

between 1.7 and 2.4, all size fractions become equally mobile. The results for the $1/2\phi$ mixture are similar to those for 1ϕ, but merely appear stunted because of the smaller grain size range.

Figure 8 presents the fractional transport rates for representative samples of the Oak Creek transport data of Milhous (1973). To provide a clear plot of the 22 samples, the geometric means of several transport samples for each τ/τ_c value are plotted on Figure 8. The trend in these data is similar to that in our data, in that the fine and middle fractions are close to equal mobility at the lowest τ/τ_c, and become less mobile at higher τ/τ_c. The coarse fractions are underrepresented in the transport at all flows, which go to a maximum of $\tau/\tau_c = 1.43$.

<u>Test of Bimodality</u>

Figure 9 presents the fractional transport rates of the bimodal mixtures. The variation in q_{bi}/f_i for MC50 in Figure 9a is similar to that of 1ϕ and Oak Creek. At low values of τ/τ_c, both fine and coarse fractions are underrepresented. At values of τ/τ_c greater than roughly two, q_{bi}/f_i values for all sizes fall within a factor of two.

The trend in Figure 9b for FC50 is more complex than those of the better sorted mixtures because τ_{ri} varies with grain size in FC50. At $\tau/\tau_c < 4$, the finest fraction and all of the coarse fractions are underrepresented in the transport. In the range $4 < \tau/\tau_c < 6$, nearly equal mobility appears to hold for all fractions. This value for the transition to equal mobility is twice that observed in MC50 and the unimodal mixtures. However, if τ/τ_c is computed using the value of τ_c for the CUNI fractions (which is roughly 2.6 times that of the FUNI fractions and, therefore, the entire mix), the transition to equal mobility occurs at $1.70 < \tau/\tau_c < 1.95$. The trend in Figure 9c for FC70 is similar to that of FC50. The finest fraction achieve equal mobility at $\tau/\tau_c \approx 4$, whereas the coarsest fraction does not achieve equal mobility at τ/τ_c as high as 5.25 ($\tau/\tau_c = 2.35$ using τ_c for the coarse fractions).

That all sizes should achieve equal mobility at some large value of τ/τ_c has been suggested by Parker and Klingeman (1982), who demonstrated that, if the same exponent in $q_{bi} \propto (\tau/\tau_c)^\beta$ holds for all fractions at large values of τ/τ_c, the transport rates of each fraction will asymptotically approach the same value at large τ/τ_c. Using the common value of $\beta = 1.5$, the value of τ/τ_{ci} at which all sizes would achieve the same transport rate is in the range of 10 and 50, if τ/τ_{ci} is computed using the largest value of τ_{ci} for the mixture. That equal mobility is observed in real mixtures at considerably smaller values of τ/τ_c is a result of relative size effects on the fractional transport rates.

BED CONFIGURATION

Although much information has been collected on the effect of grain size on bed configuration in unisize beds, little is known about the effect of sediment mixtures on bed configuration.

A consistent sequence of bed forms was observed in runs with the lognormal sediments. At the lowest flows, transport occurred over a plane bed. At higher flows, long low dunes appeared that were regular in appearance with straight, flow-perpendicular crests. The height of these bed forms increased with flow strength to a maximum of several centimeters, and the spacing between bed forms decreased with flow to values on the order of 80 to 120 cm. Taken together, these trends caused the bed forms to steepen with increasing flow (Figure 10). The bed forms also became more irregular in shape at higher flows, becoming quite three-dimensional at the highest flows.

The plane bed—dune sequence in the bed configuration is consistent with that expected for sediments with a mean size of 1.83 mm (Southard and Boguchwal, in press). There is also a sorting effect evident in Figure 10. For the unisize mixture MUNI, the plane-bed to dune transition occurred at a higher flow strength than in the mixed-size sediment, and the MUNI bed forms were shorter and steeper than those in the mixed-size sediments. Bed forms appeared at the weakest flow strength in the poorly sorted 1ϕ mixture, and the steepness of the bed forms at strong flows was smaller than in the better sorted sediments. The range of sizes in the mixture (and therefore the sorting)

may play two roles in explaining this trend. The presence in the mixture of fine grains, which tend to form shorter, steeper bed forms at lower flows than coarser grains, may trigger bed-form development at lower flows than would be the case with a better sorted mixture of the same mean size. The presence of coarser grains in the mixture may limit the depth of trough scour and, therefore, bed-form height and steepness at higher flows.

The bed forms found in the bimodal sediments are significantly different and more complex than those found in the lognormal sediments. For the MC50 mixture, a plane bed was observed at the lowest flows. Bed-load sheets were evident over a narrow range of τ, followed by dunes at the highest flows we observed. The bed-load sheets were similar to those observed in the field by Whiting et al. (1988) These bed forms are somewhat ephemeral and characterized by a height equal to one or two CUNI grain diameters, a spacing of 1m to 2m, and, most distinctively, a spatial pattern of grain size segregation wherein the leading edge of the bed form is primarily coarse-grained, while the rest of the bed form is either well mixed or primarily fine-grained. Grain paths within these bed forms are distinctive in that many grains, particularly finer ones, often do not travel all the way to the downstream end of the bed form as they would in a dune or ripple, but are continuously deposited and reentrained on the back of the bed form.

At low flows with the FC mixtures, the FUNI fractions were transported over an essentially immobile bed of CUNI. At higher flows, the FUNI became organized into irregular, short, low dunes (typically 0.5cm to 1.5cm high, 15cm to 30 cm long, with a spacing of 30cm to 100cm) migrating over a coarse , planar bed composed of more CUNI than the bulk mix. These isolated FUNI bed forms persisted at the highest flows we used, even in the presence of substantial transport of the CUNI grains. The distinctive feature of these bed forms is that the CUNI grains were typically not deposited within the bed form, but instead traveled quickly over the fine-grained, bed-form surface. The proportion of CUNI grains trapped in the lee of the bed form varied as a function of bed-form size and shape.

The primary distinction between the MC and FC bed forms is the degree to which the different modes interacted in the bed configuration. In the MC runs (modes separated in size by a factor of 2.8), both modes were found within the bed forms. In the FC runs (modes separated in size by a factor of 8.0), sediment from the two modes did not substantially interact within the bed forms, even when the transport rates of the two modes were equivalently large. Instead, the FUNI grains formed well-sorted, short, steep dunes (with a bed-form height similar to that observed with a bed composed entirely of FUNI), while the CUNI grains remained in planar sections of the bed (a plane bed is the equilibrium bed configuration for CUNI at the flows we examined). Because of the more complex and spatially variable bed configurations found with the bimodal sediment, it is not possible to construct a simple quantitative description of the mixture effect on the bed forms, as done with Figure 10.

BED SURFACE TEXTURE

Samples of the bed surface texture were taken following each run with the 1ϕ mixture. When bed forms were present, the samples were taken from the bed-form troughs. The troughs form the exposed portion of the surface over which the bed forms move and, therefore, represent the surface on which, or through which, size-dependent exchange of sediment may occur as the bed adjusts towards equilibrium.

Figure 11 presents the variation with τ/τ_c of the bed surface texture for the 1ϕ runs. Plotted on the figure are the percentage of grains on the bed surface finer than 4.0mm and 1.0mm. For reference are two lines that represent the equivalent values for the well mixed bed at the start of each run. The proportions of fine and coarse fractions in the final bed-surface texture increase with flow strength up to roughly $2\tau_c$. As τ_o increases, the fine fractions increase from a proportion smaller than present in the well mixed bed and approach a proportion equivalent to that of the start-up bed, while the coarse fractions become increasingly overrepresented on the bed surface with increasing

τ_o. At $\tau > 2\tau_c$ the coarse surface layer is completely broken up; both fine and coarse fractions are present on the bed surface in a proportion close to that of the well mixed bed.

Previous discussions of the development of an equilibrium coarse surface layer have been criticized because bed-surface samples were compared to the bulk mix, thereby involving a comparison between areal and volumetric bed samples (Parker, 1980; Parker et al., 1982b; Day and Egginton, 1983). Because of this sampling problem, the existence of an equilibrium coarse surface layer was questioned (Day and Egginton, 1983). Because the same sampling method was used for both the start-up and equilibrium samples, Figure 11 clearly shows that the bed-surface layer coarsens during adjustment towards equilibrium.

INTERACTION AMONG FRACTIONAL TRANSPORT RATES, BED CONFIGURATION, AND BED SURFACE TEXTURE

Figure 12 shows values of the percentage of the transport finer than 4.00mm and 1.00mm for the 1ϕ mixture at both equilibrium transport conditions and initial transport sampled shortly after each run was started with a completely mixed sediment bed. At values of $\tau < 2\tau_c$, when the bed and transport adjust from the well mixed start-up condition toward an equilibrium condition, the mobility of both the fine and coarse fractions consistently decreases with time, moving away from a condition of equal mobility. This process cannot be explained with a conceptual model that states that all size fractions must adjust toward equal mobility for equilibrium transport. However, an equilibrium transport system must operate at equal mobility only in the special case of a perfect feed system in which the material of the sediment bed is used as the feed material. Feed and recirculating flumes can produce very different equilibrium transport conditions, even when the same initial flow and sediment bed are used. In non-feed systems, equilibrium transport may occur at conditions far from equal mobility, as demonstrated in our experiments (Figure 6b, 7, and 9).

A clear interaction among fractional transport rates, bed configuration, and bed surface texture was observed in the 1ϕ runs. Bed forms were present in all but the two samples with the smallest τ/τ_c in Figures 11 and 12. Bed forms and a coarse surface layer can clearly coexist. The bed forms observed in the the highest two runs in Figure 11 were the largest and most irregular in shape. Greater irregularity in the magnitude of the scour in the lee of these bed forms contributed to the break up of the coarse surface layer. Comparison of Figures 11 and 12 shows that the presence of an armor layer coincides with an underrepresentation of the fine and coarse fractions in the transport, and that the break up of the coarse surface layer corresponds to the onset of nearly equal mobility in the fractional transport rates.

CONCLUSIONS

Incipient motion

Incipient motion occurs at nearly the same shear stress for all size fractions in a wide range of unimodal and weakly bimodal sediments. For these sediments, the critical shear stress for each size fraction is well represented by a value slightly smaller than the Shields value for D_{50} of the mixture. This observation was made for the Oak Creek data by Parker et al. (1982) and can be extended to a wide range of experimental sediment mixtures. Mixture sorting was demonstrated to have no effect on τ_{ci} for two sediments that differed only in their grain-size distribution standard deviation.

For strongly bimodal mixtures, the size independence of the fractional critical shear stress is no longer maintained when the modes are separated in size by a factor between 2.8 and 8.0. For modes separated in size by a factor of 8.0, we observe that the finer fractions begin moving at a shear stress considerably smaller than the coarse

fractions, and that the values of τ_{ri} depend on the proportion present in each fraction. These results suggest that incipient motion of individual fractions in bimodal sediments depends on at least four parameters: the absolute size of the sediment (represented, perhaps, by D_{50}), the relative size of each fraction (D_i/D_{50}), the separation in size between the two modes, and the proportion in each mode.

Established Motion

All sediments discussed here showed a similar pattern in fractional transport rates. Over a finite range of τ/τ_c, both fine and coarse fractions are underrepresented with respect to the bed grain-size distribution during equilibrium transport conditions. At higher values of τ/τ_c, all fractions show equal mobility. This consistency in the variation in fractional transport rates with τ_o is striking in that it arises from both lab and field data, from bed grain-size distributions covering a wide range of sorting, and for lognormal, fine-skewed, and bimodal grain-size distributions. The occurrence of distinctly non-equal mobility ($p_i \neq f_i$) is also striking for those sediments for which all sizes begin moving at nearly the same shear stress. The variation of q_{bi} with τ_o depends not only on τ_{ci}, but on relative grain-size effects controlled by the bed configuration and the presence of a coarse surface layer. Based on our laboratory data, the value of τ/τ_c at which all fractions achieve equal mobility appears to be about 2.0 for the lognormal and more narrow bimodal sediment, and in the range of 1.7 to greater than 2.35 for the more poorly sorted bimodal sediment.

Bed Configuration

Mixture sorting was observed to have a clear and measurable effect on the size and shape of bed forms in the unimodal sediments we investigated. Bed-form height was observed to decrease, while bed-form spacing and spacing/height ratio increased with mixture sorting. In the bimodal sediments, a different, more complex, and spatially variable sequence of bed forms was observed. For the sediments whose modes were separated in size by a factor of 8.0, the different modes appeared to act in a relatively independent fashion, with the fine fraction forming starved dunes and the coarse fraction forming a planar surface. For the bimodal sediment whose modes were separated in size by a factor of 2.8, a bed form sequence of plane bed — sheets — dunes was observed and involved both fractions.

Bed Surface Texture

A distinct coarse surface layer was unambiguously shown to develop as the 1ϕ unimodal sediment bed adjusted to equilibrium. We observed the coarse surface layer to become better developed as τ increases, up to the point where intense flow in the lee of bed forms broke up any surface grain-size sorting. At flows where bed forms and a coarse surface layer coexist, bed forms may contribute to the development of a coarse surface layer by increasing the supply of actively transported coarse grains. The coarse surface layer need not serve as a regulator of the mobility of different size fractions except in the special case of a sediment-feed flume. Its presence in our recirculating flume data is a result of size-dependent variation in grain mobility, development of a partial static armor, and a geometrically necessary vertical sorting process that occurs when at least some grains of all sizes are in motion.

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TABLE 1

Properties of Grain Size Distributions

Mixture	Mixture Type	Dm (mm)	D15 (mm)	D50 (mm)	D85 (mm)	Sorting (ϕ)	Sorting (geom.)	Reference
MUNI	≈unisize	1.87	1.63	1.86	2.20	0.20	1.15	*Wilcock (1987)*
1/2 Ø	lognormal	1.82	1.25	1.83	2.59	0.50	1.41	*Wilcock (1987)*
1Ø	lognormal	1.85	0.893	1.83	3.88	0.99	1.99	*Wilcock (1987)*
CUNI	≈unisize	5.31	4.68	5.28	6.17	0.18	1.13	*Wilcock (1987)*
FUNI	≈unisize	0.662	0.538	0.670	0.807	0.26	1.20	*Wilcock (1987)*
MC-50	Strongly bimodal	3.19	1.75	2.55	6.00	0.79	1.73	*This Paper*
FC-50	Strongly bimodal	1.88	0.600	2.00	5.88	1.51	2.85	*This Paper*
FC-70	Strongly bimodal	1.30	0.550	0.750	5.54	1.45	2.73	*This Paper*
DAY A	Weakly bimodal	1.50	0.336	1.82	6.05	1.77	3.41	*Day (1980a)*
DAY B	Weakly bimodal	1.17	0.318	1.57	3.49	1.45	2.73	*Day (1980a)*
MISRI N1	≈lognormal	2.37	1.15	2.36	4.91	1.00	2.00	*Misri et al. (1984)*
MISRI N2	Rectangular	3.78	1.21	3.81	11.9	1.44	2.71	*Misri et al. (1984)*
MISRI N3	≈lognormal	4.09	1.70	4.00	10.1	1.18	2.27	*Misri et al. (1984)*
SAF	Weakly bimodal	2.10	0.828	2.16	6.04	1.29	2.45	*Dhamotharan et al. (1980)*
OAK CREEK	Long fine tail	13.1	2.37	19.5	56.6	2.10	4.29	*Milhous (1973)*

Table 2. General Hydraulic Properties of Experimental Runs

Mix	Flume Width (cm)	Run	Water Discharge cms X10^3	Flow Depth cm	Mean Velocity cm/s	Slope X10^4	τ_0 Pa	qb initial g/ms	qb equil. g/ms	Bed Configuration
MUNI	60	A2	30.6	11.4	44.8	9.6	0.89	na	1.85E-03	Plane
	60	A3	35.3	11.4	51.8	12.4	1.15	na	3.59E-01	Plane
	60	A4	39.7	11.5	57.6	14.2	1.37	na	3.19E+00	2D dunes
	60	A5	43.6	11.8	61.8	21.5	1.73	na	8.91E+00	2D dunes
	60	A6	48.8	12.0	67.5	24.2	2.08	na	2.30E+01	2D dunes
	60	A7	55.7	12.8	72.4	29.3	2.53	na	4.64E+01	2D-3D dunes
1/2 Ø	60	B1	28.7	11.0	43.6	10.0	0.91	na	1.44E-02	Plane
	60	B2	30.4	11.2	45.3	10.3	0.87	na	3.70E-02	Incipient 2D dunes
	60	B3	35.6	11.2	52.8	12.6	1.17	na	1.12E+00	2D dunes
	60	B4	39.3	11.3	58.0	16.8	1.47	na	5.98E+00	2D dunes
	60	B5	43.4	11.6	62.2	21.4	1.78	na	1.37E+01	2D dunes
	60	B6	47.7	11.7	67.9	33.4	2.38	na	2.58E+01	2D dunes
	60	B7	54.9	11.5	79.4	49.2	3.74	na	9.70E+01	2D-3D dunes
1Ø	60	C1	28.6	11.1	43.1	10.4	0.97	na	2.31E-02	Plane
	60	C2	30.3	10.9	46.5	11.1	1.00	na	3.30E-02	Plane
	60	C3	35.0	11.2	52.4	18.2	1.32	na	2.32E+00	Incipient 2D dunes
	60	C4	39.1	11.2	58.0	21.4	1.63	na	1.07E+01	2D dunes
	60	C5	43.4	11.0	65.5	28.0	2.17	na	2.97E+01	2D dunes
	60	C6	47.8	10.9	72.9	33.0	2.78	na	5.94E+01	2D-3D dunes
	60	87-1	29.4	11.3	43.5	10.9	1.03	1.67E-01	1.59E-02	Plane
	60	87-2	31.5	11.1	47.5	12.5	1.17	3.42E-01	4.68E-01	Plane
	60	87-3	33.7	11.3	49.9	14.4	1.26	2.50E+00	1.96E+00	2D dunes
	60	87-4	35.9	11.2	53.4	18.4	1.36	9.94E+00	4.50E+00	2D dunes
	60	87-5	37.7	11.3	55.9	21.4	1.56	7.88E+00	9.69E+00	2D dunes
CUNI	60	D1	39.2	11.1	59.1	25.5	2.47	na	3.35E-03	Plane
	60	D2	43.2	11.0	65.4	31.0	2.99	na	3.45E-02	Plane
	60	D3	47.7	10.8	73.3	38.2	3.63	na	5.55E-01	Plane
	60	D4	54.8	11.1	82.6	49.1	4.77	na	9.55E+00	Plane
FUNI	60	E1	17.8	10.8	27.5	2.9	0.25	na	2.47E-03	Plane
	60	E2	19.6	10.8	30.2	3.2	0.26	na	2.45E-02	Plane
	60	E3	21.7	10.8	33.4	3.6	0.30	na	1.89E-01	Plane
	60	E4	25.2	11.2	37.7	5.8	0.38	na	1.75E+00	2D dunes
	60	E5	27.7	11.4	40.4	14.1	0.54	na	4.12E+00	3D dunes
	60	E6	30.2	11.2	44.9	17.5	0.72	na	6.31E+00	3D dunes
MC-50	15.6	1	8.9	11.1	51.6	29.8	1.8	na	3.28E-03	Plane
	15.6	2	9.5	11.2	55.2	31.9	2.0	na	2.22E-02	Plane
	15.6	7	10.2	11.4	57.5	29.8	2.1	na	1.12E-00	Plane
	15.6	6	11.8	11.0	69.1	43.0	3.1	na	1.80E+01	Spatial size segregation
	15.6	3	12.9	10.7	77.8	60.5	4.0	na	6.30E+01	Bed-load sheets
	15.6	5	14.2	10.5	86.8	75.7	5.0	na	1.31E+02	Isolated trains of 2D dunes
	15.6	4	15.4	10.6	93.3	99.9	6.0	na	2.23E+02	Isolated trains of 2D dunes
FC-50	60	12	22.3	11.3	33.0	6.8	0.64	na	6.90E-03	Plane
	60	11	25.1	10.9	38.4	8.9	0.85	na	3.47E-02	Plane
	60	10	30.5	11.5	44.3	10.2	1.09	na	2.31E-01	Plane
	15.6	7	6.7	11.1	39.1	16.2	1.04	na	3.17E-01	Plane
	15.6	2	7.4	10.8	44.6	19.6	1.32	na	1.15E+00	Plane
	15.6	1	8.6	11.2	49.7	25.8	1.67	na	2.56E+00	Plane
	15.6	5	9.0	10.8	53.8	26.8	1.89	na	8.00E+00	Incipient FUNI dunes
	15.6	4	10.2	11.1	59.7	30.9	2.28	na	1.33E+01	Isolated FUNI dunes
	15.6	3	11.1	10.8	66.0	42.4	2.89	na	3.88E+01	Isolated FUNI dunes
	15.6	6	12.1	10.7	72.8	42.5	3.32	na	7.95E+01	Isolated FUNI dunes
	15.6	8	13.7	11.0	80.7	46.8	3.95	na	1.32E+02	Isolated FUNI dunes
FC-70	60	1	19.4	11.2	28.8	4.7	0.47	na	1.80E-03	Plane
	60	2	22.7	11.3	33.7	5.3	0.61	na	5.02E-02	Plane
	60	3	24.7	11.2	36.8	6.2	0.73	na	2.40E-01	Plane
	60	4	26.8	11.2	39.8	6.3	0.82	na	5.08E-01	Plane
	60	5	29.2	11.2	43.7	8.1	1.00	na	1.06E+00	Plane
	60	6	32.5	11.2	48.6	9.6	1.23	na	3.67E+00	Isolated FUNI dunes
	60	8	37.9	11.2	56.5	15.2	1.73	na	1.42E+01	Isolated FUNI dunes
	60	7	40.5	11.1	60.6	15.8	1.94	na	2.65E+01	Isolated FUNI dunes
	60	10	45.4	11.1	68.2	20.4	2.47	na	4.91E+01	Isolated FUNI dunes

FIGURE 2. Reference shear stress for unimodal sediments
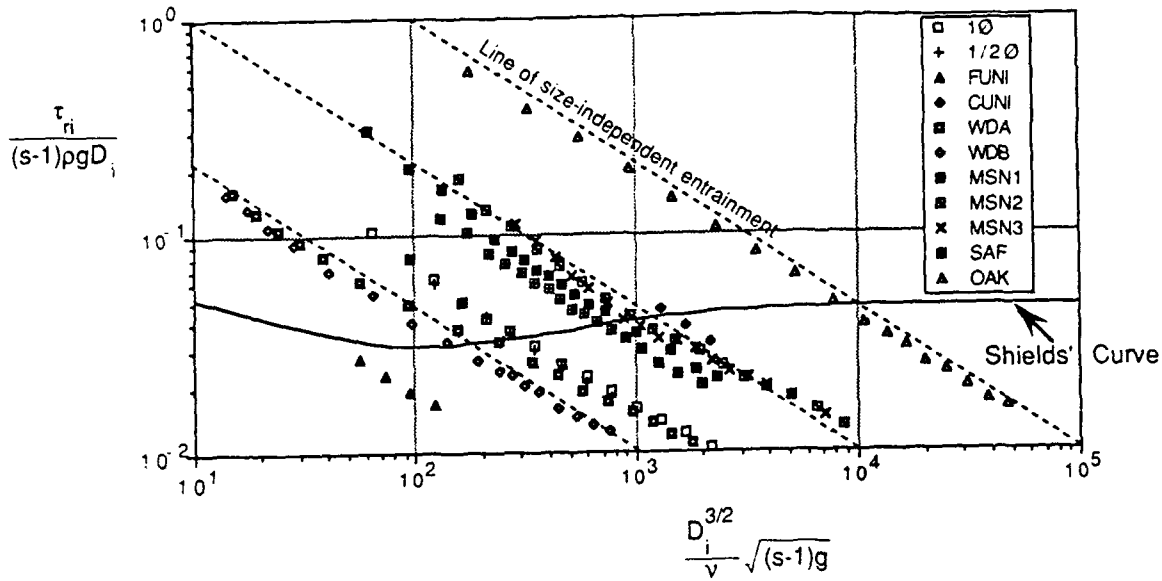

FIGURE 1. SUMMARY OF EXPERIMENTAL MIXTURES
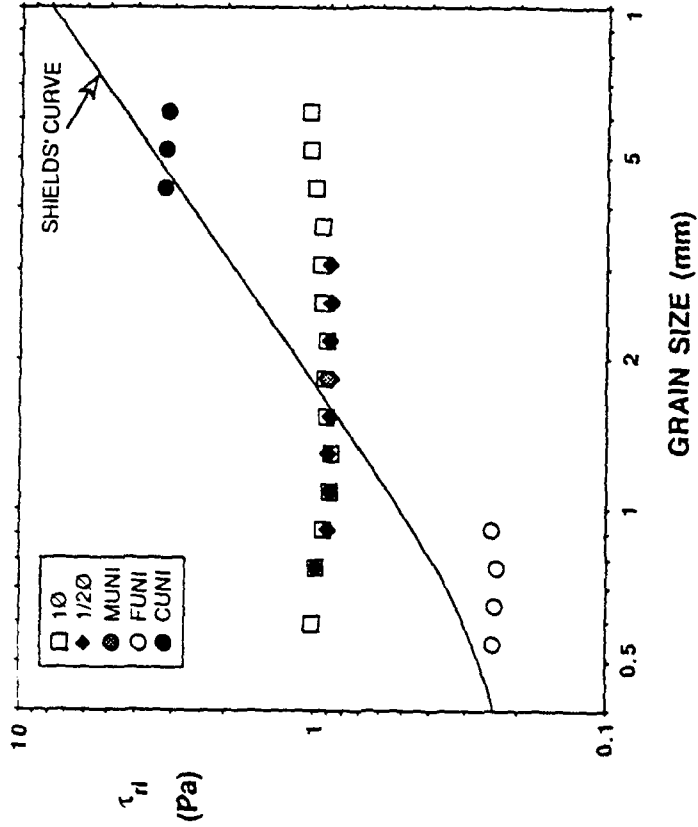
Figure 3: Reference Shields parameter vs. S*: Unimodal and weakly bimodal sediments

After Wilcock and Southard (1988)
Data Sources given in Table 1

Figure 4: Reference Shields' Parameter for D50 of mixtures shown in Figure 3.
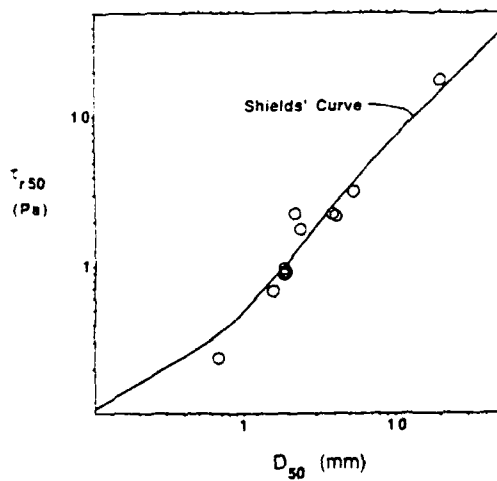

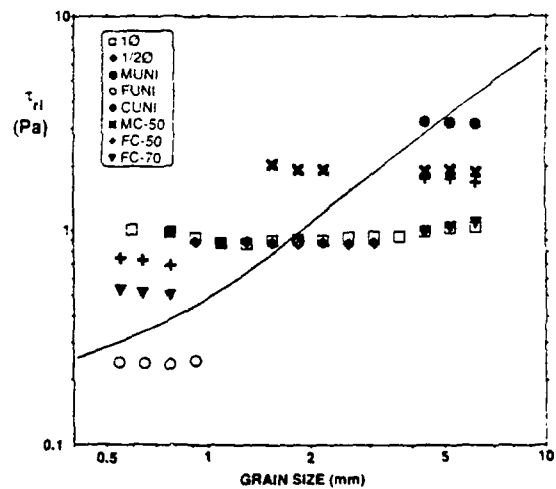Figure 5. Reference shear stress for unimodal and bimodal sediments

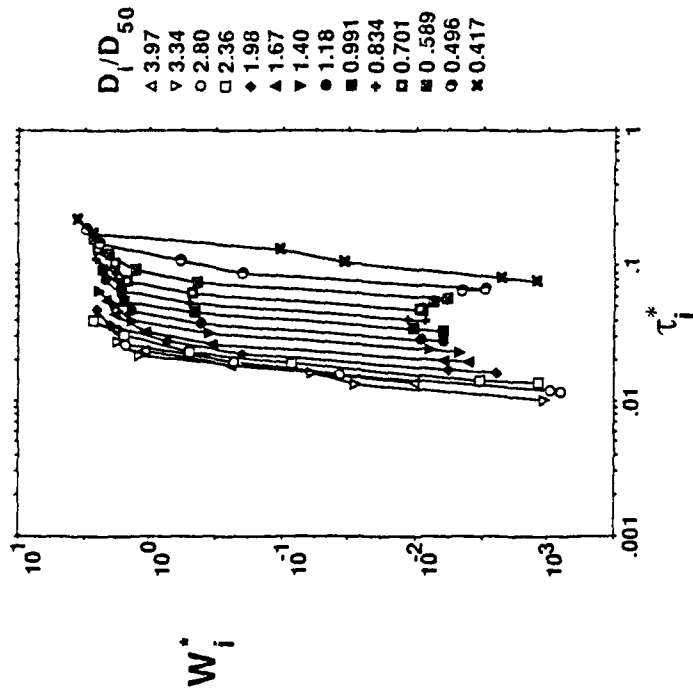Figure 6b: 1φ Fractional Transport Rates
Dimensional Plot

Figure 6a: 1φ Fractional Transport Rates
Dimensionless plot, undistorted axes
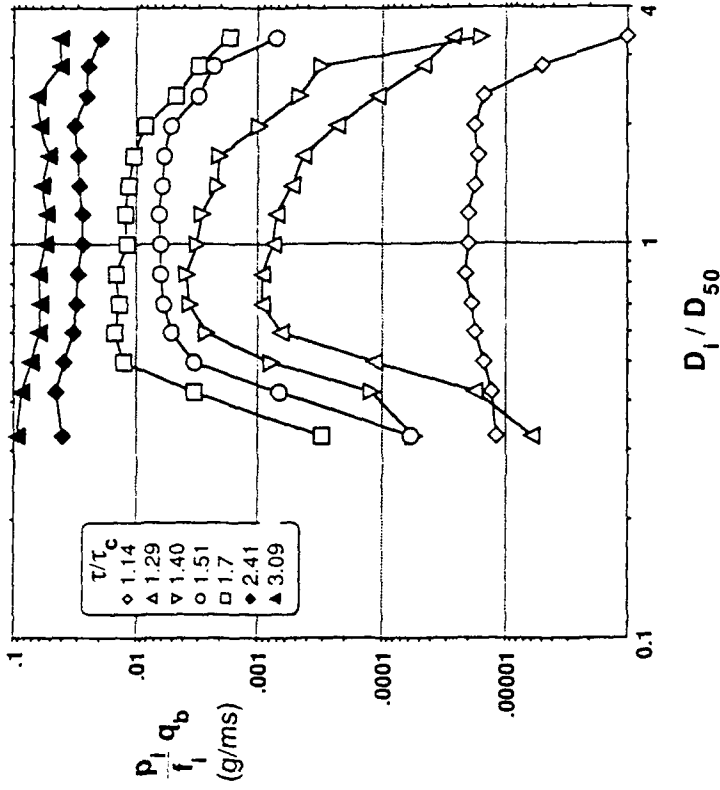
Figure 7. 1/2φ Fractional Transport Rates

$$\frac{p_i}{f_i} q_b$$

(g/ms)

τ/τ_c
◇ 1.01
▲ 0.97
▽ 1.31
○ 1.64
□ 1.98
◆ 2.65
▲ 4.16

D_I / D_{50}
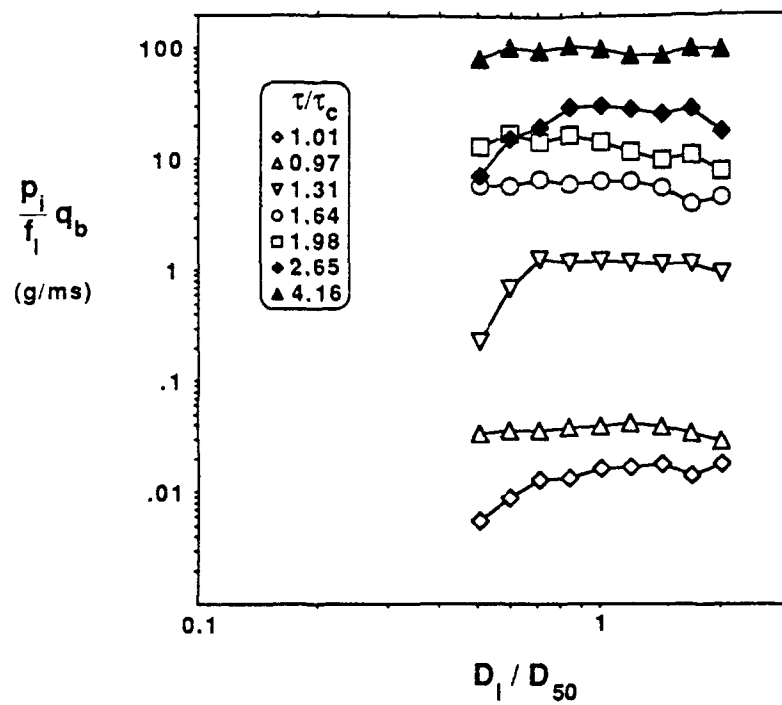
Figure 8. Oak Creek Fractional Transport Rates

$$\frac{p_i}{f_i} q_b$$

(g/ms)

τ/τ_c
◇ 1.07
▲ 1.10
▽ 1.27
○ 1.43

D_I / D_{50}
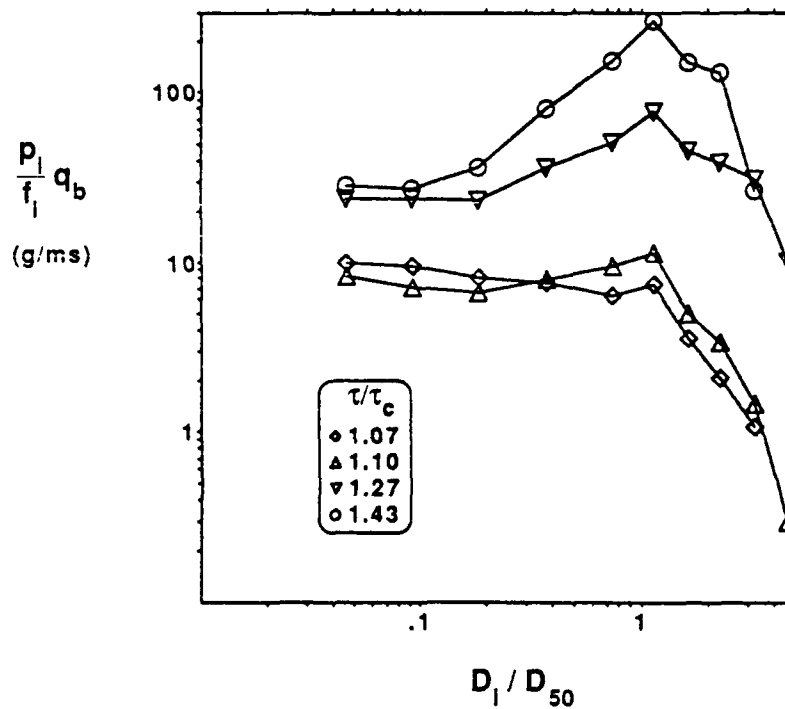
Figure 9: Fractional Transport Rates of the Bimodal Mixtures
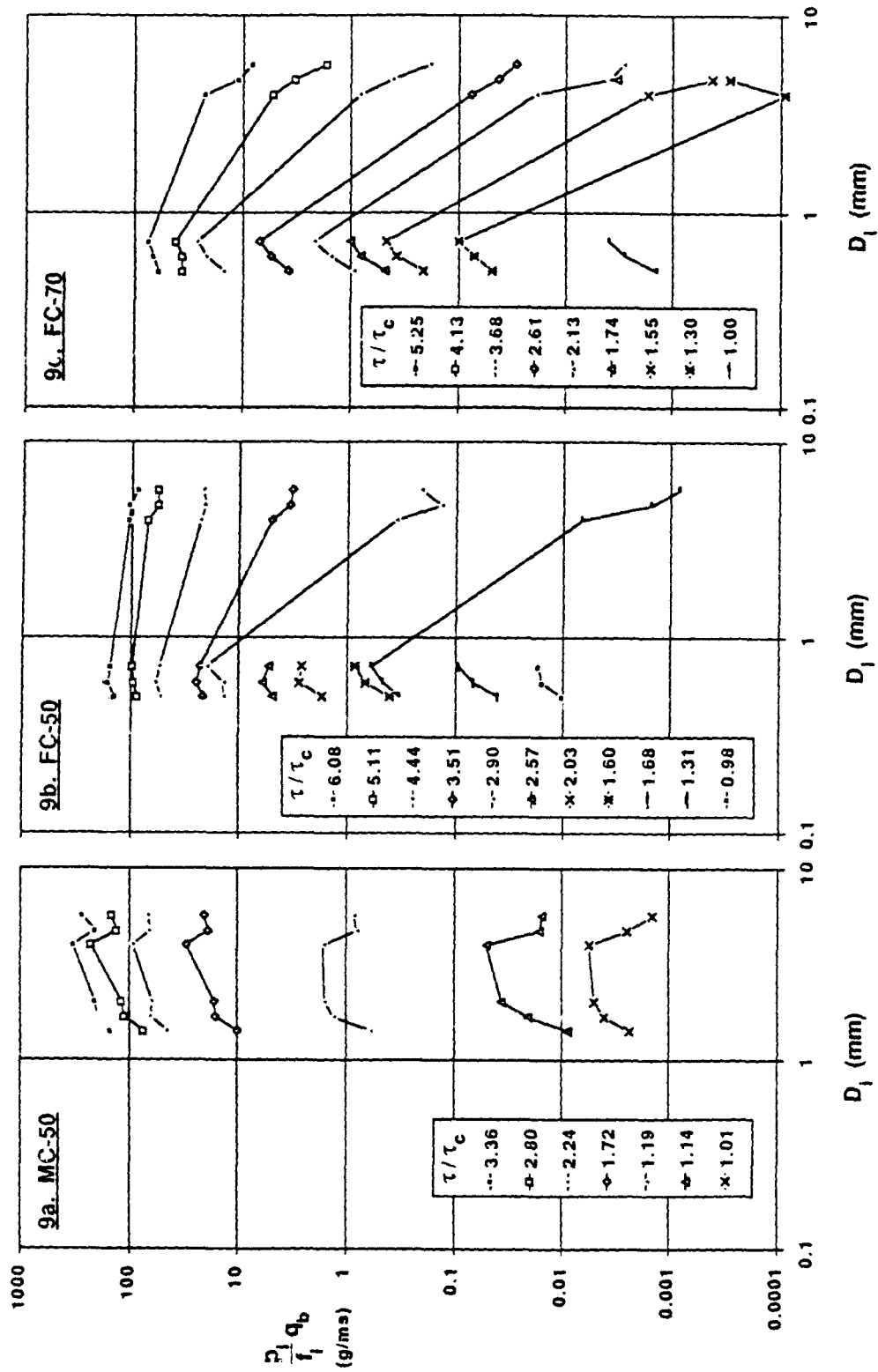
Figure 10. Ratio of bed-form spacing to height in unimodal sediment
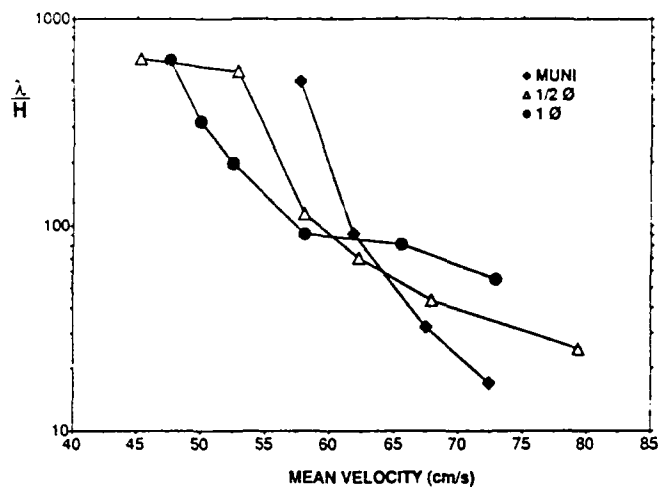
Figure 11. Variation of armor composition with τ/τc for the 1φ sediment
Percentage of fine and coarse sediment on the bed at the end of runs
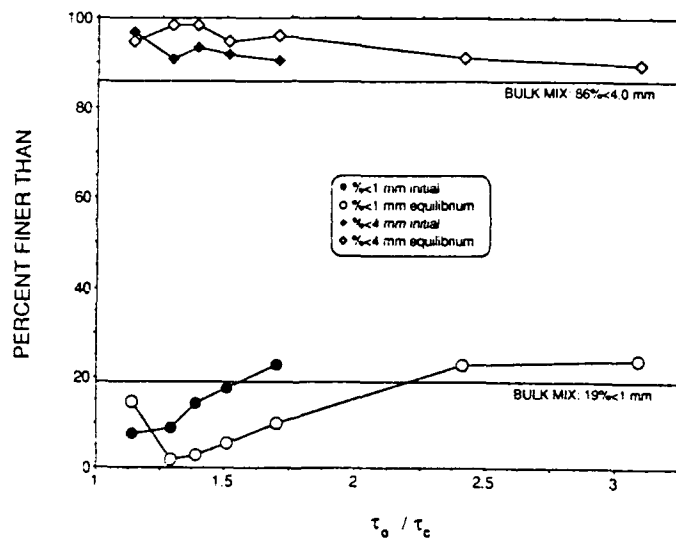
Figure 12. Variation of bed-load composition with τ/τc for the 1φ sediment
Percentage of fine and coarse sediment in transport at the end of runs
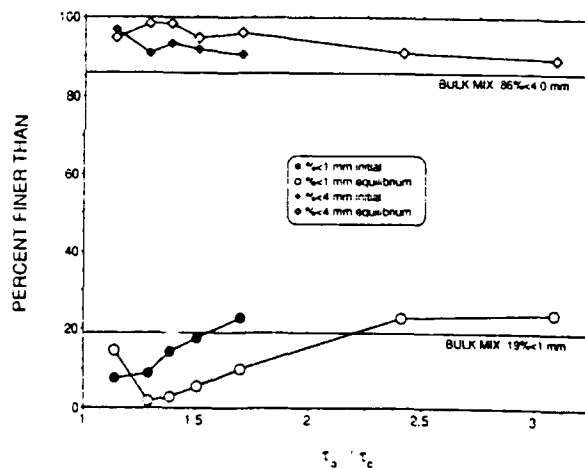
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May 30, 1990

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Dear Dr Billi:

Enclosed is a substitute Page 20 for my Gravel-Bed Rivers Workshop manuscript. The rest of the manuscript should have arrived (along with my registration) by Friday, June 1. Please replace page 20 of the manuscript you have received with the enclosed page. Thank you.

Sincerely,

Peter R. Wilcock
Assistant Professor

Figure 10. Ratio of bed-form spacing to height in unimodal sediment
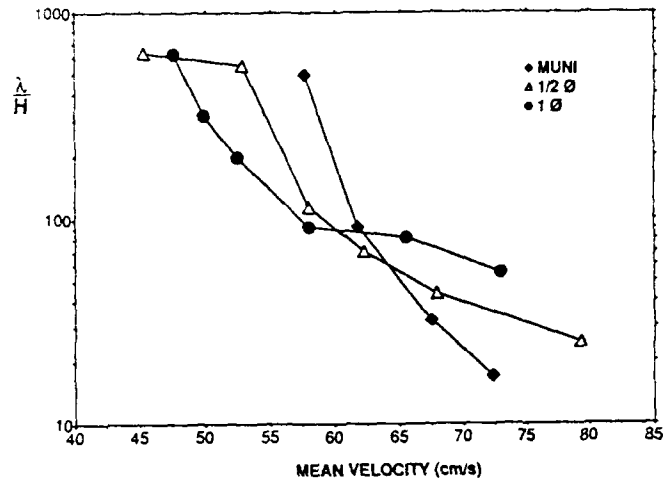
Figure 11. Variation of armor composition with τ/τc for the 1φ sediment
Percentage of fine and coarse sediment on the bed at the end of runs
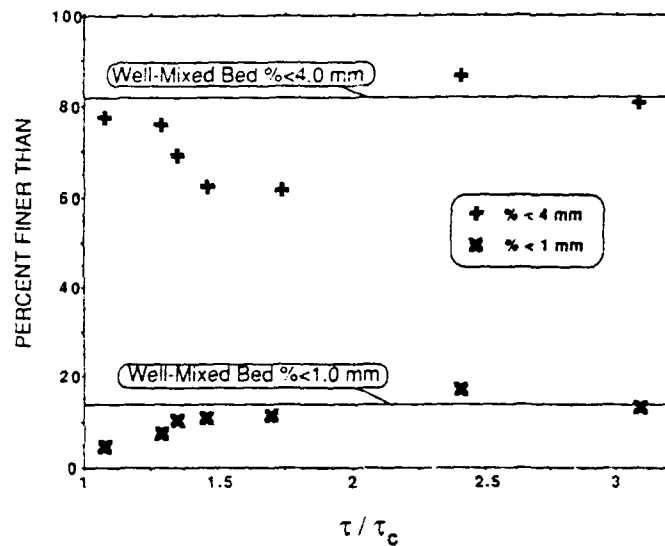
Figure 12. Variation of bed-load composition with τ/τc for the 1φ sediment
Percentage of fine and coarse sediment in transport at the end of runs
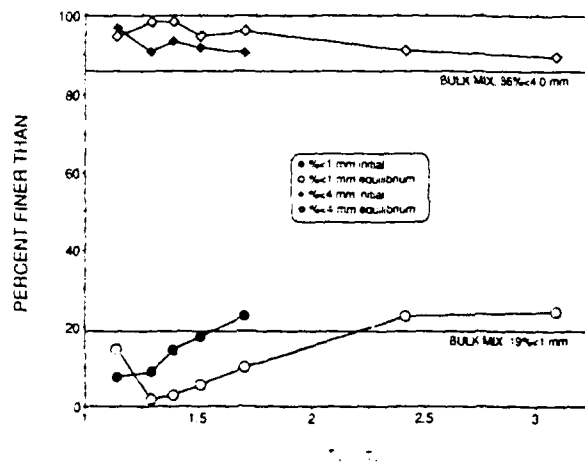
Fractional Transport Rates of Bed Load on Goodwin Creek

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ABSTRACT

Transport rates of eight size fractions of bed material were studied on Goodwin Creek from 1984 - 1988. It was found that D_{50} of the sediment in transport increased in size from 0.77 to 7.77 mm with increasing bed shear stress, the slopes of the transport rates of individual size fractions vs. bed shear stress (W_i^* vs. τ_i^*) decrease with decreasing grain size, and reference bed shear stresses of individual grain sizes were dependent on relative size as well as absolute size (the plot of τ_{ri}^* vs. D_i/D_{50} had a slope of -0.80). Agreement between mean measured transport rates from Goodwin Creek and several published transport relations was poor. A technique for calculating bed load transport separating the sand and gravel fractions is proposed.

INTRODUCTION

The Goodwin Creek Research Watershed is a 21.4 km^2 instrumented watershed that is operated by the USDA National Sedimentation Laboratory in Oxford, Mississippi. The purpose of the watershed is to provide data needed to estimate the impact of upstream land use and watershed processes on sediment supply and transport in stream channels and how this affects channel stability.

The ability to measure or predict the rate of bed load transport of a stream is an important part of the research program at Goodwin Creek. An accurate determination of the rate of bed load in transport is necessary to calculate net erosion from upstream areas as well to assess the effects of the transported sediment on downstream reaches of the channel. Improving our ability to accurately predict bed load transport rates on Goodwin Creek and also on other streams was the reason this study was undertaken.

In most studies of fractional transport rates in gravel-bed streams, peak bed load transport rates have been below 0.3 kg s^{-1} m^{-1} (e.g. Parker et al., 1982; Andrews, 1983; Tacconi and Billi, 1987; Wilcock and Southard, 1989). One exception is the study of Ashworth and Ferguson (1989) in which peak measured rates were up to 3.5 kg s^{-1} m^{-1}. Goodwin Creek, with bed shear stresses in excess of seven times the critical value and maximum measured cross-sectionally averaged bed load transport rates of 3.0 kg s^{-1} m^{-1}, offers another data set from a very active gravel-bed stream.

Most bed load transport relations in the literature characterize the importance of the bed material by one sediment size, usually the median size of the bed material. A few transport relations, most notably that of Einstein (1950), have taken into account the whole sediment mix. Parker, Klingeman, and McLean (1982) concluded that in some gravel bed streams one size fraction may in fact be

sufficient to characterize the transport of the bed material. The transport relation of Parker et al. (1982) was found to not adequately predict bed load transport on Goodwin Creek. The Goodwin Creek data show that a procedure which uses two transport relations, one for the sand and one for the gravel portions of the sediment yielded calculated transport rates close to those measured. The intermediate complexity of this two part relation may yield adequate transport rate predictions in instances where using one grain size has failed and where using data on a larger number of size fractions is not warranted.

FORMULATION OF THE PROBLEM

The concept of equal mobility as pertaining to fractional transport of bed material sediment in gravel-bed streams has been much discussed and debated since it was proposed by Parker et al. (1982). Equal mobility shall be defined here as consisting of two parts: equal entrainment mobility, and equal transport mobility. Equal entrainment mobility is defined as the case when all the sediment sizes in the bed material begin to move at the same flow strength. Equal transport mobility is defined as the case when all sediment sizes are transported according their relative proportion in the bed material (or the size distributions of the bed material and bed load are identical). If one condition of equal mobility for a given channel is fulfilled it does not imply that the other one is. For example Wilcock and Southard (1988) found that near perfect equal entrainment mobility was present in their laboratory flume runs, however, equal transport mobility was approached only at the highest flow strengths used in their flume experiments (Wilcock and Southard, 1989).

The transport of mixed size sediments in an alluvial stream channel is a complicated problem. As soon as movement of the bed sediment begins the size distribution of the bed surface changes as a function of the flow strength. The bed surface size distribution has been shown to be coarsest at low flow strengths and to become finer at higher flow strengths (Kuhnle, 1989). Superimposed on this process is the formation and growth of bed forms which may coexist with the mobile bed pavement for some flows or destroy it at higher flows (Wilcock and Southard, 1989). The flow strengths at which individual sizes of the bed material begin to move is another complex part of this problem. Several studies have found very close approximations to equal entrainment mobility (e.g. Parker et al., 1982; Andrews, 1983, Wilcock and Southard, 1988), while others have found that there is still an effect of relative size at entrainment conditions (e.g. Ashworth and Ferguson, 1989; this study).

A better understanding of the problem may be gained by considering the important variables which control the process:

$$q_{bi} = F(r_0, d, \rho, \mu, \rho_s, g, D_i, D_i/D_{50}) \tag{1}$$

where q_{bi} = the bed load transport rate of one size fraction in mass per time per unit width, r_0 = the boundary shear stress, d = mean flow depth, ρ = the density of the fluid, μ = dynamic viscosity of the fluid, ρ_s = the density of the sediment, g = acceleration of gravity, D_i = the size of a particular fraction of the sediment, D_i/D_{50} = grain size relative to the median grain size (D_{50}), and F symbolizes a functional relationship. These nine variables can be represented by six dimensionless groups:

$$\frac{((\rho_s/\rho)-1)gq_{bi}}{\rho_s \ u_*^3 \ f_i} = F\left[\frac{\tau_0}{(\rho_s-\rho)gD_i}, \frac{d}{D_i}, \frac{g^{1/3}D_i}{\nu^{2/3}}, \frac{\rho_s}{\rho}, \frac{D_i}{D_{50}},\right] \tag{2}$$

where $\nu = \mu/\rho$; $\tau_0 = \rho gRS_0$; $u_* = (\tau_0/\rho)^{1/2}$; $S_0 =$ slope of energy grade line; $R =$ hydraulic radius; $f_i =$ fraction of the ith size contained in the bed material. Equation 2 can be referred to in short-hand notation

$$W_i^* = F(\tau_i^* , d_i^*, D_i^* , \rho_s/\rho, \ D_i/D_{50}) \tag{3}$$

If equal mobility is assumed, the variables W_i^* and τ_i^* may be replaced by W^* and τ_{50}^*, which use q_b and D_{50}, respectively, and represent the entire sediment mix. The variable D_i/D_{50} may be dropped and the variables D_i^* and d_i^* may be replaced by D_{50}^* and d_{50}^*.

For some gravel-bed streams the assumption of equal mobility is a good approximation of reality, however, for Goodwin Creek data the hypothesis of equal mobility breaks down at low flow strengths when not all of the bed material grain sizes are in motion. This is probably because the bed sediment has a significant mode in the sand size range. For gravel-bed streams with sand above some critical percent in the bed material, transport models may be expected to under-predict for low flow strengths. Possibly a better way to treat a gravel-bed stream with a bimodal distribution in the bed material may be to treat the sand and gravel portions as two separate entities each with their own characteristic grain size that interact with each other in different ways depending on the relative proportions present in the bed material and on the strength of the transporting flow. Within the sand and gravel fractions all sizes are assumed to be equally mobile.

It is the hypothesis of this report that when there is a significant sand mode in a gravel bed stream, then the transport rate can be adequately represented by using a two part transport relation: one for the sand material and one for the gravel material. These two transport functions each have a weighting coefficient:

$$w_s = F(f_s) \tag{4}$$

$$w_g = F(f_g) \tag{5}$$

where f_s and f_g are the fraction of sand and gravel in the bed material, respectively, and

$$q_b = w_s(TRS) + w_g(TRG) \tag{6}$$

with TRS - a transport relation for sand and TRG - a transport relation for gravel. For the limited range of conditions considered here it will be assumed that

$$TRS = F(\tau_{50s}^*) \tag{7}$$

$$TRG = F(r^*_{50g})$$ (8)

with the subscripts '50s' and '50g' standing for the median sizes of the sand and gravel portions of the bed material, respectively.

STUDY AREA

The Goodwin Creek Research Watershed is located in the bluff-hills region of north central Mississippi, a region that is characterized by relatively steep slopes and a wide variety of erosion and sedimentation problems. The watershed includes fourteen gauging stations in the form of concrete "V" shaped supercritical flow flumes. These structures serve dual functions as grade control and gauging structures but were designed primarily to facilitate measurements of flow rates and sediment transport rates. Detailed descriptions of the watershed, its instrumentation, and the sediment sampling programs are contained in Bowie and Sansom (1986), Willis et al. (1986), and Kuhnle et al. (1988).

The channels on the watershed are ephemeral with only low base flows occurring most of the time. Flow in the channels sufficient to move the coarsest bed sediment has occurred an average of 12.1 times per year (1985-1988), usually during runoff events caused by the intense rainfall of the convective storms common to this region. As a result of the flow deriving from convective storms flow hydrographs tend to be steep with only rare periods of steady flow in the channels.
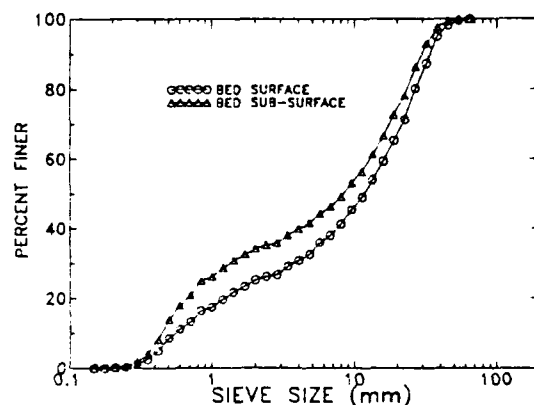
Fig. 1. Grain size distribution for bed material sediment of Goodwin Creek, station 2. Each curve is an average of 10 samples.

All samples considered here were collected at station no. 2 of the watershed. The drainage area upstream of station 2 is 17.9 km². The channel at station 2 is about 25 m wide and 3.0 m deep. The surface layer (sampled at the thickness of the largest grains, 64 mm) of the bed material at station 2 has a $D_{50} = 11.7$ mm and the subsurface material has a $D_{50} = 8.3$ mm (Fig. 1). Both bed surface and sub-surface size distributions have prominent modes at 26.9 and 0.5 mm. The mean bed slope of the reach upstream of station 2 is 0.0033.

SAMPLING EQUIPMENT

The supercritical flow measuring structure at station 2 is a "V" bottom concrete
structure with 5:1 side slopes for 4.7 m each side of the center and 2:1 slopes
thereafter. The structure has a 4 percent longitudinal slope which gives
supercritical flow for all expected discharges and establishes fixed
relationships between depth of flow and flow discharge rate within the
supercritical section. The high flow velocities inhibit sediment deposition and
provide a bottom reference for sediment transport measurements. Bed load has
been sampled at station 2 using modified Helley-Smith (MHS) samplers (Helley and
Smith, 1971; Druffel et al., 1976) suspended from foot bridges at the upstream
end of the supercritical flow flume.

The entrance nozzle of the MHS samplers has been redesigned from a square 7.62
cm orifice to a trapezoidal shape to rest firmly on the sloping floor surface of
the supercritical flume without any void area beneath the orifice. The area of
the inlet orifice is 58.06 cm^2, which is the same as the original Helley-Smith
sampler. The ratio between the outlet and inlet orifices of the MHS samplers is
3.54. The MHS sampler is attached to a rigid strut suspension on a foot bridge
over the upstream end of the concrete flume. This sampling arrangement
eliminates the problems of sampler location and of sampling on uneven bed
surfaces. A quick-release clasp on the 0.25 mm mesh sampler bag allows samples
to be collected with a maximum frequency of about one every 2 minutes.

The hydraulic efficiency of the MHS sampler has been investigated in a laboratory
flume at the National Sedimentation Laboratory. The channel used was 0.71 m
wide, flow depths were all about 0.4 m, and mean flow velocities ranged from 0.4
to 1.2 m s^{-1}. The hydraulic efficiency of the MHS sampler was evaluated by first
taking velocity profiles 2.5 cm inside the inlet orifice of the sampler and at
the same vertical and horizontal locations 0.61 m upstream of the inlet orifice.
The velocities were measured with a pitot tube with a horizontal length of 0.46
m that was inserted through a slit in the top of the collection bag of the
sampler. The pitot tube was connected to a differential pressure transducer that
was sampled by a portable computer. Flow visualization using dye injection into
the inlet orifice of the sampler demonstrated that the slit in the bag had a
negligible effect on flow through the bag. Velocities were taken at five
vertical locations in the sampler as close as possible to the two end walls and
at three equally spaced locations at the interior of the orifice. Five
velocities were taken at each vertical, equally spaced from the highest and
lowest position within the sampler orifice. The velocities at each point were
calculated from averages of 60 samples spaced 1 sec apart of the voltage output
of the pressure transducer. Discharges through the inlet orifice and through an
equal volume in the free stream were calculated for the area around each velocity
measurement. Velocities that were adjacent to the walls of the sampler were
weighted by the zero velocity assumed at the wall.

Hydraulic efficiencies were very close to 1.4 x that of the free stream for mean
flow velocities from 1.22 to 0.85 m s^{-1}. For mean flow velocities of 0.64 and
0.44 m s^{-1} the hydraulic efficiency was 1.2 x the free stream (Table 1). This
is somewhat less than the 1.5 x reported by Druffel et al. (1976) for the
standard model of the Helley-Smith sampler. No attempt was made to measure bed
load sampling efficiencies of the MHS sampler. Hubbell et al. (1985) calculated

that the standard version of the Helley-Smith sampler over-sampled by a factor of about 1.5. This is presumably due to the fact that the hydraulic efficiency is also 1.5 and the width of sediment sampled would have been greater than the width of the sampler. Some scouring of the bed may also have occurred at the sampler location due to its high hydraulic efficiency (Pitlick, 1988). Scouring of the bed with the MHS samplers could not have occurred due to the concrete flume base that was used in this study as a sampling platform. Furthermore the sediment moving through the supercritical flume is moving at a high velocity on the smooth bottom and much of it may have had enough inertia such that it would not be expected to be drawn into the MHS sampler from a width greater than that of the sampler. In any case a bed load sampling efficiency ratio of one was assumed for this study.

Table 1. Summary of Hydraulic Efficiency Tests of MHS Sampler.

Flow discharge (m^3 s^{-1})	Flow depth (m)	Mean channel flow velocity (m s^{-1})	Mean flow vel. through orifice (m s^{-1})	Mean flow vel. free stream (m s^{-1})	Hydraulic efficiency ratio
0.347	0.398	1.23	1.45	1.04	1.39
0.291	0.392	1.05	1.20	0.88	1.37
0.234	0.391	0.85	0.97	0.70	1.39
0.177	0.393	0.64	0.68	0.57	1.20
0.121	0.393	0.43	0.48	0.40	1.20

SAMPLING TECHNIQUES

Sampling with the MHS samplers consisted of lowering the sampler to the flume bottom for a measured time interval such that the sampler fills approximately one-third full. For the period 1985 - 1986 samples were collected only at the centerline of the supercritical flow flume. Beginning in 1987 lateral samples across the right side of the 1:5 sloping section of the flume were taken at each flow strength. Each set of lateral samples was collected starting at the flume centerline and at 1.5 m lateral intervals ending either at the water's edge or the first sample with negligible sediment. These lateral samples were taken sequentially as close together in time as possible. Each set of lateral and center samples was analyzed together to give one mean cross section transport rate. Lateral samples were only collected on the right side of the structure. It was assumed that the distribution of sediment on the other side of the structure was the same. Each sample was stored in a labeled sample can for later drying, weighing, and sieving at the laboratory.

DATA TRANSFORMATION

The MHS sampling was conducted from November of 1984 to September of 1988. From that period 488 samples were analyzed from 21 separate transport events. A total of 187 sets of lateral samples were collected from 12 of these events with 301 samples collected only at the centerline for the other 9 events. Mean cross section transport rates and mean cross section size distributions were calculated for each of the lateral sets of samples. This was done by summing the area under

the sampled lateral transport rates connected by straight lines (Fig. 2) and computing a weighted average for the sediment size distribution. The lateral samples were divided into 3 groups based on depth of flow in the supercritical flume (0.3-0.6 m, 0.6-0.9 m, > 0.9 m) and least-squares linear relations between the mean cross section rate per unit width and the sampled rate at the center of the flume were calculated for each group. The data and least squares line are shown for flume depths > 0.9 m in Figure 3. In order to reduce the effect of outlying points on the least squares equation, points that were more than 2 standard deviations away from the fitted line were dropped and the regressions were repeated. A total of four points were dropped from the regressions by this procedure. The linear relations between mean cross section rate per unit width and sampled rate at the flume centerline of the flume (Table 2) were then used to convert the 301 centerline samples to mean cross section rates per unit width.
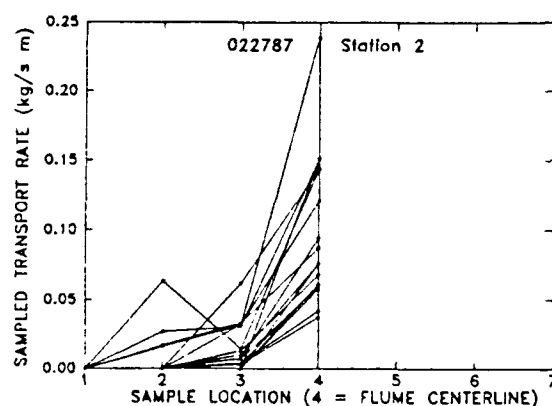
Fig. 2. Plots of lateral sample sets from 2/27/87 transport event. The distance between each sample location is 1.5 m. Lateral samples were taken only on one side of the flume centerline.
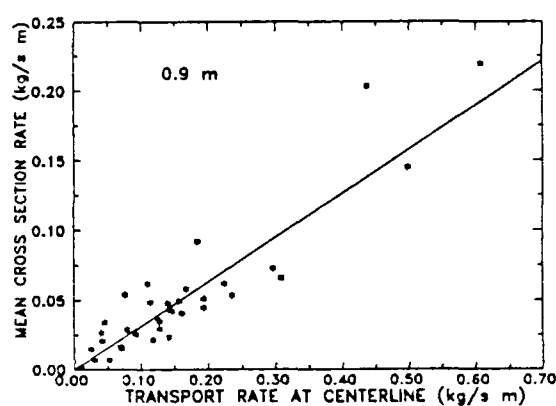
Fig. 3. Example of data used to calculate linear relations between sampled rate at flume centerline and mean cross section rate. The data for flume depths > 0.9 m is shown.

Table 2. Summary of Linear Equations Used to Convert Centerline Samples

Flume depth range (m)	# points	#points removed from regression	slope	R squared
0.3 - 0.6	85	0	0.3050	0.90
0.6 - 0.9	40	2	0.2628	0.76
⟩ 0.9	35	2	0.3162	0.85

The bed shear stress (r_0) for each mean cross section transport rate was calculated at the mid-point of the sample collection times using a mean hydraulic radius vs. flow depth relation from 10 surveyed channel cross sections and water

surface slope data from a 106 m long straight reach 63 m upstream from the supercritical flow flume. Water surface slopes were calculated for each sample time using the data from 4 USGS bubble gauges connected to pressure transducers in a 91 m length of the 106 m long surveyed channel reach. Data from the four bubble gauges was collected every minute by the remote telemetry system of the watershed.

DATA ANALYSIS

Transport Samples

Bed load transport rates calculated from the MHS data taken at station 2 are shown in Fig. 4. A mean transport rate curve for the data was defined by averaging all of the samples in 5 N m^{-2} wide strips from 10 to 40 N m^{-2}. The data was averaged to eliminate the natural variability of the bed load transport processes for gravel (Hubbell, 1987; Kuhnle and Southard, 1988; Kuhnle et al., 1989; Whiting et al. 1988) and to define the mean transport rate for a given flow strength. For flow strengths up to 40 N m^{-2} the number of transport samples collected is assumed to be sufficient to obtain representative average values for transport samples taken in that range of flow strength. Above 40 N m^{-2} not enough samples have been collected to define an accurate mean value and thus this data was not used in this study.

Weighted means of the median size of sediment in transport were also calculated for 5 N m^{-2} wide strips of bed shear stress (Fig. 5). The median size of each sample was weighted by the mass of the transport sample. For a given bed shear stress the median sizes of individual transport samples ranged from a fraction of a mm to nearly 20 mm (Kuhnle et al., 1989). The D_{50} values ranged from 0.84 to 7.77 mm.
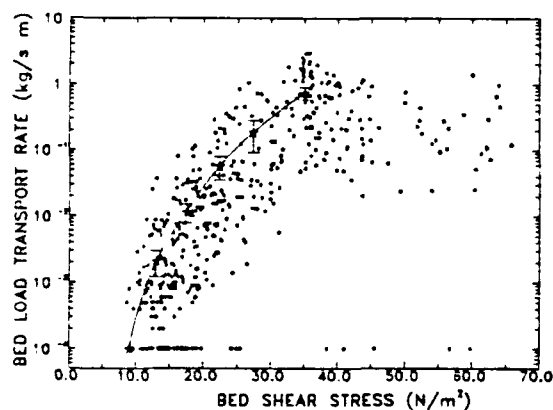
Fig. 4. Plot of 488 bed load samples from #2. Star points are average values from all data in each 5 N m^{-2} wide interval of bed shear stress. Error bars are the 95% confidence interval for the population mean. Solid line is a cubic spline equation fit between the points. Zero values plotted as 10^{-4}.
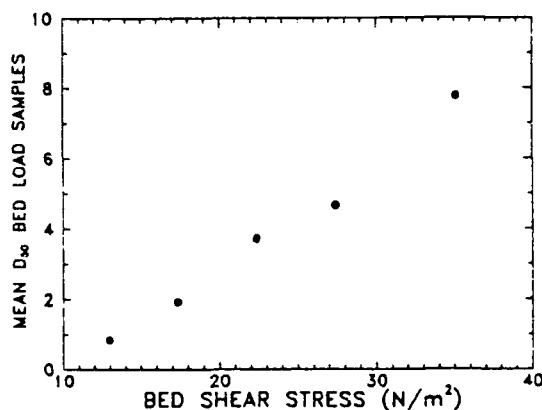
Fig. 5. Mean weighted D_{50} of bed load samples versus bed shear stress.

Fractional Transport Rates

It was necessary first to determine how the size distributions of the 301 centerline samples related to the mean cross section size distributions. When the weighted average size distribution for each set of lateral samples was calculated and compared to the size distribution of the corresponding centerline sample, it was found that the centerline and weighted average size distributions were nearly the same. Therefore the size distributions of the 301 centerline samples were taken as being equal to the mean cross section size distributions of the bed load. The transport rate of each of the eight size fractions was then calculated for the 488 transport samples. These samples were then averaged in increments of 5 N m^{-2} of bed shear stress using the same technique used to average the total bed load transport rate samples.

Eight size fractions of the bed load MHS transport data from station 2 were considered. Figures 6 and 7 show plots of W_i^* vs. τ_i^* and τ_{ri}^* vs. D_i/D_{50}, respectively. The subscript 'i' denotes that the mean diameter of that particular size fraction has been used as the characteristic grain size in that dimensionless parameter. The slopes of the lines in Figure 6 show a progressive decrease with decreasing size (Table 3). The dimensionless reference bed shear stress was defined as $\tau_{ri}^* = \tau_i^*$ when $W_i^* = 0.002$ (Parker et al., 1982). The reference bed shear stress can be taken as a surrogate for critical bed shear stress. The slope of the line in Figure 7 is -0.8 which is somewhat less than the -1 value needed for perfect equal entrainment mobility.
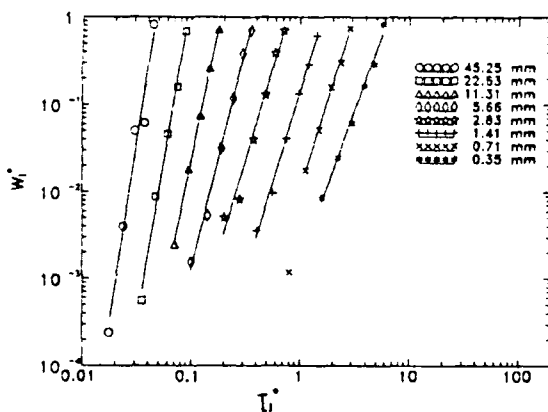
Fig. 6. Transport rates for eight size fractions of Goodwin Creek. One point for the 0.71 mm size fraction was not used in the transport relation.
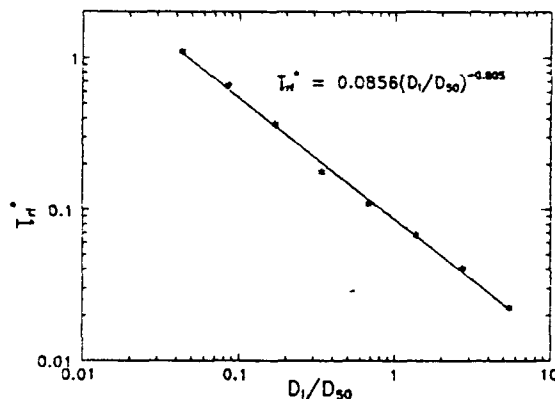
Fig. 7. Dimensionless reference bed shear stress versus relative grain size.

The mean ratios of the fraction of the ith size range in the bed load - p_i, to the fraction of the ith size range in the sub-surface bed material - f_i are plotted against bed shear stress in Figure 8. The relative proportion of the transport rate of each size fraction is shown in Figure 9. Figures 8 and 9 show that the finer sizes are over represented and the coarser sizes are under

represented in the bed load for low bed shear stresses with all p_i/f_i values tending toward one (equal transport mobility) at higher bed shear stresses.

Table 3. Least Squares Fit to $W_i^* = a(\tau_i^*)^b$, and τ_{ri}^*.

geometric mean size of fraction (mm)	b	a	τ_{ri}^*
45.25	8.27	9.36×10^{10}	0.0223
22.63	7.46	4.79×10^7	0.0406
11.31	6.11	2.83×10^4	0.0674
5.66	5.07	1.45×10^2	0.1099
2.83	4.21	2.80×10^0	0.1787
1.41	4.19	0.13×10^0	0.3664
0.71	3.99	1.03×10^{-2}	0.6627
0.35	3.51	1.40×10^{-3}	1.0972
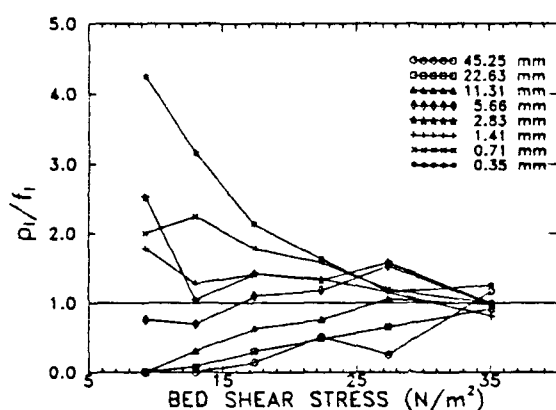
Fig. 8. Ratio of fraction of i-th size sediment in bed load to fraction of i-th size sediment in bed material sub-surface versus flow strength.
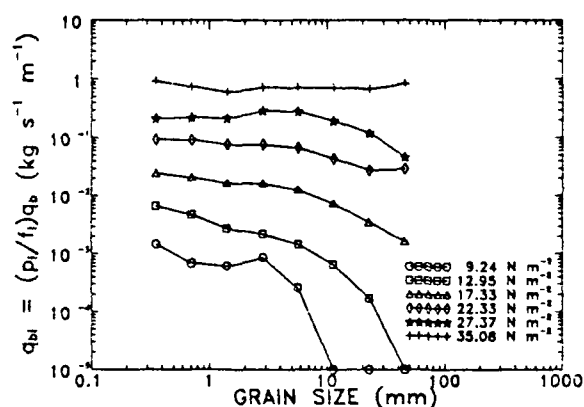
Fig. 9. Fractional transport rates for six flow strengths. Zero values plotted as 10^{-5}.

Prediction of Bed Load Transport Rate

The accurate prediction of gravel bed load transport rates on natural streams is a difficult task (e.g. Parker et al., 1982; Gomez and Church, 1989). The published transport relationships of Meyer-Peter and Müeller (1948), Parker et al. (1982), and Bagnold (1980) were compared to the mean transport relationship of Goodwin Creek graphically shown in Figure 4. None of the three transport relationships used does an adequate job of predicting bed load transport on Goodwin Creek (Fig. 10). The bimodal nature of the bed material at Goodwin Creek may be one of the reasons for the failure of the published transport relations. To test the feasibility of this hypothesis a transport relation which is composed of two parts was formulated for the Goodwin Creek data.

Instead of formulating new transport relations for sand and gravel, two existing transport relations were used with weighting coefficients derived from the Goodwin Creek data. The two transport relations used were Yalin (1963) for sand, and Parker et al. (1982) for gravel. The boundary between sand and gravel was defined as 2 mm.
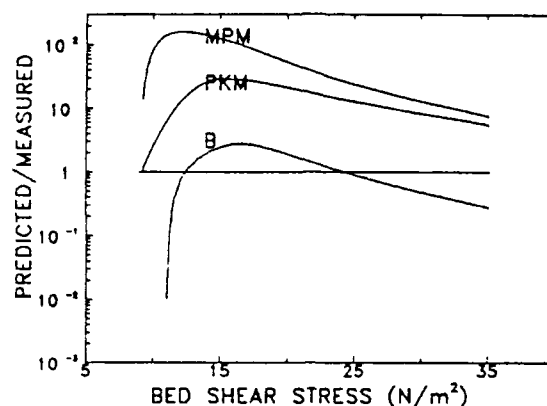
Fig.10. Three transport relations compared to mean Goodwin Creek relation. The horizontal line at 1.0 equals perfect agreement. B - Bagnold (1980), PKM - Parker, Klingeman, McLean (1982), MPM - Meyer-Peter and Müeller (1948).

The calculation of critical bed shear stress values for the two transport equations was based on the bed surface grain size distribution. The critical bed shear stress used for the sand (r_{cs}) was 9.0 N m^{-2}, which was empirically derived from the data. The critical bed shear stress used for the gravel fraction (r_{cg}) of the load was 17.24 N m^{-2}. This value was slightly larger than the reference bed shear stress value calculated for the 45.25 mm size fraction.

A critical part of this two part transport relation is the value of the two weighting factors. After empirically fitting the weighting factors that yielded a best fit with the transport data it was determined that these values were close to the portions of sand and gravel contained in the bed surface material. These values were used (w_s = 0.25, w_g = 0.75) with the two transport relations to calculate the sand and gravel load (Table 4).

Table 4. Bimodal Transport Model Comparison with Goodwin Creek Data

r_0 (N/m^{-2})	net calc. sand (kg/s m)	ratio sand (calc/meas)	net calc. gravel (kg/s m)	ratio gravel (calc/meas)	total calc. sand+gravel (kg/s m)	ratio total (calc/meas)
9.00	0.0000	----	0.0000	----	0.0000	----
12.95	0.0032	1.95	0.0000	0.00	0.0032	1.54
17.33	0.0127	1.84	0.0008	0.16	0.0135	1.16
22.33	0.0293	1.02	0.0302	1.05	0.0595	1.04
27.37	0.0512	0.75	0.2370	2.02	0.2882	1.55
35.08	0.0931	0.38	0.8804	1.79	0.9735	1.32

Discussion

Recent work has shown that the reference shear stress (a surrogate variable for the critical shear stress for movement) of an individual size fraction in a bed with a mixture of sizes is dependent on its relative size as well as its absolute size. Work by Miller and Byrne (1966), Li and Komar (1986), Komar and Li (1986), and Wiberg and Smith (1987) have shown that the interaction of the different sizes in a mix act to reduce or even eliminate the differences in the values of reference bed shear stress of the various sizes. The current understanding of the reduction of the reference shear stresses of the different sizes concerns the different pivot angles of the grains (which are inversely related to relative grain size) and the relative protrusion of the grains into the boundary layer of the flow (which causes a greater force to be exerted on the larger grains for a given flow). When equal entrainment mobility is reached a plot of dimensionless reference bed shear stress vs. relative size (Fig. 7) will have a slope of -1. In fact several researchers have reported data from field streams (Parker et al., 1982; Andrews, 1983; Andrews and Erman, 1986) and from laboratory flumes (Wilcock and Southard, 1988) with slopes very near -1. Ashworth and Ferguson (1989) and the data of White and Day (1982) analyzed by Wilcock and Southard (1988) had slopes of -0.74 and -0.81, respectively. These last two values, which were very close to that calculated for Goodwin Creek, indicate that the larger grains in the sediment mixes still required larger bed shear stresses for movement than the smaller ones. Reasons for this difference may be related to the shape of the size distribution of the bed material sediment.

Deviations from equal transport mobility were also found for bed load transport on Goodwin Creek. The slopes of the transport relations of the size fractions from 0.35 to 2.83 mm are only slightly different from one another, while the slopes of the transport relations for sizes above 2.83 mm show a larger deviation among them (Table 3). The combination of the larger grains needing a greater bed shear stress for entrainment and having a steeper sloping transport rate relation than the smaller grains yields unequal transport mobilities until high bed shear stresses are reached. The pattern of the slope variations of the Goodwin Creek data is very similar to that found by Parker et al. (1982) for Oak Creek. The entrainment mobility of Goodwin Creek, however, is different from that of Oak Creek.

The patterns of transport mobility of the different size fractions for Goodwin Creek data (Fig. 9) are similar to those from laboratory flume experiments of Wilcock and Southard (1989, Fig. 4, p. 1646). The lowest transport rates of both data sets show that the coarse grains are under represented in the transport. For intermediate flows the Goodwin Creek data shows only coarse grains under represented in the bed load not both coarse and fine grains under represented in the bed load shown by the flume data. Both data sets approach perfect equal transport mobility at the highest flows.

The two part transport relation used to calculate bed load transport rate on Goodwin Creek should be taken only as one possible way to deal with gravel-bed streams with bimodal bed material size distributions. The prediction of reference bed shear stresses for the sand and gravel parts of this model is one problem which needs to be studied further. For sand the empirically derived reference bed shear stress could be approximated from the relation: $\tau_{rs0}^* = 0.045$,

with D_{50} of the bed surface used. The reference bed shear stress for gravel could be approximated from the equation: $\tau_{r90}^* = 0.03$, with D_{90} of the bed surface used. In other words the interaction of the two size fractions causes different grain size information to be needed for calculating entrainment and transport rates. These two equations are only for one stream and may not hold for other cases.

Much work also remains to be done in order to determine values of the two weighting coefficients for cases other than Goodwin Creek. The two logical endpoints of this model would be a stream in which there is no sand in the bed material where $w_s = 0$, and $w_g = 1$, and one in which there is no gravel in the bed material where $w_s = 1$, and $w_g = 0$. An experimental program is currently in progress at the Sedimentation Laboratory to test if this technique works for bed material size distributions that cover the range from pure sand to pure gravel.

Conclusions

1. Fractional transport data from Goodwin Creek have shown that deviations from equal entrainment mobility exist on the stream. Bed surface armoring processes have not removed all of the differences in the reference bed shear stresses (τ_{ri}^*) of the size fractions present in the bed material.

2. For low flow rates on Goodwin Creek the coarse fractions of the sediment were under represented in the transport ($p_i/f_i < 1$), the fine fractions were over represented in the transport ($p_i/f_i > 1$), and equal transport mobility is approached as bed shear stresses increase.

3. A two part transport model has been shown to be a possible way to predict transport on gravel-bed streams with significant sand present in the bed material (bimodal size distribution). For Goodwin Creek the weighting factors needed for this model can be taken as the proportion of sand and gravel sediment in the bed surface layer. Within certain ranges the proportions of sand and gravel will likely be important for other streams with bimodal bed material size distributions. Prediction of reference bed shear stresses for this model will probably need to take into account the size distribution present at the surface of the bed of the stream. Dimensionless parameters scaled to D_{50} and D_{90} of the bed surface size distribution were used to arrive at the empirically derived values for reference bed shear stress for the sand and gravel fraction, respectively. It is not known if these parameters will hold for other situations.

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NOTATION

The following symbols are used in this paper:

D_{50} — median grain size;

D_i — mean grain size of the ith grain size subrange;

$D_i^* = g^{1/3} D_i/\nu^{2/3}$;

$D_{50}^* = g^{1/3} D_{50}/\nu^{2/3}$;

d — mean flow depth;

$d_i^* = d/D_i$;

$d_{50}^* = d/D_{50}$;

F = a functional relationship;

f_g = fraction of gravel in bed material;

f_s = fraction of sand in bed material;

f_i = fraction of the ith size sediment contained in the bed material;

g = acceleration of gravity;

p_i = fraction of ith size sediment contained in bed load;

q_{bi} = bed load transport rate of ith size fraction in mass per time per unit width;

q_b = total bed load transport rate in mass per time per unit width;

R = flow cross-sectional area/wetter perimeter = hydraulic radius;

S_0 = slope of the energy grade line;

TRS = bed load transport relation for sand;

TRG = bed load transport relation for gravel;

$u_* = (\tau_0/\rho)$ = bed shear velocity;

$W^* = ((\rho_s/\rho)-1)gq_b/\rho_s\,u_*^3$ = dimensionless total bed load;

$W_i^* = ((\rho_s/\rho)-1)gq_{bi}/\rho_s\,u_*^3\,f_i$ = dimensionless bed load in i^{th} size range;

w_g = weighting coefficient for transport relation for gravel;

w_s = weighting coefficient for transport relation for sand;

μ = dynamic viscosity of fluid;

$\nu = \mu/\rho$;

ρ = density of the fluid;

ρ_s = density of the sediment;

$\tau_0 = \rho gRS_0$ = bed shear stress;

τ_{cs} = critical bed shear stress for sand portion of bed material;

τ_{cg} = critical bed shear stress for gravel portion of bed material;

$\tau_{50}^* = \tau_0/(\rho_s-\rho)gD_{50}$; = dimensionless bed shear stress for median size of bed material

τ_i^* = $\tau_0/(\rho_s-\rho)gD_i$ = dimensionless bed shear stress for i^{th} size range;

τ_{ri}^* = dimensionless reference bed shear stress for i^{th} size range,
(τ_i^* = τ_{ri}^* when W_i^* = 0.002);

τ_{r50}^* = dimensionless reference bed shear stress for D_{50} of bed surface sediment;

τ_{r90}^* = dimensionless reference bed shear stress for D_{90} of bed surface sediment;

τ_{50s}^* = dimensionless bed shear stress using median grain diameter of sand fraction;

τ_{50g}^* = dimensionless bed shear stress using median grain diameter of gravel fraction;

MONITORING OF PARTICLE MOVEMENT ON
VIRGINIO GRAVEL-BED STREAM

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ABSTRACT

A continual operation bedload measurement station was installed near Baccaiano in the Virginio gravel-bed stream, which has a grain-size about D_{50} = -5 φ and a straight shape with alternate bars. The slope of the stream at this point is about 0.008 and the basin has an area of 40 km². The introduction of a marked sample of about 3000 pebbles, upstream the measurement station, makes possible to study the relationships between bedload transport with the pebbles passage through the station itself and furthermore with their position in the bed after each flood.

The results of this study confirm the pulsating rate of the bedload transport and its poor correlation with grain-size distribution and discharge. For each grain-size classes the distance travelled by the pebbles, their embedding and, in some cases, their mean velocity, was measured.

1. INTRODUCTION

The main part of the bed-load transport in a gravel-bed river has a grain-size considerable inferior to that characteristic of its river bed.

The most coarse grain-sizes, constituting as a rule only a small part of the ground bed-load trasport, are an important contributory factor in the river-bed dynamics since they form the base of the channel, and the bars and riffles are generally formed by them. The means and quantity of pebbles transported in a gravel-bed river are little known. Its measurement is very difficult and apart from the little data measured in experimental basins and that gathered in laboratory water-channels, it has been almost impossible to measure on a large scale with a great enough level of accuracy and using present-day methods, in natural river-beds Emmett 1981; Tacconi, 1982).

The results obtained experimentally in natural river-beds between bed-load transport and other parameters such as discharge or hydraulic features of the current, are much lower than expected (Billi and Tacconi, 1987).

The variability of the coarse sediment transport is very high and it is characterized by a pulsating nature linked more to morphological-sedimentary features of the river-bed than to variations in discharge. (Klingeman e Milhous, 1970; Hayward and Sutherland, 1974; Meade et Al., 1981; Reid et Al., 1985; Tacconi and Billi, 1987).

2. GENERAL DESCRIPTION OF RIVER-BED AND EXPERIMENTAL STATION

Virginio Creek, situated in the Arno basin in Tuscany near Baccaiano, was equipped with a measurement gauge in continual operation for bed-load ($\varphi > -1$) supplied with a vortex tube trap and readmission system for sediment in the downstream river-bed. The basin upstream has a surface area of 40 km, in which are present marine pliocene deposits composed of silts, sands, pebbles and used for cultivating grain, vines and olive trees. It has a mediterranean climate with an average annual precipitation of 900 mm.

The river-bed upstream from the measurement station is straight with alternating bars, joined to riffles and pools with bankes covered by thick vegetation. The average size width-wise with a high-water level is 12 m with a gradient of 0.008.

The results of the studies made at the measurement station up until 1985 have published (Tacconi and Billi 1987; Billi and Tacconi 1987). In brief, the studies carried out have shown for Virginio Creek:
- the extreme variability of coarse sediment transport;
- its pulsating nature with a cycle of about 30';
- the low correlation between sediment and liquid transport discharges;
- the different means of transport for high discharges where the river-bed completely mobilizes itself forming a conveyor belt with solid discharge occurring in a single event like the total

discharge measured over several consecutive years;
- the successive and progressive evolution of the bed with the formation of alternating bars and the development of the armour in the low-water channel, the close similarity between grain-size in the bedload transport and the subarmour.

The present note refers to the measurements and to the studies carried out in 1985. In this period there were only four flood events studied and, indeeed along with an exceptional period of drought, the Local Authorities starded the work of dredging the river-bed which they claimed necessary. Inspite of its ineffectiveness, this work transformed the river-bed in a channel of trapezoidal sections destroying its natural shape and rendering for a certain amount of time any significant measurements impossible.

3. METHODOLOGY

The operations possible for measuring sediment transport at the measuring station are the following:
1) Analysis and control of transverse sections in river bed section under study.
2) Grain size measurements of armour and subarmour.
3) Continual measurement of bedload transport during flood events ($\varphi > -1$).
4) Drawing of sample and grain size analysis of bedload transport during flood events.
5) Continual visual analysis of bedload transport with analysis of any further marked pebbles (transit time, grain size, weight followed by immediate reintroduction in down-stream river-bed).
6) Introduction of a marked sample upstream from measuring station in section under control.
7) Direct survey in river-bed of the cloud formed by sample distribution downstream from point of introduction, and both up and downstream from measurement station both before and after the subsequent flood events.
8) Other operations such as the measurement of liquid discharge of suspended sediment transport, velocity, etc..

3.1 Observations and Measurements

The experiment made in 1988 consists of the introduction of a marked sample into a section 250 m upstream from the measurement station and its subsequent analysis: - evolution of rver-bed; - measurement of bedload transport; recording of passage made by clasts in marked sample during the floods from the measurement section; - measurement of the distances travelled by all visible clasts (not those buried in river-bed) in each grain-size class of marked sample after the 4 flood events.

For operational purposes, not all the observations and measurements sediment transport during the events under study continued throughout the entire duration of the event.

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3.2 Features of the River-bed at the time of Experiment

The means adopted, the nature and the quantity of the bedload transport depend to a larger extent on the state of the river-bed before the event than on the liquid discharge (Tacconi, Billi, 1987).

As mentioned in the introduction, the river-bed under study was subjected to maintenance works by the Local Authorites in 1986. The trapezoidal section that was artificially formed after the various flood events causing considerable sediment discharge, was again modified in Spring 1988 and the river-bed returned to its initial natural form (proving the ineffectiveness, if not the damage, of the Autority's dredging); the river-bed developed once more its former straight form with bars alternating with the low flow channel, and an apparent armour and sequences of riffles and pools.

The marked sample was introduced in a section 250 metres upstream from the section for measuring sediment discharge. A lateral bar on the left bank was present here placed almost 30 m from the introduction section. The low water channel was 4 m wide and was recognizable from the rest of the bed through the presence of a lateral step measuring about 10 cm; the grain-size of the river-bed examined corresponding with the lateral bars was the following:

Armour
Mean Diameter (φ) = -5.961852
Standard Deviation (φ) = .6869485
Skewness = .5583912
Kurtosis = 3.413433

D10 (φ) = -5.045822	D10 (mm) = 33.03268
D50 (φ) = -6.024881	D50 (mm) = 65.11334
D90 (φ) = -6.831544	D90 (mm) = 113.8937

Subarmour
Mean Diameter (φ) = -4.929065
Standard Deviation (φ) = 1.287_1
Skewness = 1.393888
Kurtosis = 5.186635

D10 (φ) = -3.276591	D10 (mm) = 9.690636
D50 (φ) = -5.17537	D50 (mm) = 36.13613
D90 (φ) = -6.284161	D90 (mm) = 77.9329

About the low-level bed, the armour grain size was measured just before the observation period.

Mean Diameter (φ) = -6.383727
Standard Deviation (φ) = .6338497
Skewness = .6787533
Kurtosis = 3.149776

D10 (φ) = -5.505165	D10 (mm) = 45.41713
D50 (φ) = -6.480906	D50 (mm) = 89.31964
D90 (φ) = -7.181181	D90 (mm) = 145.1279

The sample was introduced by throwing the clasts at random across the whole low-water channel, 4 meters wide in a cross-wise strip 2 meters wide.

3.3 Features of the marked sample

The preparation of the marked sample was preceded by a preliminary survey that involved sampling channel and bars sediments in the river-bed tract upstream the measuring station.

The marked sample was prepared and called "Grain size", sizing between -4 φ and -7 φ, and composed of 3935 pebbles weighing a total of 337 kg.

Besides this "grain-size", a "morphometric" sample was introduced composed of 300 pebbles sizing between -4.5 φ and -5.5 φ, representative of the three forms in Folk's classification (spheric, discoidal, elongated).

The pebbles were examined in the same river-bed to ensure identical lithological features (limestones) of those normally present in bedload transport and they were subsequently given epossidic colour using two yellow colour components.

The individual clasts of the morphometric sample were numbered and coloured red enabling close study of their movements; the morphometric sample was introduced in order to study the effect of pebble morphometry on mobility; that study is not included here.

TABLE I
CHARACTERISTICS OF MARKED SAMPLE
GRAIN SIZE SAMPLE (YELLOW)

φ	N° Clasts	Weight kg	Mean Weight
-4 ÷ -4.5	2000	20	10
-4.5÷ -5	1000	28	28
-5 ÷ -5.5	500	32	64
-5.5÷ -6	250	50	200
-6 ÷ -6.5	125	80	640
-6.5÷ -7	60	127	2117
-4 ÷ -7	3935	337	85.6

MORPHOMETRIC SAMPLE (RED)

φ	N° Clasts	Weight kg	Morphometry
-4.5÷ -5.5	100 100 100	8.5	Elongated Discoidal Spheric
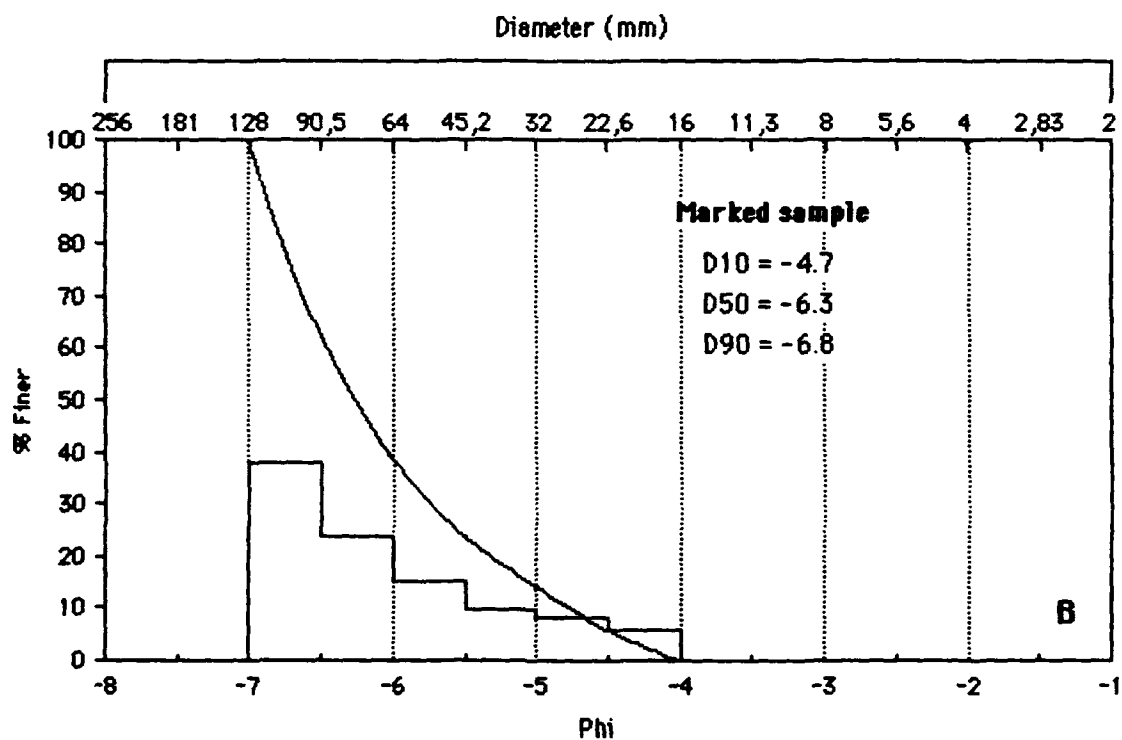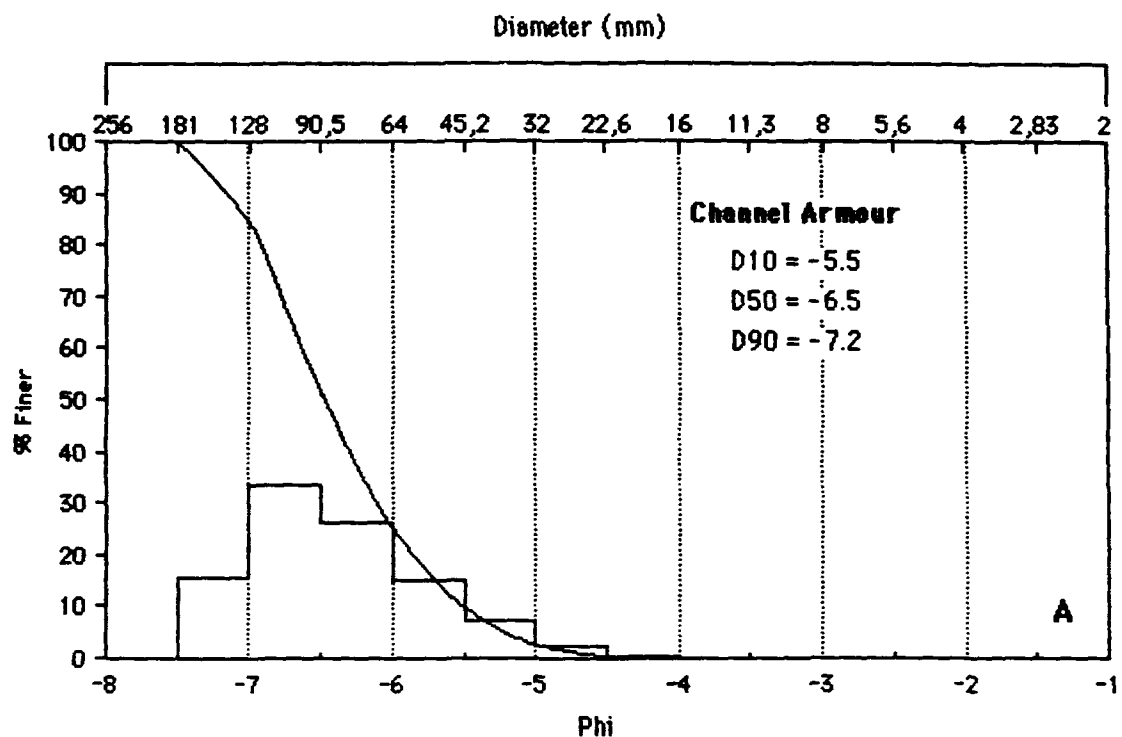
Fig. 1. Granulometric distribution of the armour channel (A)
and of the marked sample (B).
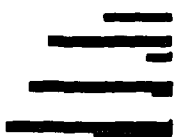
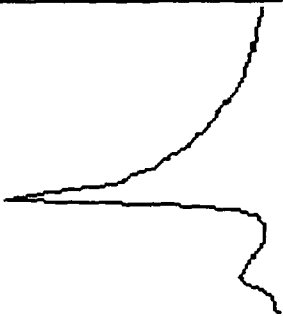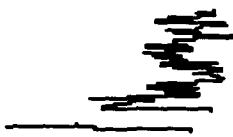| Flood number	1	2	3	4
Date	13/4/88	20/5/88	28/5/88	5/6/88
Rainfall	No measurements			No measurements
Discharge				
Bedload	No measurements			Total Bedload: 450 kg
Peak Discharge (mc/s)	3	17	6	4,9
Max Bedload (kg/min)		580	125	184
Observations		345	331	
Pebbles Mobility — Range		1 - 666	1 - 744	2 - 776
Pebbles Mobility — Mean		234,6	246,4	192,2

Fig. 2

4. OBSERVED FLOOD EVENTS

The sizes measured for each of the four flood events are concisely drawn up in Fig. 2.

The main hydrological features and solid discharge for each flood are subsequently described.

4.1 Flood n° 1 - 14.04.1988

This is a low intensity event with a maximum discharge of about 3 mc/sec; and duration of 1440 minutes. The sediment discharge was not measured.

4.2 Flood n° 2 - 20.05.1988

This is the highest intensity flood of the four observed and it is associated to four main precipitation events.

A first peak of about 4 mc/sec was registered followed by the main peak with a discharge of 17 mc/sec.

The recording of the sediment discharge is not complete; in fact when the measuring station was put into use, some marked pebbles were found that must have travelled along a section of the order of 250 m.

Moreover, 30 minutes after the start of the recordings, due to operational problems connected with the high rate of bedload trasport, measuring was stopped for 50 minutes. A maximum bedload transport of 580 kg/min and 2110 kg in 5 minutes was recorded.

This is the highest level of sediment discharge ever recorded at the measuring station from its installation.

The total sediment discharge measured was about 43260 kg. After this interruption, measuring continued for more than 48 hours during which further data was not drawn. The characteristic events measured are indicated in Fig. 3.

4.3 Flood n° 3 - 28.05.1988

Inspite of its weaker intensity, the recording of data for this flood event comprises all the sizes.

The lenght of the flood wave is 420 minutes, with a peak discharge of 6 mc/sec.

The maximum sediment discharge measured is 125 kg/min while the total sediment discharge is 2780 kg.

A few isolated arrivals of bedload may be noted that have preceded the sediments' real state of continual motion.

During the principle event of sediment discharge following the first isolated episodes, the initial condition of sediment movement is easily recognizable, with a liquid discharge of little more than 2 mc/sec (Fig. 4).

The sediment discharge is exhausted quickly in this event after the peak period of liquid discharges.
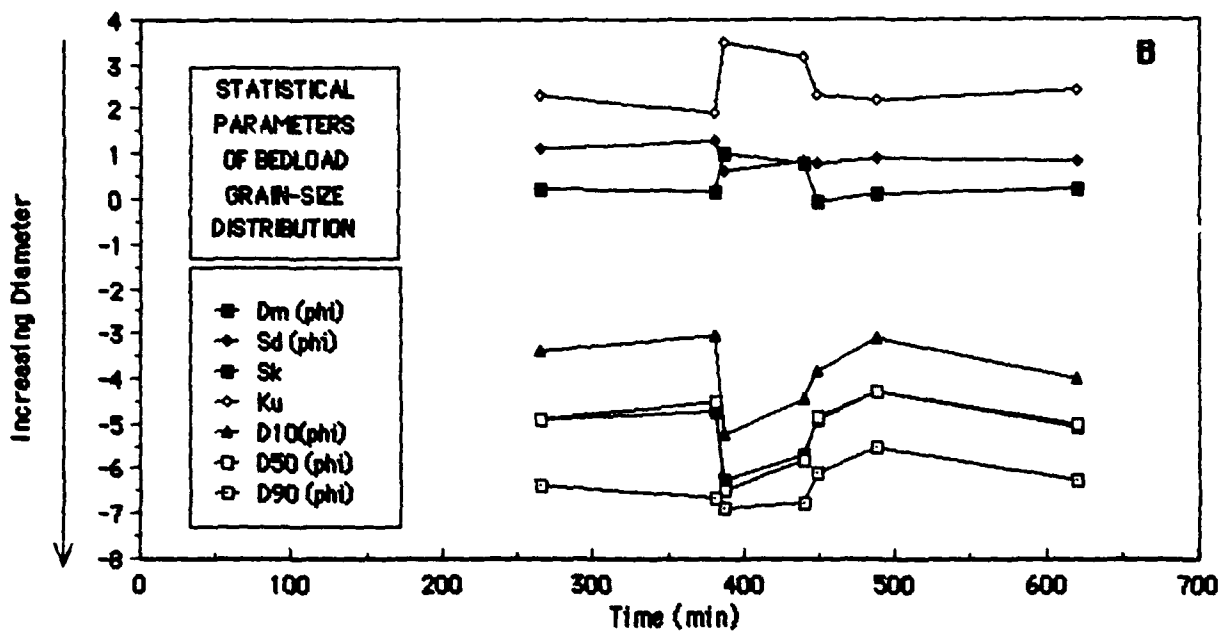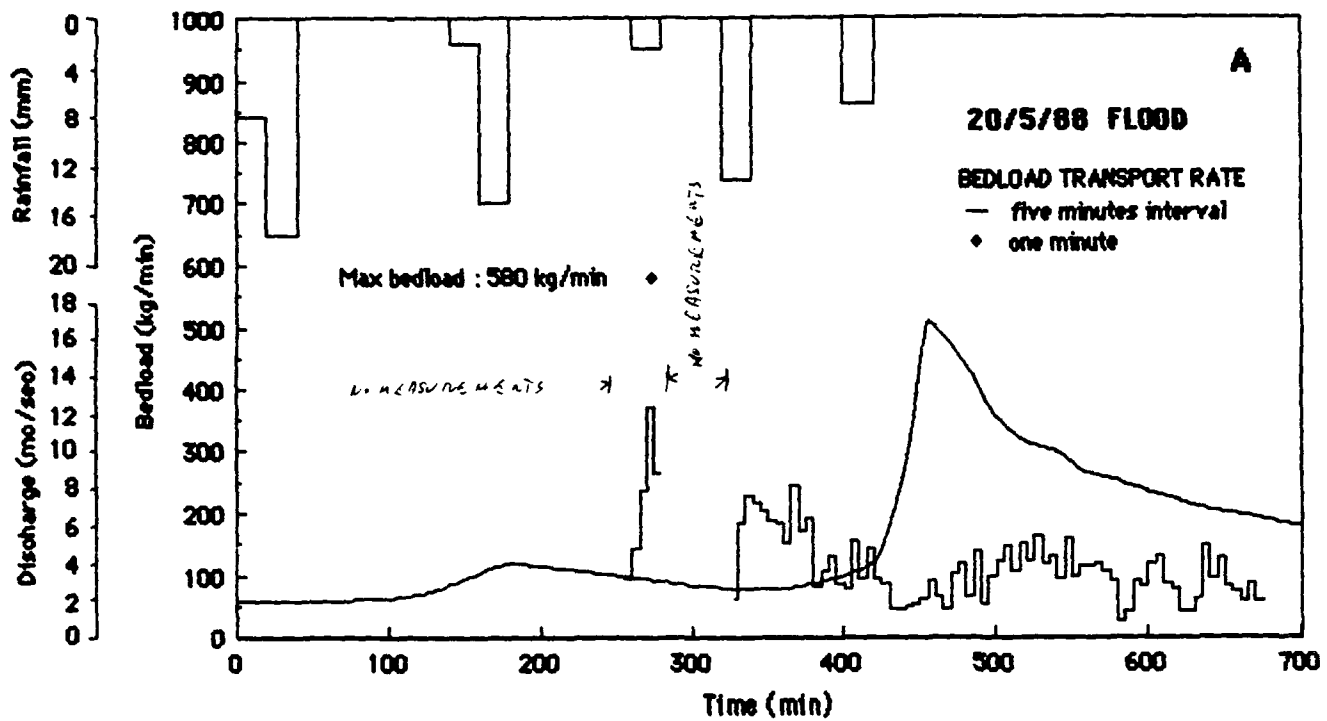
8

Fig.3 . 20/5/88 flood : A) rainfall, discharge and bedload;
B) statistical parameters range for the bedload samples.
Dm: mean diameter; Sd: standard deviation; Sk: Skewness; Ku: Kurtosis;
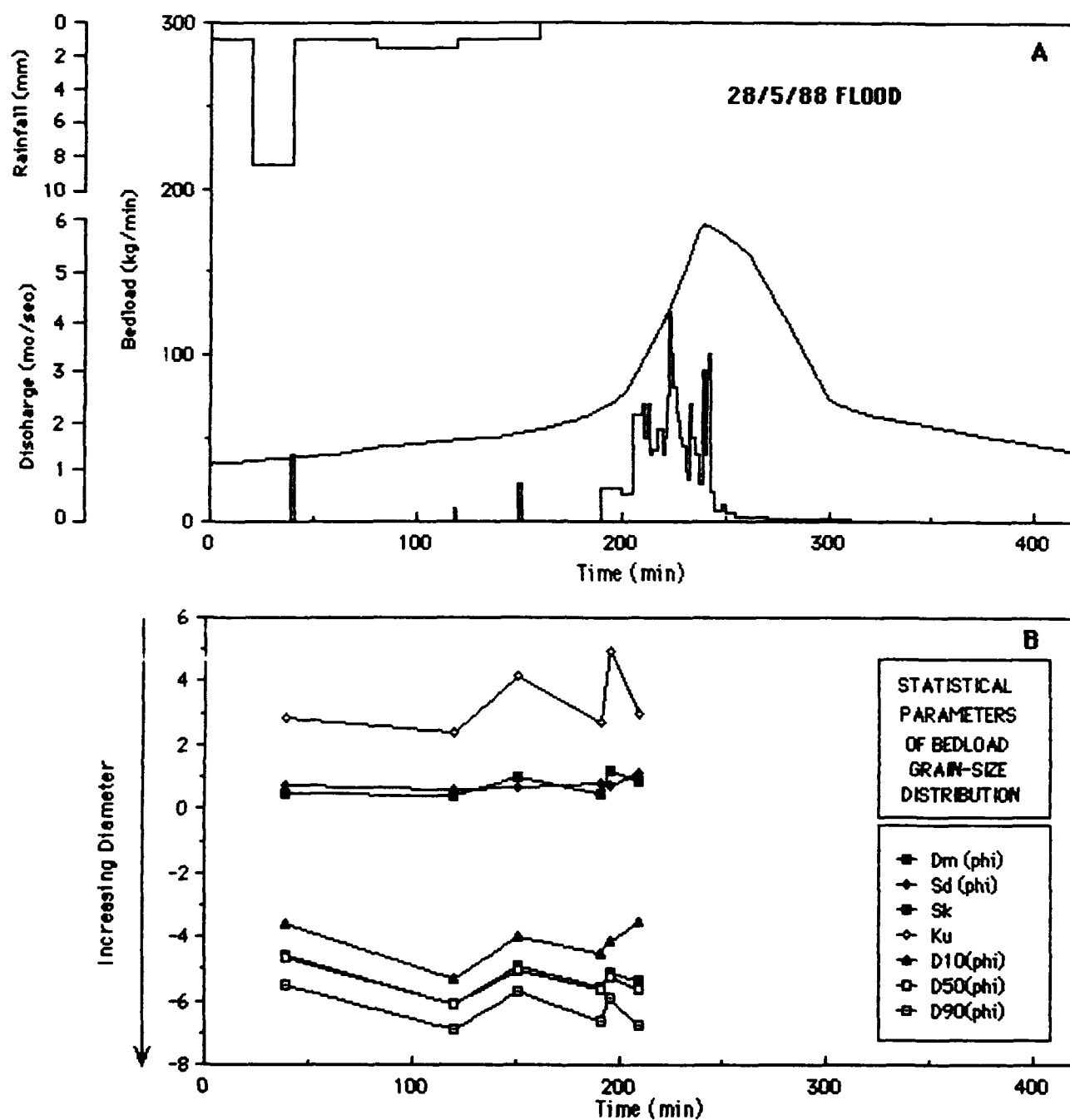D10,50,90: diameters for 10,50,90 percentiles.

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Fig.4 . 28/5/88 flood : A) rainfall, discharge and bedload;
     B) statistical parameters range for the bedload samples.
     Dm: mean diameter; Sd: standard deviation; Sk: Skewness; Ku: Kurtosis;
     D10,50,90: diameters for 10,50,90 percentiles.

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4.4 Flccd n° 4 - 05.06.1988

This event lasts a considerable amount of time, with a peak discharge of 4.9 mc/sec. Although data measuring the sediment discharge continuously is not available, it was observed that at the end of the event the vortex-tube-trap contained 450 kg of material instead of its maximum capacity of 700 kg. The material trapped inside may be put down to the total bedload transport.

This material was moreover composed of fine sediments and "mud balls", linked to a few pebbles of medium-sized dimension. This data leads one to think of an event characterized by a prevailing suspended sediment transport, linked to a prolonged washing away from its sides. The modest amount of bedload may be explained by a higher level of armouring of the river-bed that remained stable with relatively high discharges.

One supposing is also connected to the particular shape of the hydrogramme that does not present abrupt variations in the discharge even in the growth stage.

Moreover it could be a bedload due to lateral armour of material deposited and remaining stable on the armour. This can be supported by mean of the marked pebbles analysis; in fact the number of marked pebbles after flood n° 4 is much less then before, this can be due to the pebbles burying the previous ones.

5. MOVEMENT OF THE MARKED PEBBLES

As we have already noted, both the "grain-size" and the "morphometric" part of the marked sample was introduced into the river-bed by randomly throwing into a section 250 m upstream from the measuring station.

In order not to create obstacles in the current, the sample was distributed over a strip 1 m wide. The marked clasts together with the riverbed clasts were moved about during the following floods, remaining partly buried under the river-bed and partly visible on the surface, embedded amongst the other clasts or leaning over and free to move about.

The visible clasts on the river-bed and partly visible on the surface, embedded were examined after the various floods determining their dimension and the distance they had travelled.

During the periods in which the bedload was measured, the clasts were recorded on their arrival and then thrown immediately back into the river-bed downstream from the measuring station; during the periods in which the measuring station was not in operation, the same clasts together with the bedload passed freely on from the measuring station.

This situation enabled us to follow "the cloud" of clasts even in the section downstream from the measuring station and determine its characteristics taking into account the maximum distance it had travelled down. The position of the clasts relative to each individual flood is demonstrated in Fig. 5 (Enclosed in text).

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5.1 "Grain-size Sample"

It is noted how the flood n° 1 brought about only a slight redistribution of the marked sample with a maximum shifting of about 18 m from its introduction point (Fig. 5).

This small flood did not determine at all the bedload discharge or determined only a very small amount. Although measurements are not available for this bedload, this flood is significant since it moves the marked sample about placing the clasts in a position that is closer to their natural condition; in other words, the flood already places the sample in the optimal situation for determining successive measurements that are not distorted by its very presence.

The flood n° 2 caused the greatest movement of marked pebbles as a result of the greater amount of sediment discharge; the marked clasts were discovered on the surface up to 666 m away the introduction point.

Following the floods n° 3 and n° 4 the maximum distance in which the marked pebbles were found was 744 m and 776 m respectively. Whilst the barycentre of the pebble cloud resulted in being a certain distance away from the introduction point: 235 m after the 2nd flood; 246 m after the 3rd flood and 192 after the 4'.h flood.

The total number of pebbles found after each flood is the following:

2nd flood n° 345;   3rd flood n° 331;   4th flood n° 184

The river-bed survey continued for another 1000 m downstream from the last pebble found in the river-bed, consequently the part of the sample that was not raised must have been buried in the bed of that channel section.

5.2 "Morphometric Sample"

As we mentioned with reference to the "Grain-size" sample, a sample composed of 300 clasts with φ composed between $-4.5 \div -5.5$ coloured red and individually number was introduced. The number of findings in the river-bed of such clasts is as follows:

2nd flood n° 13;   3rd flood n° 15;   4th flood n° 5

Having being numbered individually, it was possible to follow the movements of each single clast with greater precision away from the introduction point, and the distances between the 2nd and 3rd and those between the 3rd and 4th makes it possible to distinguish which pebbles out of the 300 present had remained on the surface and which ones embedded. The mean distance travelled by the clasts in "morphometric" sample is:

flood n°2 = 247 m;   flood n° 3 = 292 m;   flood n° 4 = 252 m

The results are printed in Table 2.

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Table 2 - Passage of Clasts in the "Morphometric" Sample

1	2	3	4	5	6	7
42	54.5	74		19		
100	88	202		114		
54	105	298		193		
75	142	151	227	9	76	0.608
6	202	300		98		
38	52	57	57	5	0	0
19	71	77		6		
69	312	421		109		
52	375	632		259		
73	195	195		0		
47	280	291		11		
61	477	495		18		
3	491	559	576	68	17	0.136
56		184	232		48	0.384
34		208	233		25	0.200

1) Number of Marked Pebble
2) Position after the flood of 20.05.88
3) Position after the flood of 28.05.88
4) Position after the flood of 05.06.88
5) Passage taken in the flood of 20.05.88
6) Passage taken in the flood od 28.05.88
7) Mean velocity (m/min) of marked pebbles during the flood of 28.05.88

6. OBSERVATIONS AND DISCUSSIONS

6.1 Observations of Bedload

The two events that we have bedload trasport measures for confirm the phenomena's extreme temporal variability and its impulsive nature.

As we have already observed in past flood events, (Tacconi, Billi 1987) once the critical initial discharge has been overcome, the movements of the fluid and sediment discharges appear to continue on their own during the floods and generally the peak sediment discharges precede the highest fluid discharges. Once begun, the movement, across a partial or total mobilization of the armour, seems to trigger an its successive partial stabilization and remobilization of the river-bed matter, yet which is still the subject of study and discussion.

During the Flood n° 3 a few instantaneous arrivals of bedload may be noted, which greatly precede the initial situation of sediment motion, generalizad and continued in time.

These arrivals could be linked to:
- partial lateral erosion of bars a/o banks;
- lateral contributions from tributaries;

- movement of clasts deposited over armour during terminal stage of preceding flood (20/5).

The limited duration and intensity of the main transport event after this probably excludes and actual disintegration in the armour but leads one to think of its settling in the same way suggested for the preceding impulsive arrivals.

6.2 Marked Sample Transport

There are many observations to be made here, yet we should remember that during the Flood n° 4 in the section of river-bed under study, there was on anomalous situation in relation to the others, that concerned the depositing of sediment in the presence of a very low bedload transport.

6.2.1 Number of Rediscovered Pebbles

- Grain-size Sample -

The first observation regards the number of pebbles rediscovered in the river-bed after floods (Tab. 3). This number after the flood n° 2 is equal to 8.8% of the total; after the flood n° 3, it is equal to 8.4% of the total and after the flood n° 4, 4.7% of the total. This last percentage may be explaned through the phenomena of depositing sediment described earlier. This means that 91.2% of the pebbles after the food n° 2 remained embedded; after the flood n° 3, 91.6% of the pebbles remained embedded, and after the flood n° 4, 95.3% of the marked pebbles. The rate of embedment is very high.

If we look at this phenomena with relation to each grain-size class, we see how this rate is very high for the small grain-sizes and diminishes according to exponential law with the rise of the clasts' dimensions. (Tab. 3 and Fig. 6).

One of the causes for this in Virginia Creek could be the smaller chance of being embedded during the transport of larger pebbles together with sorting and the formation of the armour; on the other hand, their greater exposure to the current makes them more easily transported which in turn justifies their relatively high mobility in relation to the smaller grain-sizes.

These are only a few suppositions that need verification made in the face of a far larger number of observations.

The percentage in terms of the weight of the clasts rediscovered after the flood n° 2 is 34.3% (115752 g); after the flood n° 3, it is 31.3% (105449 g); after the flood n° 4 it is 29.2% (98572 g).

During the measurement of the bedload in the flood n° 2 in the face of a bedload transport of 43260 kg, the transited sample (from the measurement section in that period (PC), was 24150 g, equal to 0.0558% of the bedload measured.

$$PC/TS \times 1000 = 0.0558 \qquad (1)$$

Table 3

Clast rediscovered after the floods - Grain-seze Sample

	φ						
	$-4\div-4.5$	$-4.5\div-5$	$-5\div-5.5$	$-5.5\div-6$	$-6\div-6.5$	$-6.5\div-7$	$-4\div-7$
N° of clasts in Sample	2000	1000	500	250	125	60	3935
Total weight of clasts in sample (g)	20000	28000	32000	50000	80000	127000	337000
After the Flood n° 2 of 20.05.1988							
Number of clasts rediscovered	72	73	74	45	36	36	345
Weight of clasts rediscovered (g)	720	2044	4736	9000	23040	76212	115752
Percentage of clasts rediscovered	3.6	7.3	14.8	18.0	28.8	60.0	8.0
Percentage of clasts embedded	96.4	92.7	85.2	82.0	71.2	40.0	91.2

After the Flood n° 3 of 28.05.1988							
Number of clasts rediscovered	57	99	60	48	36	31	331
Weight of clasts rediscovered (g)	570	2772	3840	9600	23040	65625	105449
Percentage of clasts rediscovered	2.9	9.9	12.0	19.2	28.8	51.6	8.4
Percentage of clasts embedded	97.1	90.1	88.0	80.8	71.2	48.4	91.6

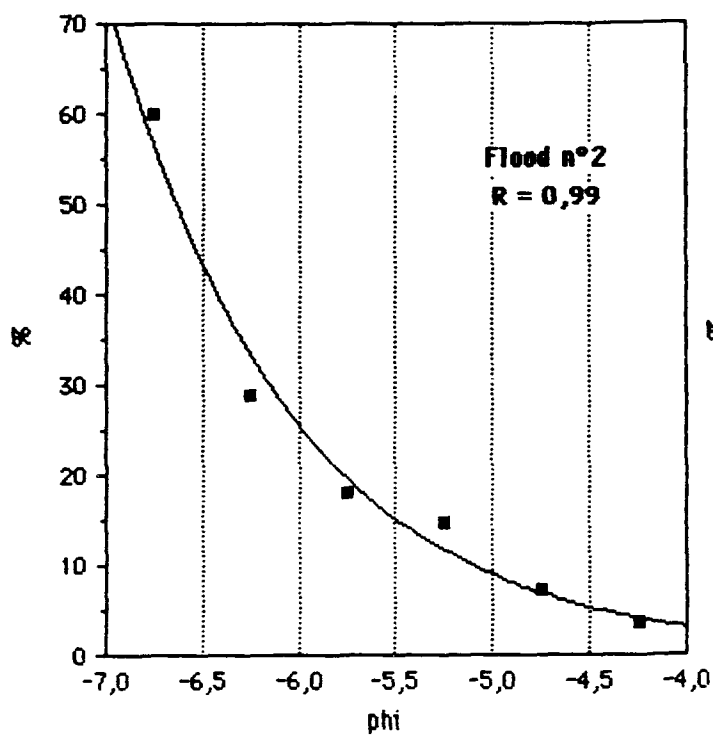| **After the Flood n° 4 of 05.06.1988**							
Number of clasts rediscovered	30	35	34	28	21	36	184
Weight of clasts rediscovered (g)	300	980	2040	5600	13440	76212	98572
Percentage of clasts rediscovered	1.5	3.5	9.0	11.2	16.8	60.0	4.7
Percentage of clasts embedded	98.5	96.5	91.0	88.8	83.2	40.0	95.3
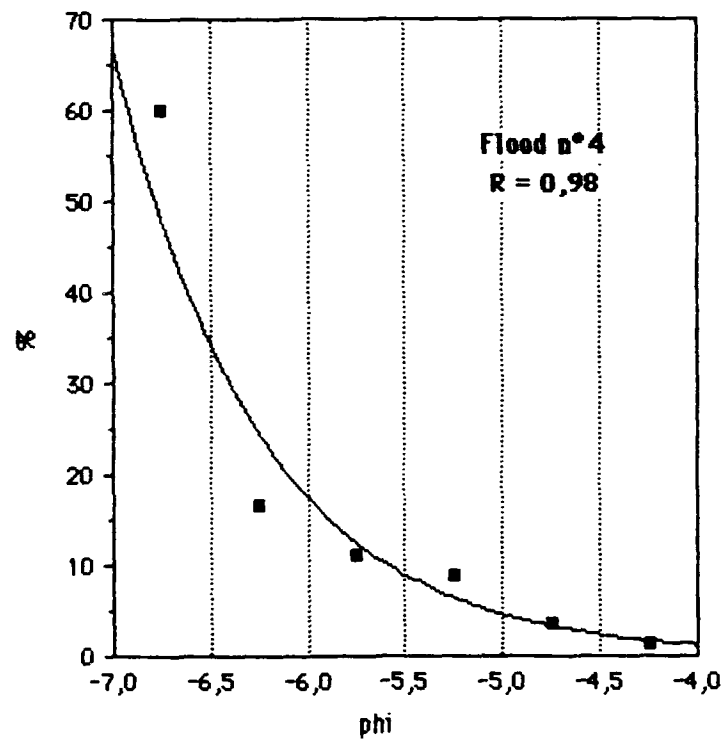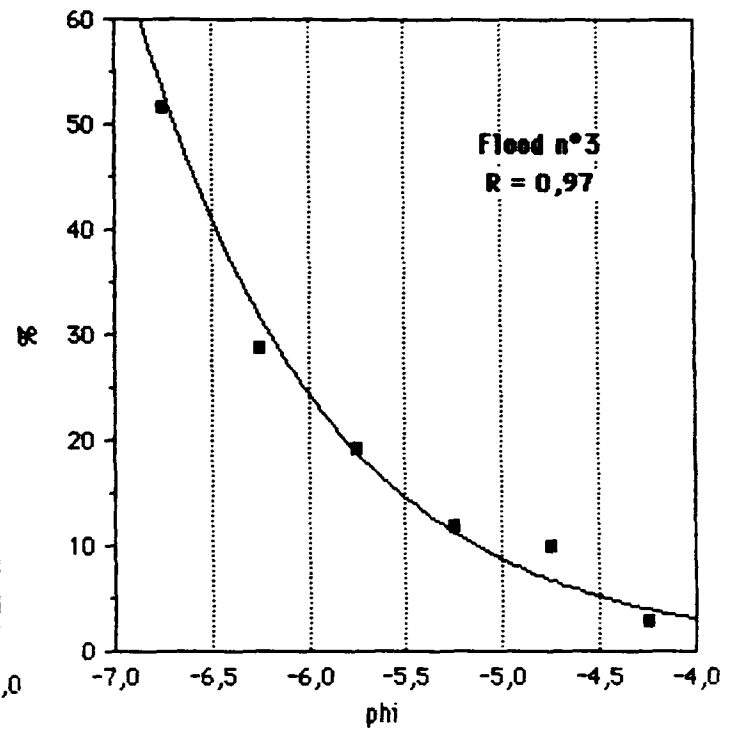
Fig.6. Phi : size of marked pebbles; % : number pebbles renvenued (in %)
in respect of the total marked sample for each grain-size class.

- Morphometric sample -

The total number of pebbles in the morphometric sample is 300 units. After the flood n° 2, 13 pebbles (4.3%) were visible, after the flood n° 3, the preceding 13 pebbles remained on the surface and 2 more pebbles that had been embedded arrived making a total of 15 (5%); the flood n° 4 determined the embedment of 10 pebbles and since there were no new only 5 pebbles remained visible (1.6%).

In Fig. 5 the pebbles of the morphometric sample are indicated together with the identification number and their morphometry.

6.2.2 Position (Shifting) of marked pebbles

- Grain-size sample -

A synthetic view of the clasts' movements in the grain-size sample is shown in Fig. 7 while in Table 4 the statistical parameters of their movements are displayed subdivided into grain-size classes.

The scattering of data is very high; in Fig. 7 the positions of the pebbles subdivided by grain-size classes are shown.

Regarding the three floods observed, in all the grain-size classes there are pebbles that have practically not moved after the floods. The average distances away from the introduction point are almost alike for the grain-size classes, inferior to D_{50} with a slight rise with the increase in φ and a sharp decrease when φ is greater than D_{50}.

There is a similar behaviour, though less accentuated, in the maximum distances (Fig. 8). The surface of the river-bed where marked pebbles were found after the flood n° 2 is almost 2660 m². If we take as the occupied surface of a visible pebble in the river-bed that corresponding to the circle with diameter φ, the total surface area covered by the marked pebbles ($\Sigma\varphi$), inside this area, where they are distributed, is 0.76 m² equal to 0.0336% of the total surface area.

$$\Sigma\varphi/S \times 100 = 0.0336 \qquad (2)$$

- Morphometric Sample -

By means of the number marked on each single clast in the morphometric sample, it was possible to follow the individual movements and the exposure (both on the surface and on those embedded) for each clast after the floods.

Unfortunately the sample, included in a single class of φ (-4.5÷-5.5 φ) was composed of a limited number of components (300) and therefore no statistics may be gathered. The data draw is similar to that of the grain-size sample. In the flood n° 3 where the bedload measurement is complete, it was possible to measure the average velocity of the few clasts available, that varies from 0 to 0.608 m/min. (Table 2).

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Table 4

Statistical data of the position in relation to the
introduction point of clasts in the Grain-size sample

after the flood n° 2 of 20.05.1988	φ						
	-4+-4.5	-4.5+-5	-5+-5.5	-5.5+-6	-6+-6.5	-6.5+-7	-4+-7
N° of marked clasts gathered	72	73	74	45	36	36	345
Distance from the introduction point - mean (m)	245.4	284.3	301.8	192.2	249.4	80.7	234.7
Distance from the introduction point - median	208.0	226.0	279.5	144.0	204.5	54.5	203.0
Distance from the introduction point - minimum	1.0	2.0	1.0	1.0	2.0	1.0	1.0
Distance from the introduction point - maximum	662.0	666.0	665.0	658.0	584.0	495.0	666.0
Distance from the introduction point - st. dev.	193.3	186.9	210.0	185.5	180.2	94.5	194.4
Distance from the introduction point - variance	37367	34961	44104	34393	32459	8930	38565

after eh flood n° 3 of 28.05.1988	φ						
	-4+-4.5	-4.5+-5	-5+-5.5	-5.5+-6	-6+-6.5	-6.5+-7	-4÷-7
N° of marked clasts gathered	57	99	60	48	36	31	331
Ditsance from the introduction point - mean (m)	276.7	265.6	307.0	197.5	221.9	112.8	246.4
Distance from the introduction point - median	233.0	225.0	303.0	161.5	166.5	111.0	207.0
Distance from the introduction point - minumum	7.0	1.0	1.0	2.0	1.0	4.0	1.0
Distance from the introduction point - maximum	723.0	744.0	699.0	674.0	602.0	498.0	744.0
Distance from the introduction point - st. dev.	166.3	195.6	193.7	178.6	173.8	100.6	185.7
Distance from the introduction point - variance	27671	38289	37528	31914	30189	10127	34492

| after the flood n° 4 of 05.06.1988	φ						
	-4+-4.5	-4.5+-5	-5+-5.5	-5.5+-6	-6+-6.5	-6.5+-7	-4÷-7
N° of marked clasts gathered	30	35	34	28	21	36	184
Ditsance from the introduction point - mean (m)	244.1	268.1	198.5	152.8	178.3	107.9	192.2
Distance from the introduction point - median	224.5	215.0	187.5	139.0	183.0	103.0	182.0
Distance from the introduction point - minimum	3.0	2.0	3.0	2.0	3.0	2.0	2.0
Distance from the introduction point - maximum	776.0	745.0	744.0	726.0	510.0	505.0	776.0
Distance from the introduction point - st. dev.	198.3	216.4	175.7	165.2	135.8	107.5	178.8
Ditsance from the introduction point - variance	39322	46840	30887	27304	18454	11548	31963
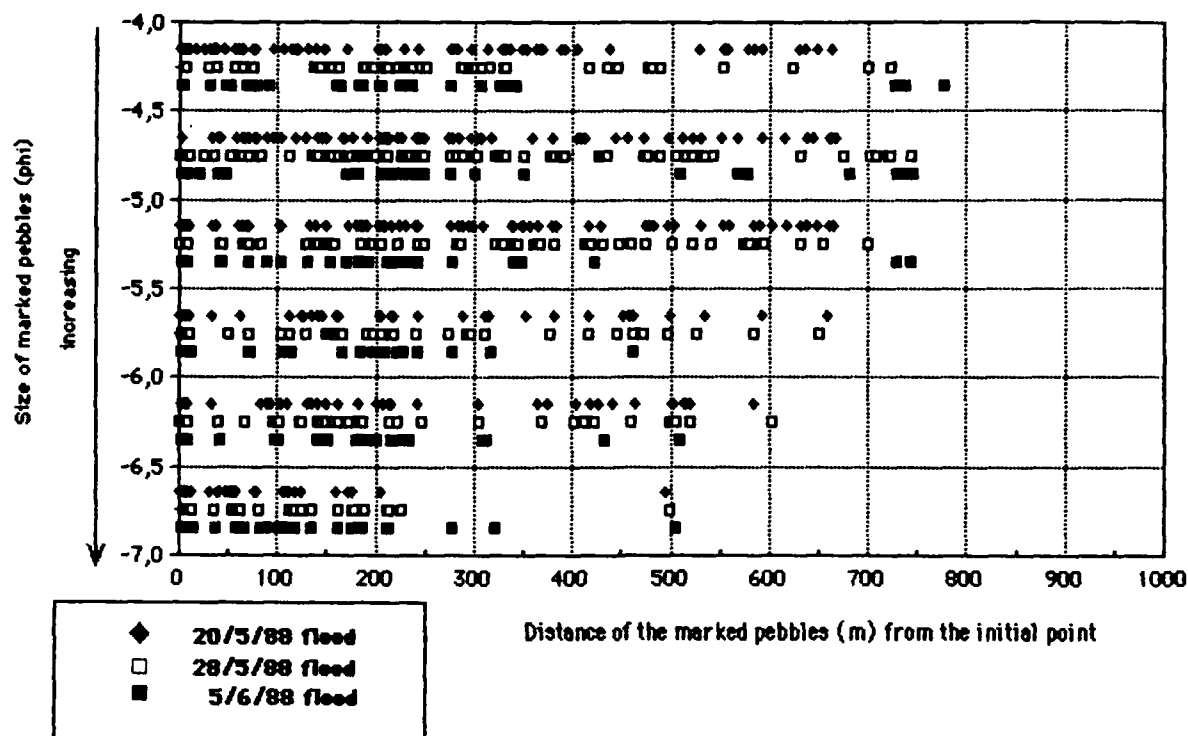
Fiq.7 Distribution of the marked pebbles after the floods
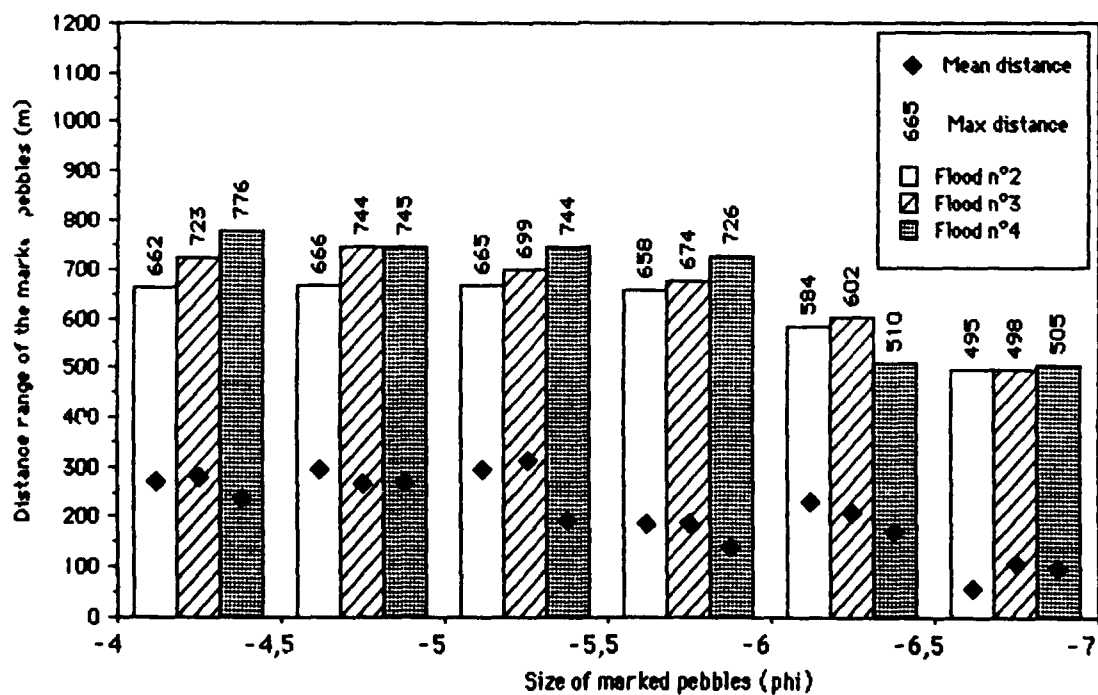in respect of their relative size.

Fig. 8 - Distance Range of the Marked Pebbles (m) in respect of
their relative size

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7. FINAL REMARKS AND FUTURE RESEARCHES

The coarse bedload transport in the Virginio stream seems a more complex phenomena then the fine one. Apart from extreme case, the bedload transport tooks place during very short periods during the floods, and it has a poor correlation with bedload transport during a flood could be indicated as the amount of pebbles going through a section by the distance travelled by them.

Because of the variability of the situations in which each single clast is encoutered during the transport (free, in clusters, embedded, in the bars, in the bed etc.) the distance travelled by the clasts, during the observed floods, it is relatively short and the meam velocity at which they moved has a rate 2 - 3 ~~times~~ less then the water velocity which, during a flood in the examined part of the stream, can reach 6÷8 m/sec.

The measurements and the observations done in the Virginio stream, with many difficulties, are not enough to fully explain the bedload transport in such a stream.

The future researches should be supported and integrated by mean of trough model studies to simulate the recorded events under a more controlled conditions. These researches will study in detail the bedload and furthermore they will deal with the following subjects:
1) defining the relationships between bedload transport and bed dynamic;
2) establishing an indirect methodology to measure the bedload transport by mean of marked samples introduction in the bed and shifting measurements.

About this last point, the Authors are very interested because of the (1) and (2) give similar figures; in other words that means the marked sample dispersion in the bedload measured by the station it is on the same rate of the sample superficial dispersion after the recorded events.

Acknowledgements

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marked sample distribution
following four flood events

schematic map

scale

0 5 10	25	50 meters

granulometric distribution of marked sample

(phi)	symbol	number of pebbles
- 4.0	●	1000
- 4.5	◉	1000
- 5.0	☆	500
- 5.5	◉	250
- 6.0	○	125
- 6.5	●	60

sample shape
(after Folk)

shape	symbol	number of pebbles
elongated	▬	100
compact	■	100
platy	▲	100

A _____ _____ **A** cross section

stream flow

G

4/13

43 21 46

5/20

5/28

6/5

Fig. 5B

tribution
od events

ap

50 meters

of marked sample

er of pebbles

1000

1000

500

250

125

60

iber of pebbles

100

100
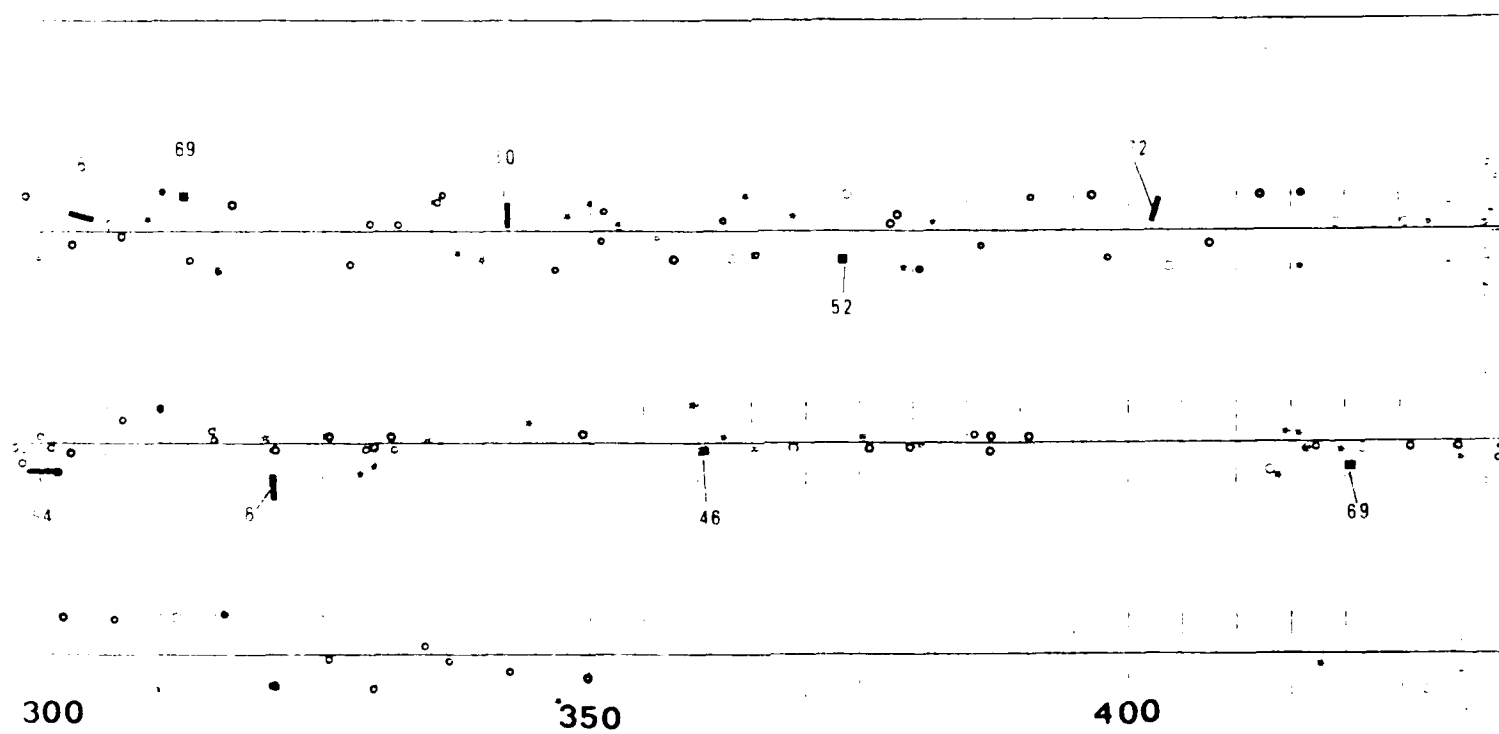
100

cross section

stream flow

G   F   E

lateral bar

4/13

43 21 46

98   38   7

5/20

42

38

5/28

38

6/5

50

Fig 5 C

Fig. 5-D

Fig 5 E

Fig. 5F

61   52

36   3

61   3

450   500   550

Fig. 56.

600   650

Fis. 5H

700   750   800
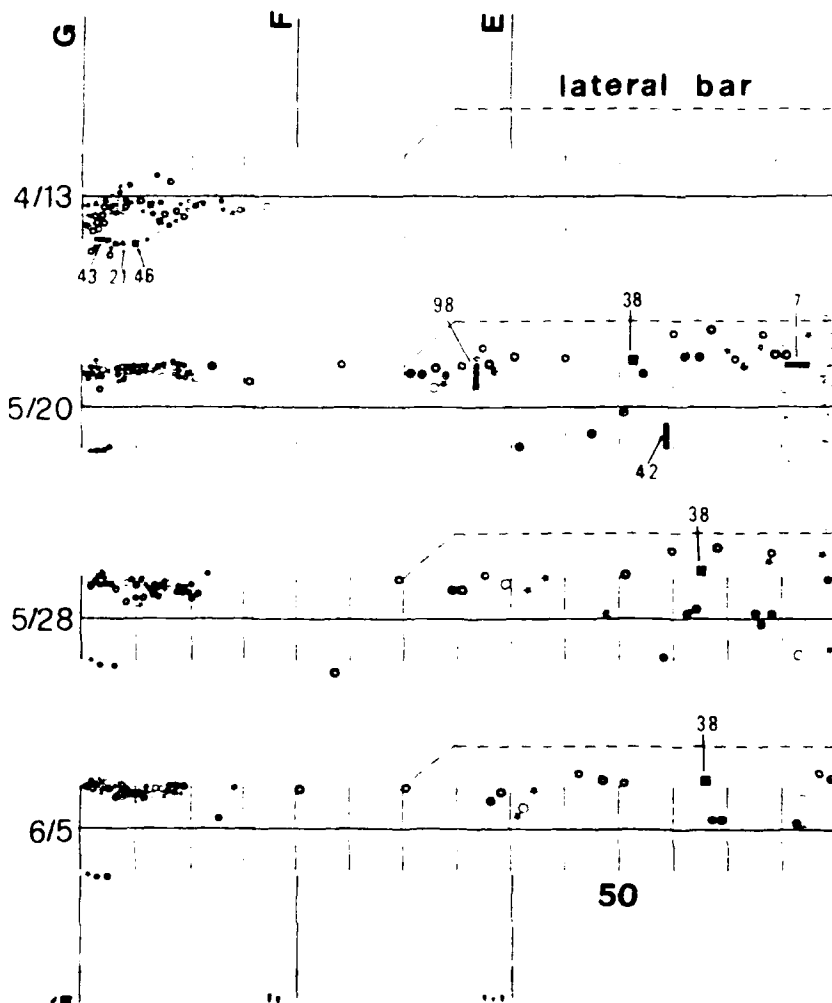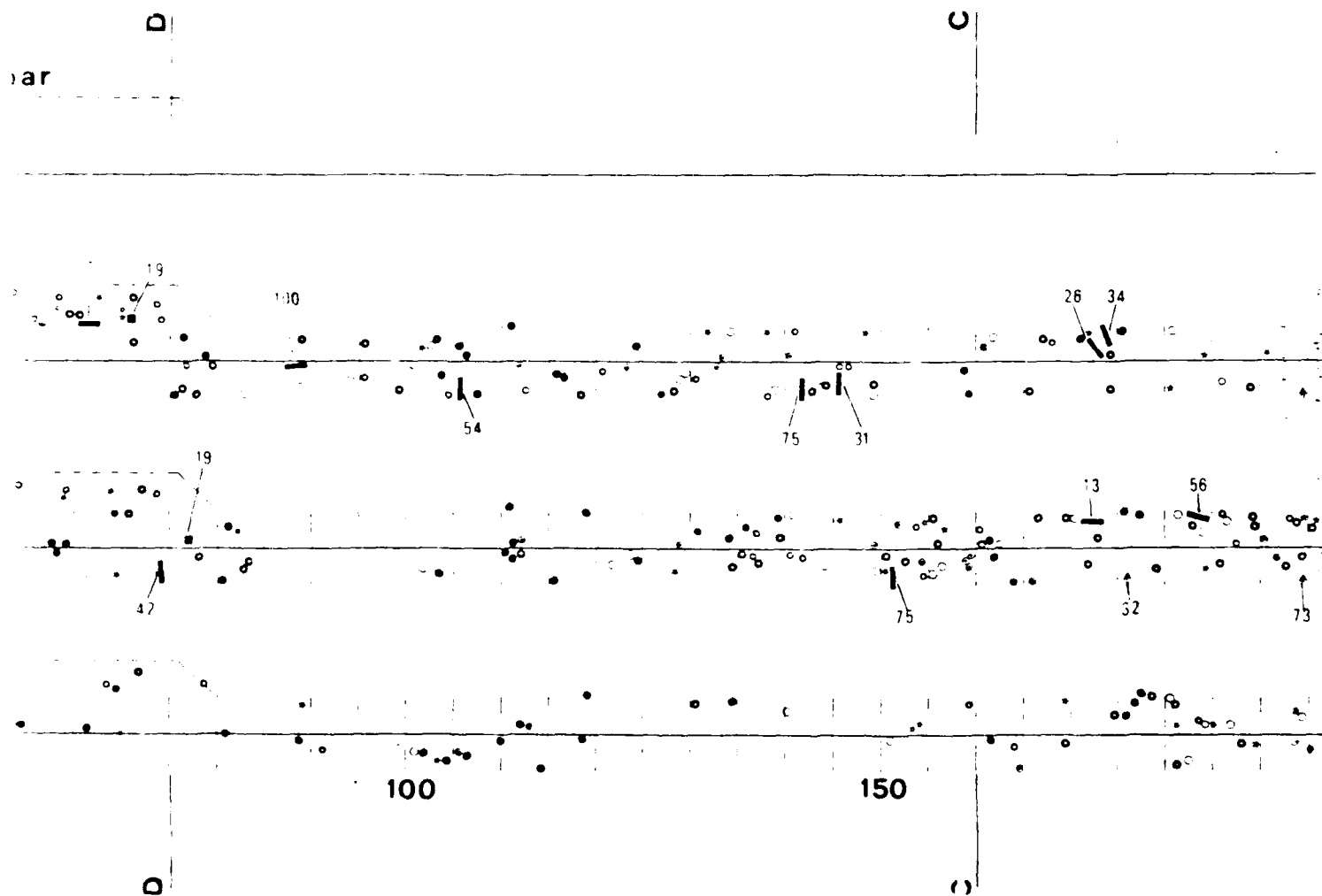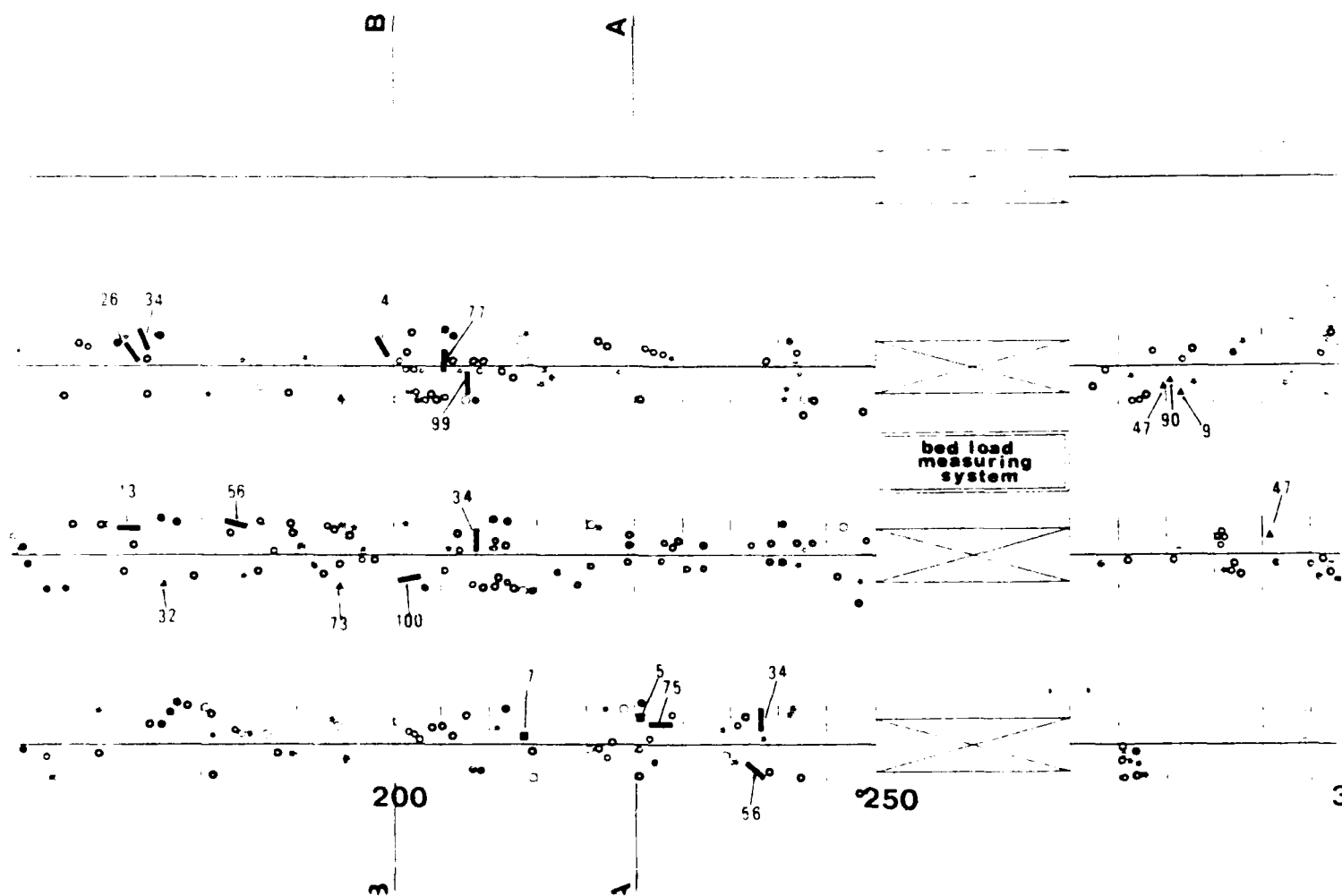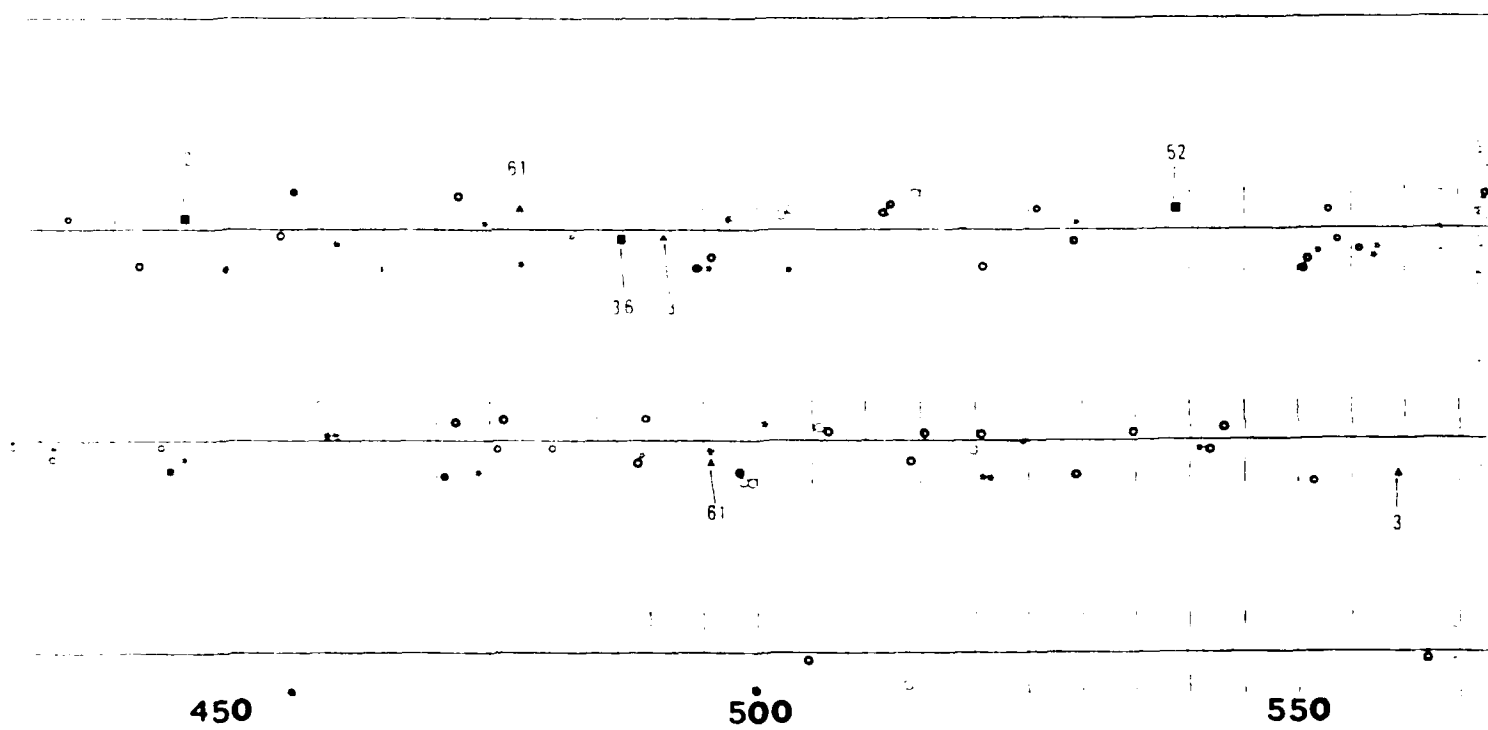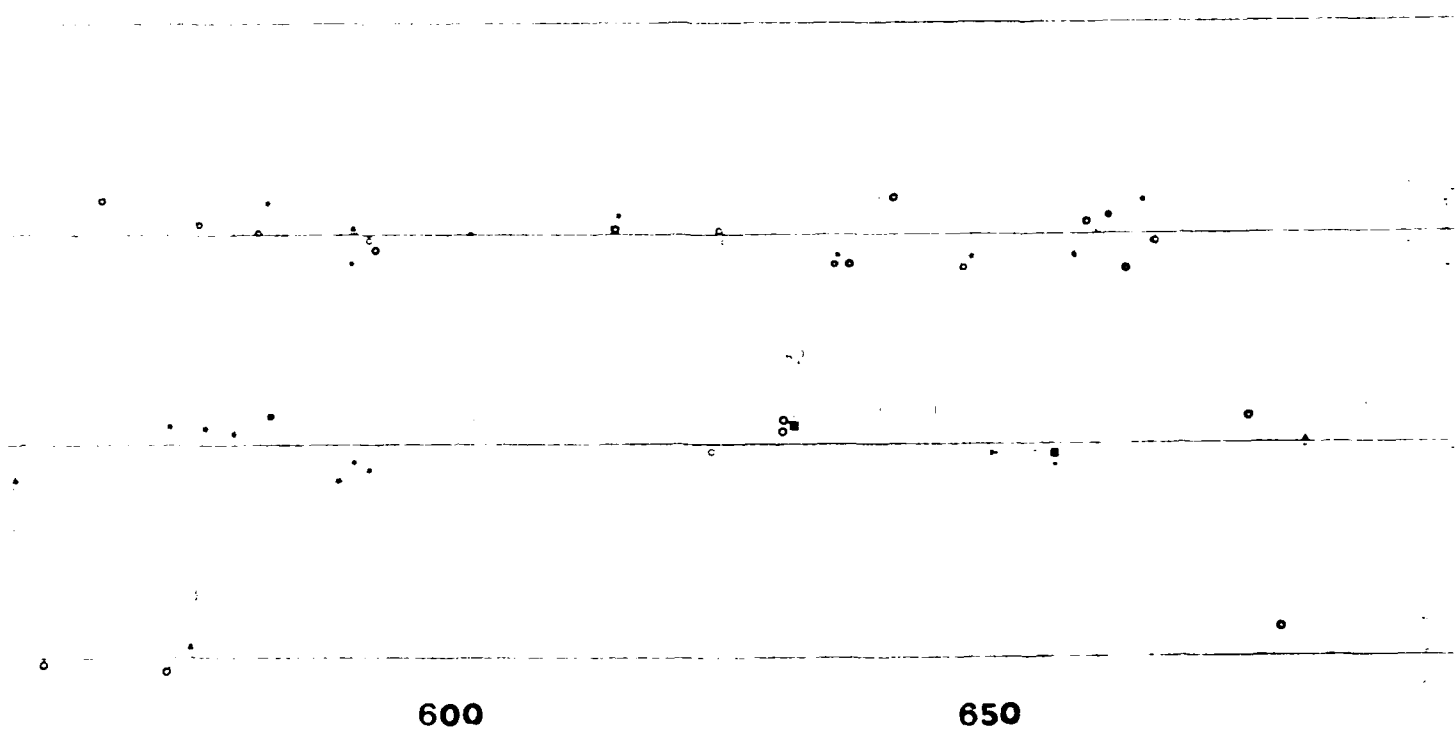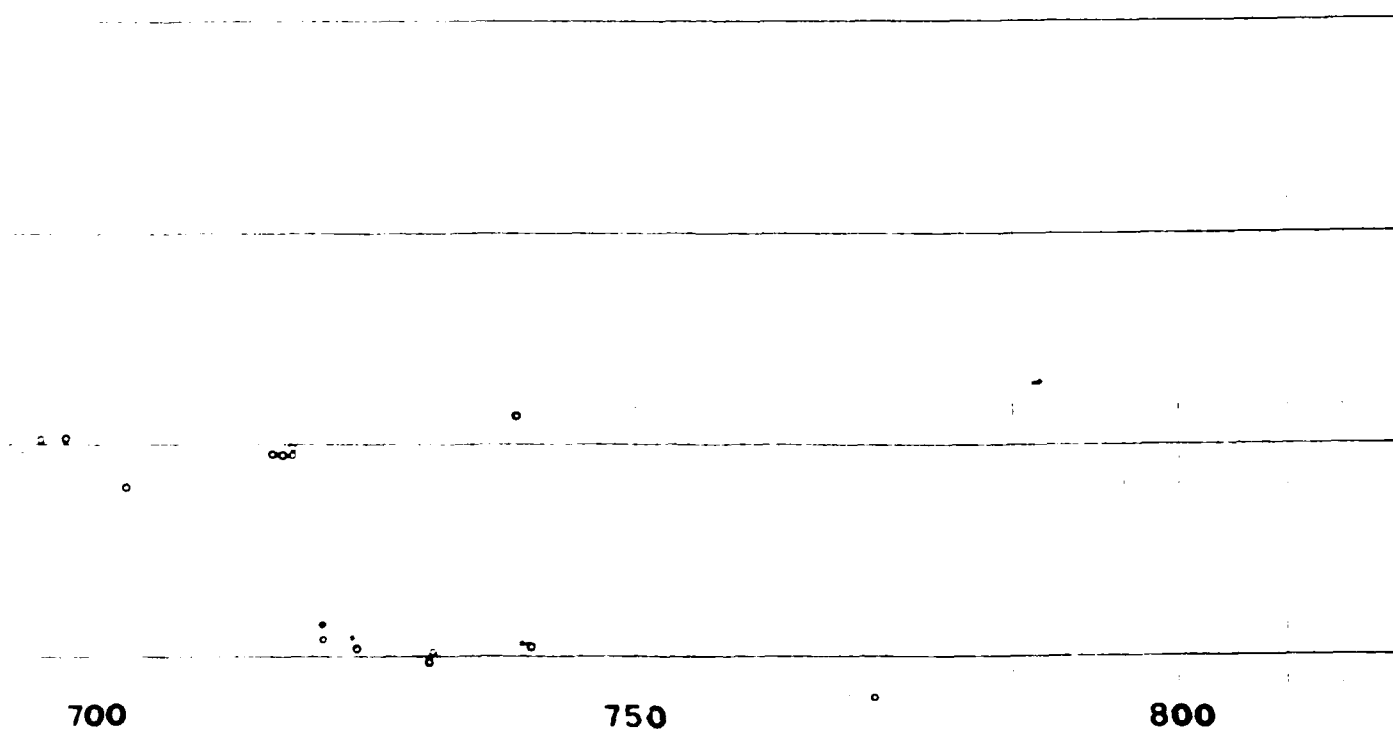
THE MOVEMENT OF INDIVIDUAL GRAINS ON THE STREAMBED

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ABSTRACT

We examine the movement of ensembles of coarse clasts at
low transport intensities, using all available results. The study
confirms the lack of relation between the distance of movement of
individual stone and their size, and emphasizes the stochastic
nature of bedload movement. The distribution of distances of
movement follows the Einstein-Hubbell-Sayre compound Poisson
model or a simple Gamma model for small displacements. For larger
displacements, bars interrupt passage of the particles. The mean
distance of movement is largely independent of mean grain size
for particles smaller than the surface D_{50}. For larger particles,
distance declines to zero very rapidly. The relation between the
mean distance of movement and excess stream power reveals the
scatter typical of all transport relations. All the results
remain noisy so that samples larger than several hundred stones
are needed to finally discriminate the characteristics of
particle displacements.

INTRODUCTION

Sediment transport consists fundamentally of movements of individual particles. H. A. Einstein initiated the modern study of grain displacements, but the topic has not been much pursued because of the difficulty to track and recover suitably large samples of grains. Recently, methods based on radioactivity, magnetism and radio telemetry have changed the situation. The motion of grains is not continuous, but consists rather of a series of steps and rest periods. Altogether, the movement of individual stones appears to be a statistically random phenomenon (Einstein, 1937). This behaviour is similar to the random fluctuations of the local velocity. Einstein (1971) asserted that it should be possible to integrate them into a description of the total pattern of motion of the sediment.

To investigate the motion of individual stones, Einstein (1937) conducted a flume study using marked particles. He found no relation between distance of movement and particle size. Assuming that the rest periods and the actual movements are independent and each negatively exponentially distributed, Einstein described the probability that individual particle would move a certain distance for a known number of steps as a Gamma variate. The distribution of particle displacements after a known interval of time and any number of steps is, then, a compound Poisson process. Using a different approach Hubbell and Sayre (1964) developed the same results.

Flume experiments by Einstein and by Hubbell and Sayre (1964, 1965), and field data from a sand bed river collected by Sayre and Hubbell (1965) showed that the exponential distributions described the travel distances relatively well. Other investigators (cf. Grigg, 1970; Yang and Sayre, 1971) suggested that step length itself followed the Gamma distribution. Substantially more complex statistical models have subsequently been proposed (cf. Shen and Todorovic, 1971; Vukmirovic and Wilson, 1977), but they incorporate information requirements that are not yet feasible to satisfy in the field.

Einstein's theory and subsequent similar approaches have gained relatively little attention and have had relatively little impact on sediment transport research. Nordin (1971) mentioned a few possible reasons for the ignorance of Einstein's work. Among them were the fact that his thesis was written in German and was translated only in the 1960s (by W. W. Sayre), that the practical confirmation and application of such an approach was not clear at that time, and that his famous bedload equation (i.e., 1950) overshadowed it. There may be a more fundamental reason. The arguments in Einstein's thesis are kinematic, and lead to no useful prediction of sediment transport based on summary flow data. But that is what most workers in the field have pursued. However, there is no conflict between Einstein's stochastic

theory and one that seeks an empirical or a quasi-deterministic correlation between mean flow conditions and bulk sediment movement. The sediment transport rate is the product of the virtual rate of travel of the individual particles and the mean depth of the active cross-section of the bed, where the "virtual rate of travel" is the mean distance of movement of the bed material during a known interval of time. In other words, the rate of transport is determined by the average characteristics of the individual displacements of the moved particles, which may very well be correlated with mean flow conditions.

In gravel bed streams, tracer programmes were started in the early 1960s (Takayama, 1965a) in attempts to evaluate bedload movement. Instead of describing the process, most of the investigators concentrated on the expected relation between the distance of movement and particle size (e.g., Keller, 1971; Schick and Sharon, 1974; Butler, 1977; Carling, 1987; Ashworth and Ferguson, 1989). Others examined the influence of the sedimentological characteristics of the bed on the movement of individual particles (cf., Laronne and Carson, 1976; Brayshaw et al., 1983). Most of the tracer work confirms the lack of relation between distance and grain size, mentioned more than 50 years ago by Einstein (1937).

The entrainment, transport and deposition of sediment are affected by a large number of interrelated variables. These can be grouped into three categories: flow, material and channel form (Fig. 1). Some material and channel morphology variables have not been satisfactorily quantified. Prominent amongst these are the bed surface structures that characteristically develop in heterogeneous gravels. Bed structure includes microstructures, packing and armouring. "Microstructure" refers to the specific arrangement of a number of clasts whereby they constrain each other from movement. Imbrication is the best known example. Combined effects of the three categories change over space and time. Under the same conditions of flow and channel morphology, bed material may exhibit a wide range of structure and packing. Accordingly, quantities, distances, and other attributes of the moving sediment vary. Figure 1 indicates the complexity of the resulting sediment transport process.

In this paper we present descriptions of displacement characteristics of coarse particles, based on relatively large samples with high recovery rates. The material moves as bedload in a "low intensity" transport regime. That is, individual grain movements are isolated events, and the characteristic, coarse surface layer remains in place (cf. Parker and Klingeman, 1982). On the basis of these results, we seek a general description for the pattern of individual grain movements and we examine the possibility for there to be a scale-free sediment transport relation based on mean travel distance. We will show that effects produced by the channel form and material variables substantially complicate the results in river channels beyond those described from experimental studies.

THE DATA

Data presented in this paper were obtained in various studies (Table 1). They derived from observations of painted and magnetically tagged stones. The rivers examined cover a wide range of bed texture and structure, and channel morphology. In addition, they cover a wide range of hydrological regimes, from flash floods in the desert to the opposite extreme of gradually varied snowmelt floods. For more descriptive material about the study rivers, see the cited references (Table 1).

In most of the field studies painted stones were used and the data are limited to the surface exposed particles only. In many of these studies, the recovery rate is low. Therefore, the recovered particles do not represent the entire sample, part of which will have become buried. The studies which used magnetically tagged stones include buried particles, in addition to exposed ones, with relatively high recovery rates. In these studies the recovered particles do represent the entire sample.

In most cases the tracer stones were larger than the bed material, but generally they represent reasonably the sizes that make up the surface layer. In the first event of each study, the tracer particles are initially placed on the bed surface and are not incorporated into the bed structure, so they move more readily than normal bed material.

DISTANCE OF MOVEMENT AND PARTICLE SIZE

Under controlled conditions in his flume, Einstein (1937) tracked the movement of painted material ranging in size between 3 and 40 mm. Some of his experiments were made using a very narrow range of sizes. Table 2 shows that, for material of the same size and under similar flow conditions, some stones moved long distances whilst others remained stationary. Stelczer (1981) similarly followed the movement of single, or few, stones in the Danube River. He relocated particles after 6 or 24 hours. Table 2 shows that a single stone, under similar flow conditions, moved varying distances in successive steps. These results demonstrate clearly the stochastic nature of the entrainment and movement of individual particles in gravel bed rivers.

The consequent lack of relation between distance of movement and particle size has been reported in many field studies. However, most of these studies used tracers with a relatively narrow range of sizes, always coarser than the bulk bed material. A wide range of sizes might still reveal some relation. Figure 2 illustrates three cases. In Nahal Hebron the range of sizes was narrow and the data do not show any particular trend. In contrast, there is the appearance of an envelope over widely graded Harris Creek data suggesting that large particles

travel shorter distances. Data of Allt Dubhaig are intermediate in both respects. All the data show that stones may move any distance beneath the envelope, so that functional relations between size and distance do not occur for individual particles.

THE INFLUENCE OF PARTICLE SITUATION

Random entrainment of individual stones is due to the turbulent flow and the arrangement of individual grains which varies widely over the bed. In gravel bed rivers grains may lean upon each other in many ways.

At the end of a flow event, some of the particles are found free from constraint on the bed surface, others are locked within the surface layer and the rest are buried within the scouring layer. These three groups define the final situation of particles after a competent flow event and the initial situation before the next event. The initial situations affect the flow required for entrainment. This might ultimately influence the travel distance of a stone during a flow event, hence the mean distance of movement of each group. On the other hand, the final situation might reflect the average distance of movement of a group, since buried -- and perhaps locked -- particles probably stop moving before ones which remain free.

Competent flow events change the particle position relative to the local bed surface elevation (Hassan, in press). Some of the initially exposed particles (free and locked) are buried, and a portion of the buried ones become exposed. Transition frequencies for 282 particles in Nahal Hebron are given in Table 3. The frequencies are related to flow magnitude. Similar results have been obtained from Carnation Creek, but as yet our results are insufficient to tell whether a general model may describe the phenomenon.

Table 4 summarizes the mean distance of movement of particles classified according to their initial or final situation from all discriminating observations known to us. In the initial event after stones placement in Nahal Hebron and Carnation Creek, particles which were buried at the end of the event had moved farther, on average, than the free ones. The reverse result was obtained from other recorded events. Grouping particles according to their final situation revealed results similar to those based on the initial situation. Although the differences between the groups are not statistically significant, there is a consistant tendency for tne locked and the buried particles to move shorter distances than the free surface stones, except in some initial events. In addition, there appears a tendency for post-event locked and buried stones to have moved proportionally less far than ones classified in these groups before the event. This outcome is interpreted to indicate that stones trapped during the event may be definitively stopped at any time -- perhaps after only a small distance of movement.

The mean distance of movement of particles buried during the 23.1.83 event of Nahal Hebron and the single event of Carnation Creek, both initial events after placement of stones on the surface, was higher than for the other groups. This result occurs because burial of initially exposed particle implies the occurrence of scour, transport and fill. On the other hand, many exposed stones remained that way because they were marooned in relatively inactive parts of the bed. The outcome is produced, then, by the unnatural distribution of stones within the scour layer before the event.

DISTRIBUTION OF DISTANCE OF MOVEMENT

The distribution of particle displacements is available for complete flood events for several of the streams of Table 1. The range in grain size of the tracer particles is relatively restricted, so the influence of grain size should not strongly affect the results (cf. fig. 2c). Two questions can be asked about the movement of grains; first, what is the characteristic displacement of the bed material, and second, what is the distribution of the entire sample after every flow event. The first question is concerned with the virtual rate of travel. In this case we deal with the moved particles only. The second question is concerned with the evolution of the streambed. In this case, the entire sample of tracer material should be considered, including the stones that did not move.

Figure 3 shows observed distributions for both cases. The data are, in general, highly skewed -- usually monotonic. Although the event distributions tend to be very irregular, distinct secondary modes appear in some examples.

The statistical model of Einstein-Hubbell-Sayre (hereafter referred to as E-H-S) for grain displacements can be examined in light of these data. Einstein represented the model as

$$f_t(X) = \exp(-X-T)\Sigma(X_2^n T^{n+1})/n!(n+1)!$$

in which $T = t/t_0 = 2<x>^2/<x_s^2>$ is the number of average rest periods in the event (a scaled event time);

$X = x/x_0$ is the number of average step lengths (a scaled distance);

$x_0 = <x_s^2>/2<x>$ is the average step length;

$t_0 = t(<x_s^2>/2<x>^2)$ is the average resting time;

$<x>$ is the average displacement

$<x_s^2>$ is the variance of the displacement.

The fitted distributions are displayed in figure 3, and the similarity between the observations and these distributions was examined using the x^2 test at the $a = 0.01$ level (chosen, in light of the noisy data, to bias the test toward acceptance of the null hypothesis). Results are reported in table 5.

Five of the nine tests indicate that the E-H-S model provides a satisfactory description of particle movements (Table 5a). The successful descriptions derive from simple rainstorm floods of moderate magnitude and small mean movement (Allt Dubhaig; Nahal Hebron; Okawa; among others), whereas unsuccessful description are associated with long, complex snowmelt events (Seale Brook; Harris Creek) and major, multi peak rainstorm flows (Carnation Creek; Nahal Hebron 23.1.83) during which stones travel farther. Results for the distribution of all stones (Table 5b) are the same, except that two of the Japanese streams now have unsuccessful fits. In these channels there was a very high proportion of zero displacements

In events with moderate displacements and simple distributions, it is possible to suppose that most stones move only 1 to 2 times, so that a simple Gamma distribution might adequately describe the distribution of movements. We tested this proposal under the same conditions as our previous test (Fig. 3; Table 5). The pattern of successes is similar to that for the more realistic E-H-S model, except that Seale Brook now is successfully predicted. There are some spectacular improvements in goodness-of-fit.

The unsuccessful cases tend to have long tails (a few stones travel a long way), and to exhibit a secondary mode of displacement lengths, very pronounced in the case of Carnation Creek. The implication is that a trap exists for mobile stones atfer some moderate displacement, but that a good many stones may bypass it. The trap locations are channel bars, the major accumulations of mobile channel-bed sediments. We have confirmed this by direct test in Nahal Hebron and by observation of stone recovery positions in Carnation Creek. In a more detailed investigation Hassan, Church and Schick (in review) showed that truncation of the long tail in some cases produces successful statistical descriptions of the short range displacement using either the E-H-S or Gamma models.

We conclude that the movement and pattern of displacement (dispersion) of bed material can be described in the short range for individual flow events by relatively simple statistical models of a random process, as originally proposed by H.A. Einstein. However, when displacement distances become comparable with the "morphological scale" of the channel, the movement of the individual stones ceases to be completely random. The large scale features of the channel distinctly influence the transport phenomenon. A more complex stochastic model will be necessary to explain this phenomenon. It will be a salutary model in that it must incorporate a description of the principal sedimentary features of the river channel -- a consequence of sediment movement that is wholly lacking in contemporary descriptions of sediment transport.

MEAN DISTANCE OF MOVEMENT AND MEAN SIZE

To examine the possibility to establish a practical correlation between particle characteristics and sediment movement, we studied the relation between the mean distance of movement and the mean size. The analysis was restricted to free, surface particles only (a particle must be freed before it can be moved). For free particles the common assumption is that the travel distance is proportion to 1/D, where D is the grain size. This relation does not hold for locked or buried stones. The data examined here cover a wide range of flow events and grain sizes, so we scaled both variables.

Despite the scaling, the data still exhibit scatter (Fig. 4), which may be due to errors in measurements and to relatively small sample sizes. In general, the scaled distance of movement does decrease with the scaled size, declining rapidly when the scaled size is greater than about 2. The average armour ratio of the data is equal to 2. In other words, when the size of the material is larger than the median size of the surface material, mobility declines rapidly. This result agrees with Einstein's (1950) assertation that the hiding factor is important for material finer than the median size of the surface material, and is consistent with the very steep envelopes in figure 2. Figure 4 also indicates that travel distance is rearly independent of grain size for smaller grain sizes for which hiding factors and traps (comprising openings between the larger particles) are apt to promote a highly stochastic process.

MEAN DISTANCE AND STREAM POWER

Sediment transport rate is the product of the virtual rate of travel and the depth and width of the active layer on the streambed. Here we present the correlation between distance of travel and a measure of flow conditions on an event basis. We used Bagnold's (1966) specific stream power as a matter of convenience and we calculated the "excess stream power" using Bagnold's (1980) methods.

Figure 5 presents all of the available data. The threshold values vary substantially and may be supposed to represent the combined effects of particle inertia and structural constraints to entrainment. The variation might also be a result of errors, especially in hydrological measurements. It is typical of the variability found in all displays of bed material transport. The functional relation between the mean distance of movement and excess stream power is $<L> = 1.3 (w-w_0)^{1.2}$, where $<L>$ is the mean distance of movement and $(w-w_0)$ is the excess stream power. Figure 6 also shows an upper envelope line for nearly all of the data. The envelope line is interpreted to represent the relation when bed structure exercises no significant influence over the threshold condition.

For comparison the figure shows some observations of gravel movement on a sand bed (Leopold et al., 1966). In sand bed rivers there is no structural constraint to movement. Individual large clasts may be transported rapidly over the sand (Fahnestock and Haushild, 1962), although individual stones may episodically become buried. For these data a poorly defined functional relation is $<L> = 78 \ (w-w_o)^{0.9}$. The distances of movement of gravel on a sand bed are an order of magnitude larger than on a gravel bed.

It is interesting to speculate that, for a transport function $q_b \ \alpha \ (w-w_o)^{1.5}$, these results imply that the mobile (scour) layer increases as $(w-w_o)^{0.3}$ in gravel bed channels, and as $(w-w_o)^{0.6}$ in sand bed channels.

SUMMARY

We have examined the displacement characteristics of coarse bed material moving as bedload in the "low intensity" transport regime. The study confirmed the lack of a simple, general relation between the distance of movement and particle size. However, observations using a wide range of stone sizes indicate that distances travelled by large particles decline rapidly. Flume and field data under similar flow conditions demonstrate the stochastic nature of bed material movements. Grouping particles according to their initial or final situation before or after a flow event indicates that free particles generally move greater distances than locked or buried ones.

The compound Poisson model of Einstein-Hubbell-Sayre and a simple Gamma model for the distribution of movements were tested against observed distributions of particle displacements from natural streams. Both models fit the data reasonably well for small mean displacements, but for larger displacements bars act as traps and the distribution is no longer completely random.

The relation between the scaled mean distance of travel and the scaled mean size indicates that travel distance is largely independent of grain size for particles smaller than $2D_{50}$ of the subsurface material. For larger particles, distance declines to zero very rapidly. The discriminating point of particle behaviour closely approximates the mean armour ratio in the study streams.

The correlation between the mean distance of movement and the excess stream power is similar to that normally obtained for sediment transport. The distance of movement of gravel on a sand bed is an order of magnitude larger than on a gravel bed.

The field data remain very noisy, and further research is needed to characterize all aspects of the particle displacements. The bed characteristics and the morphological elements of the

channel influence the statistical homogeneity and the randomness of the process. This explains part of the variability in the data. The other part is the outcome of the inherently high variance. To overcome this problem, Stelczer (1981) and Hassan, Church and Schick (in review) recomended very large samples in order to ascertain the statistical characteristics of particle displacements.

The bedload consists of movements of individual particles, yet the characteristics of the individual particles alone do not explain the complexity of the interaction between flow, material and channel (Fig. 1). As demonstrated in this paper, structure, morphology and sedimentology of the bed exercise important influence over the bedload movement. A successful model should consider all of these factors.

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Table 1

Data of gravel bed rivers with particle travel measurements

Stream	Grain size (mm) subsurface D50	surface D50	tracers	Number of events	Tracer method	Number placed	Recovery rate %	Source
Cobble gravel channels, Israel								
Nahal Hebron	35	70	45-180	4	magnet	282	90-93	Hassan (1988)
Nahal Og (West Bank)	15	35	45-180	2	magnet	250	55-56	
Sandy gravel channel, Israel								
Nahal Yael	10		pebbles-cobbles	8	paint			Lekach & Schick (pers. comm.)
Welsh upland cobble channels								
Nant Calefwr			cobbles	1	paint	35	85-100	Slaymaker (1972)
Nant Y Grader 9A	36		cobbles	2	paint	35	85-100	
Nant Y Grader 9B	34		cobbles	2	paint	35	85-100	
English upland cobble channels								
Carl Beck	50		15-130	2	paint	120	98	Carling (pers. comm.)
Great Eggleshope	20		15-130	1	paint	85	78	
Scottish upland cobble channel								
Allt Dubhaig 1	34	98	11-147	4	paint	279	49-73	Ashworth (1987)
Allt Dubhaig 2	26	46	41-238	4	paint	150	35-92	
Allt Dubhaig 3	29	41	27-153	4	paint	170	36-93	
Allt Dubhaig 4	29	42	26-170	4	paint	135	30-88	
Allt Dubhaig 5	15	23	24-135	3	paint	143	80-96	
Gravel bed river in Japan								
Fukogawa	35		22-128	5	paint	300	61-98	Takayama (1965a, b)
H.Yakawa	35		22-128	5	paint	500	08-81	
Ckawa	40		22-128	5	paint	500	42-81	
Gravel bed rivers in Canada								
Carnation Creek	25	60	16-180	1	magnet	250	80	This study*
Harris Creek	20	91	06-512	1	magnet & lithology		75	This study**
Seale Brook	40		06-250	1	paint		05	Laronne (1973)
Danube River, Hungary			11-32		radioactive	few	100	Stelczer (1981)
Flume			17-24		paint	649		Einstein (1937)

* Reach described in Church et al. (in press)
** Reach described in Chamberlin, ed. (1988)

Table 2

Flume and field data of tracer particles under similar flow conditions

Flume (Einstein, 1937; plates 2, 3, 4, 5, 6, and 7)

Number of stones	Number of measurements	$<x>$ (m)	x_S (m)	Size range (mm)	Duration (mins.)
	649	11.77	7.67	17-24	5
	643	9.08	7.07	17-24	60
	619	9.57	4.91	17-24	20
	344	7.86	6.93	< 17	5
	334	13.92	5.46	< 17	6
	291	20.91	6.60	> 24	3

Danube River (Stelczer, 1981; App. 1, table 22)

1	8	12.84	17.75	24.4	360
1	9	4.56	7.79	29.8	360
1	11	8.57	6.31	13.1	360
1	11	4.98	6.33	15.1	360
1	11	1.02	1.72	26.1	360
4	16	18.63	9.44	22-32	360
5	60	5.97	11.56	11-16	360
5	18	6.47	11.98	22-32	1440

$<x>$ mean distance of movement
x_S standard deviation of the distance of movement

Table 3

Transition frequencies for burial/ exposure of coarse clasts in Nahal Hebron (N = 282)

Event Date Q_{max} (m^{3}s^{-1})	19.1.83* 9.2	23.1.83 33.0	17.10.84 18.0	8.11.86 49.8
Exposed By Event		0.10	0.14	0.10
Remain Exposed	0.66	0.27	0.21	0.18
Total Exposed After Event	0.66	0.37	0.35	0.28
Exposure Frequencies		0.29	0.22	0.15
Buried By Event	0.34	0.39	0.16	0.17
Remain Buried		0.24	0.49	0.55
Total Buried After Event	0.34	0.63	0.65	0.72
Burial Frequencies	0.34	0.60	0.43	0.49

*first event after clast placement on the surface

Table 4

Mean distance of movement of particles grouped according to
their initial or final position before or after the event.

		Initial Position				Final position		
Event	Position	$<x>$ (m)	x_S (m)	N		$<x>$ (m)	x_S (m)	N
Nahal Hebron								
19.1.83	Free	10.2		221		9.9	14.9	135
	Locked					10.7	15.3	86
23.1.83	Free	57.9	35.2	126		51.0	35.5	72
	Locked	63.5	35.5	73		47.2	36.7	52
	buried	No Data				70.8	33.4	138
19.1.83& 23.1.83[**]	Free	69.3	37.7	126				
	Locked	73.2	37.2	73				
	Buried	79.9	32.3	51				
17.10.84	Free	18.1	18.9	65		16.4	17.0	65
	Locked	11.5	15.0	52		9.2	13.9	57
	Buried	3.7	9.8	144		5.2	13.7	139
8.11.86	Free	95.8	159.2	67		159.7	157.2	49
	Locked	54.9	124.1	34		54.6	111.9	19
	Buried	57.2	124.1	149		37.5	112.8	182
Nahal Og								
8.11.86	Free[+]	146.5	137.9	140		217.8	182.7	34
	Buried					134.3	119.2	106
6.1.87	Free[+]	19.7	47.0	24		21.0	83.4	27
	Buried	12.5	63.2	74		11.9	50.4	71
Seale Brook								
Spring 1972?	Free	86.5	51.85	242		93.3	63.6	88
	Locked					81.2	72.4	54
Carnation Creek								
Jan. 1990	Free[+]	52.6	45.8	183		36.2	42.8	72
	Buried					60.4	45.7	111

[+]   includes a few locked particles
[**] in the initial situation, two means represent the
      23.1.83 event. For the buried group we measured only
      the combined distance of the two first events
for notations see table 2

Table 5

Statistical characteristics of particle movements in field events.

River	a) Moved particles only							b) All particles				
	$\langle x \rangle$	x_s^2	T	x^2_E	n	x^2_G	$x^2_{0.01}$	$\langle x \rangle$	x_s^2	T	x^2_E	$x^2_{0.01}$
Dubhaig 2	22.8	370.2	2.81	9.4	1.05	11.3	21.7	17.5	375.6	1.63	5.9	26.2
Dubhaig 3	18.8	347.0	2.04	15.9	0.85	9.73	21.7	12.6	310.5	1.02	7.4	24.7
Fukogawa	17.8	234.4	2.71	19.5	1.19	14.2	23.2	10.7	216.4	1.05	8.1	24.7
Hayakawa	37.8	2642.0	1.08	19.0	0.57	39.4	23.2	30.0	2334.8	0.77	32.7	26.2
Okawa	5.23	10.63	5.15	9.5	2.04	14.3	18.5	2.29	11.36	0.92	83.8	27.7
Carnation Creek	52.59	2096.7	1.32	301.0	0.98	131.2	27.9	51.0	2118.8	2.46	223.0	26.2
Seale Brook	86.5	2689.1	5.57	41.1	1.14	13.3	24.7					
Harris Creek	92.9	9254.4	1.87	56.3	0.79	83.1	30.6	58.2	7814.6	1.28	148.1	30.6
Nahal Hebron												
23.1.83	53.9	1172.4	6.20	20.9	2.06	26.7	20.1	51.1	1172.4	4.46	235.1	18.5
17.10.84	19.4	300.3	2.51	18.0	0.73	4.7	18.5	16.9	316.8	1.80	18.6	20.1
8.11.86	166.1	23483.5	2.35	19.4	1.03	17.4	23.2	142.0	29261.5	1.38	16.7	23.2

for notations see text
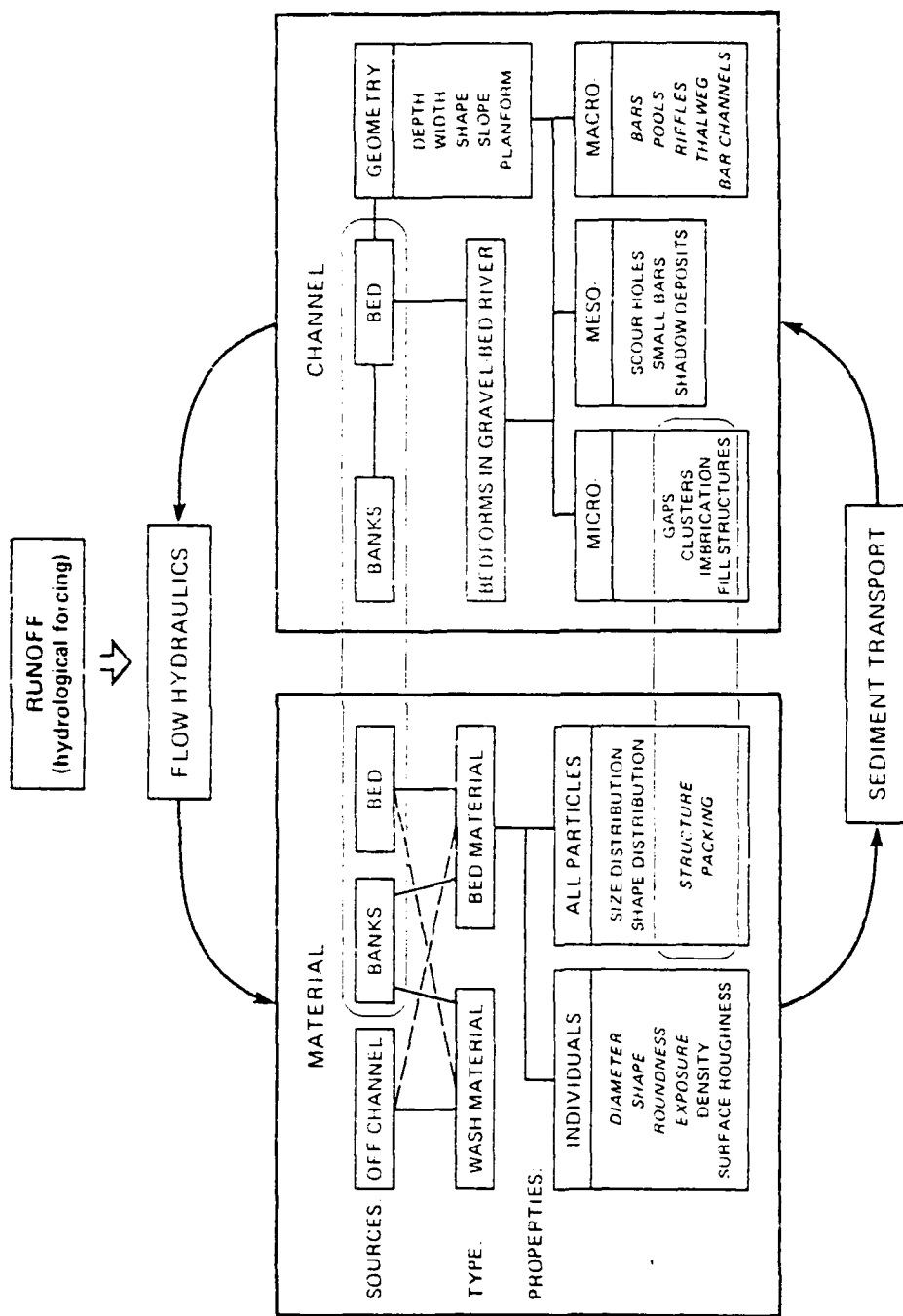
Fig. 1: Relation amongst material, flow and channel
properties in gravel bed rivers.
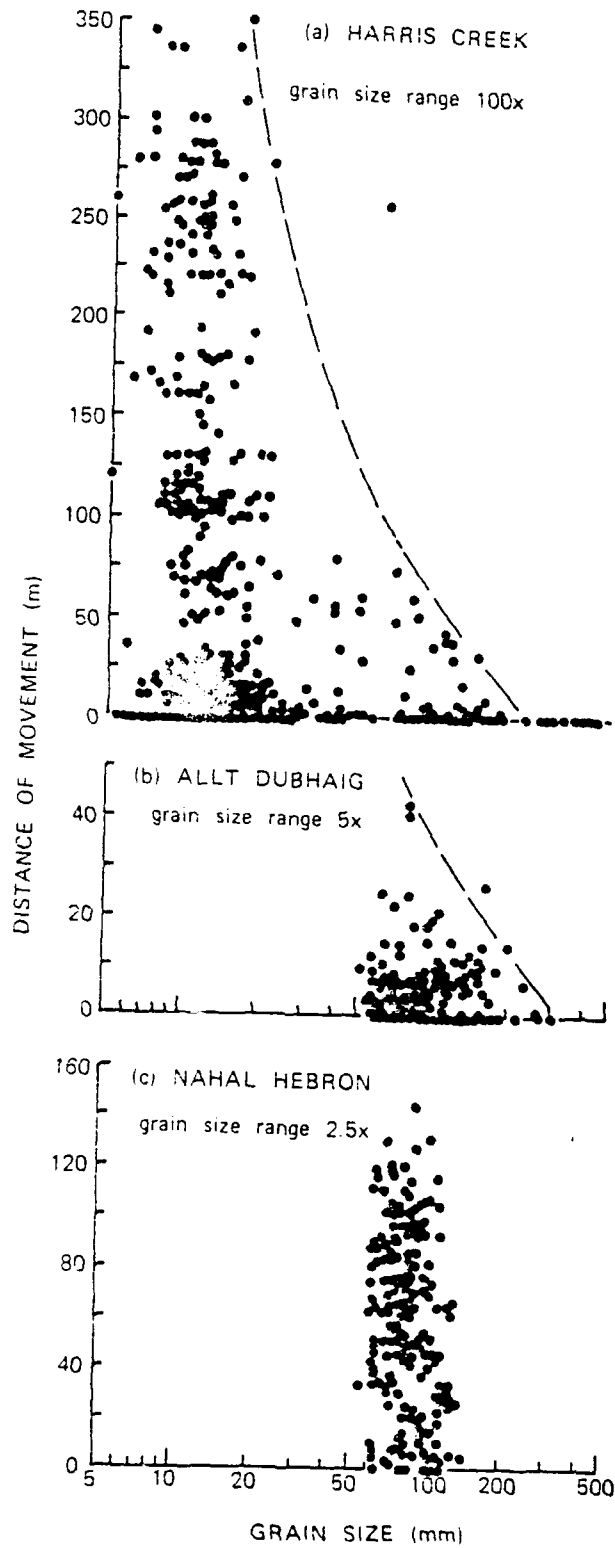
Fig. 2: Distance of travel versus particle size in (a) Harris Creek; (b) Allt Dubhaig River (data of P. J. Ashworth) and (c) Nahal Hebron. The semilogarithmic plot allows stones that did not move to be plotted.
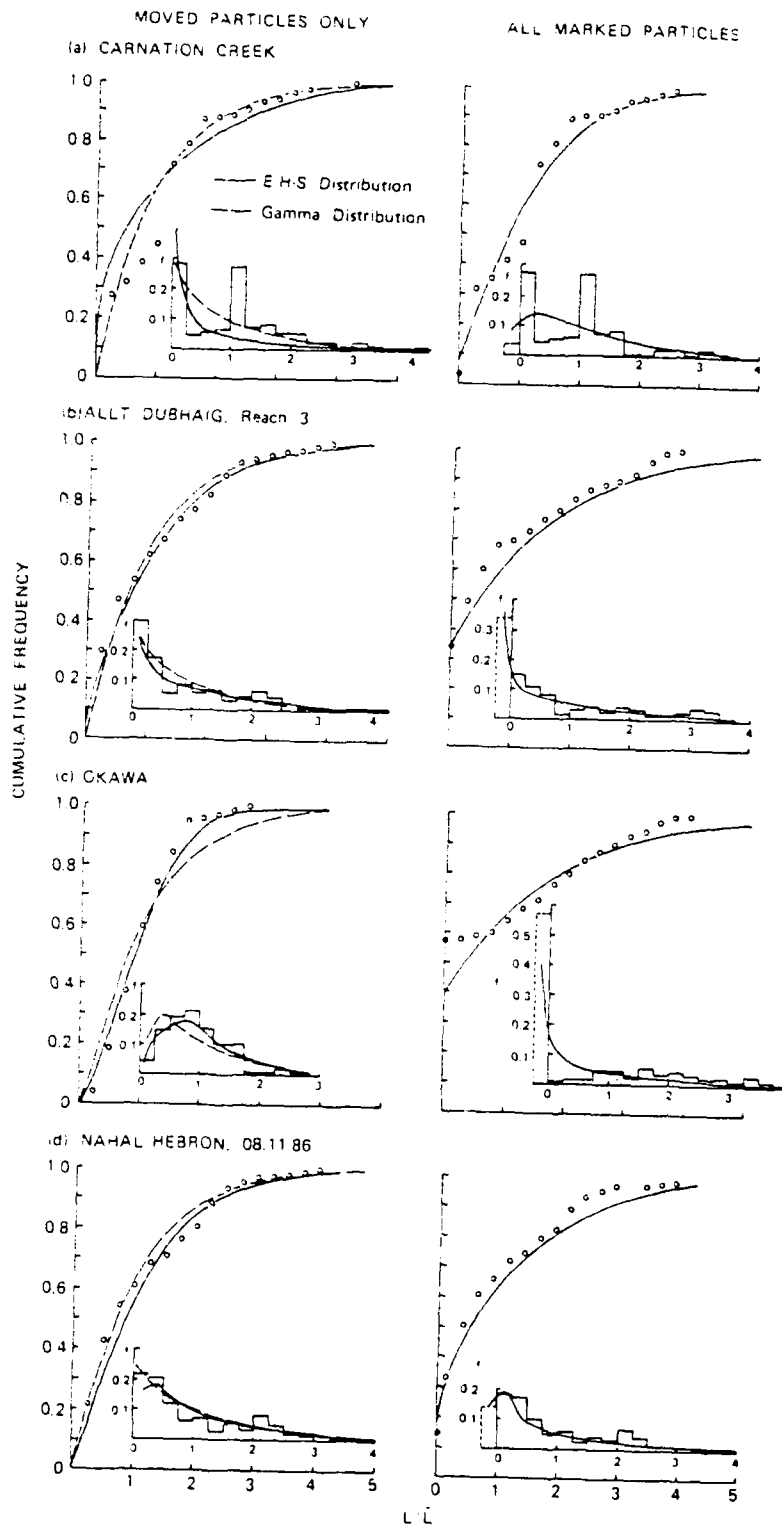
Fig. 3: Particle displacements of moved and of all
particles. The data were scaled by dividing the
distance of movement of each particle by the mean of
the event. Then the scaled data were grouped within
interval equal to 0.25 of the mean travel distance for
the examined event.

Fig. 5:

Relation between mean distance of movement and excess stream power estimated for the peak flow during flood events. Large symbols indicate particles moved and deposited by previous flows (i.e., in naturalistic positions before entrainment). The solid line is the best functional relation; the dashed line is an envelope curve for most of the data. For comparison, a relation for cobbles placed in sand bed streams is shown. The mean relation is affected by four additional points outside the limits of the diagram.

Fig. 4:

Relation between scaled mean distance of movement and scaled mean particle size. Distance is scaled by the mean distance of movement of the median size group of the surface material ($<L_{D50}>$). Grain size is scaled by the median size of the subsurface material. Data are grouped by half phi intervals. Groups containing less than 10 stones were excluded.

Workshop on Gravel-bed Rivers

Florence, 24-28 September 1990

MODELLING SEDIMENT TRANSPORT: DOMINANT FEATURES TO BE
SIMULATED IN DIFFERENT HYDROLOGICAL AND MORPHOLOGICAL
CIRCUMSTANCES

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1. Introduction

Sediment transport computations are generally needed to
simulate fluvial morphological processes, and in particular
to predict bottom aggradation and degradation.

Aggradation and degradation processes are sometimes
dramatic for gravel-bed rivers. As sediment transport in
mountain rivers is comparatively more intense and more
variable (both in space and in time) than in lowland
watercourses, not only morphological changes may be
extremely large, but the reaction time of these streams is
usually very quick.

In these circumstances, a correct evaluation of sediment
transport becomes crucial. On the other hand this evaluation
can be made by considering only the most significant
parameters that control the processes under investigation,
renouncing the description of less relevant aspects.

In this respect an important distinction should be made
according to the time- and space-resolution required by the
analysis. For example, an overall analysis at basin-scale
$(100 \div 1000 \text{ km}^2)$ needs the entire network of main streams

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(slope smaller than 5-10%) to be reproduced in the model, although with a rather coarse resolution (reaches having 200-500 m lenght). In this case, by averaging over the reach lenght, a number of simplifications can be introduced, both from the waterflow and from the sediment transport point of view.

By contrast, a detailed investigation on the silting of a small reservoir or on the sediment-trapping by a selective check-dam, requires a relatively small estension of the model (one kilometer or less), but with a space-resolution of few meters; in this case the hydrodynamic description should be much more detailed, as well as the sedimentological processes taken into consideration.

By the same token, as far as time-resolution is concerned, the analysis at basin scale mentioned before can be made either for a long period of years (e.g. to study the progressive aggradation and degradation of a torrent, respectively upstream and downstream of an artificial reservoir) or just for few hours (e.g. to study the sudden aggradation of a torrent overfed by landslides and debris flow during an exceptional storm). In the first case, by averaging the pulsations of waterflow and sediment input over the hydrological cycle, the annual sediment transport can be considered slowly variable from year to year; while in the second case the discharge and all the other parameters should be considered rapidly variable during the flood. Again, depending upon the required time-resolution (years or minutes), different simplifications may be brought into the model.

In any case, regardless the time- and space-scale of the processes to be reproduced, sediment transport of mountain rivers is <u>basically controlled by the non-uniformity of transported material and bottom material</u>.

A number of problems related to non-uniform grainsize distribution will be mentioned in the following sections. A four-layer conceptual model for the transport of a non-

uniform grainsize material at small space- and time-resolution will be subsequently presented. The possibility of simplifying the model will be finally discussed.

2. Composition of bed- and transported-material.

Intuition suggests that fine particles are more mobile than coarse particles, and indeed all the experimental formulae show that the transport rate T(d) of a uniform material increases if its grainsize, d, decreases.

What about if a certain grainsize class, di, belongs to a mixture? If one assumes that the movement of each class is not affected by the presence of the others, the transport rate of a given class should be equal to the product $\beta_i \cdot$ T(di) where T(di) is the transport rate of the same class computed as uniform material and β_i is the percentage of the same class present in the bottom. In this hypothesis, the granulometric composition of the transport results to be much finer than the granulometric composition of the bottom as the ratio

$$\beta_i T(di) / \sum (\beta_i T(di))$$

is much larger than β_i for finer diameters and much smaller than β_i for coarser diameter.

In a mixture, however, the mobility of different grainsizes is strongly equalized by the fact that finer particles are protected by the larger ones, and therefore are less subjected to the waterflow than in a uniform-size bed. To take into account this equalizing effect, different correction procedures ("hiding-and-exposure" coefficients) have been proposed (Egiazaroff, Day, Ranga-Raju, etc.). In the limit-case of equal mobility of all the grainsize classes, the composition of the river bottom is equal to the composition of the transported material.

According to some authors, an almost equal mobility is reached in many cases, especially in presence of relatively fast velocity and small grainsizes. It should be stressed, however, that there is no need to assume that the granulometry of the bottom and the granulometry of the feeding material must be necessarily the same when the equilibrium is reached. The two granulometries become equal only if all the grainsize classes have equal mobility, but more frequently this is not the case, as shown by several laboratory experiments, where the bottom composition appears to be much coarser than that of the transport. This difference is confirmed by the material forming the bed of mountain rivers , which almost invariabily is much coarser than the material of the landslides feeding the transport, as well as coarser than the material trapped in a reservoir intercepting the torrent.

3. Vertical distribution of bottom granulometry.

Another aspect related to the hiding-and-exposure effect mentioned in the previous section is the so called pavement. As defined by Parker (1982), pavement is the upper layer of the movable bed, which appears to be generally coarser than the underlaying bottom material (subpavement). Since the particles of the pavement are susceptible to movement (they are in fact transported by the waterflow), the notion of pavement is completely different from that of armouring which designates a completely immobile (at least below certain hydraulic critical conditions) protective layer.

Although the coarser particles of the pavement are not permanently at rest, still they exert a sort of shielding on the smaller particles of the subpavement. In this respect, the formation of pavement can be seen as a "vertical hiding", which seems to be much more effective than the "horizontal hiding" exerted by the coarse grains on the finer ones in their lee. In fact, this last mechanism can

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be probably neglected when compared to the shielding action exerted by the pavement.

If the "vertical hiding" is sufficiently strong, the granulometry of subpavement may be equal to the granulometry of the transported material (equal mobility conditions) although the pavement granulometry is coarser. This is what has been found by Parker in the Oak Creek. Even for the material of the subpavement, however, there is no need to be equal to the transported material. Recent experiments (Di Silvio e Brunelli, 1989) have indicated that, while the pavement appears to be coarser than the subpavement, the composition of the transported material is still much richer of fine particles than the last one.

4. Distinction between bedload and suspended transport.

Since suspended and bedload transport in equilibrium conditions depend on the same hydraulic and sedimentological parameters, it may be attractive to aggregate the two modes of transport into a "total" solid discharge, as proposed by many formulae.

A distinction between bedload and suspended transport, however, may be important in non-equilibrium conditions. As a matter of fact, the most relevant difference between the two modes of transport consists in the average travel lenght of individual particles: while the grains transported as bed load move each time by short steps proportional to their diameter, d, the grains transported in suspension travel over much longer distances proportional to Uh/w, that is decreasing with their size.

As the average travel lenght is obviously related to the distance required by the transport to reach the equilibrium with the local waterflow characteristics (adaptation distance), this distance results to be practically zero for bedload, while may be extremely long for fine particles transported in suspension.

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For a general purpose model, then, is preferible to keep separated the two modes of transport. We shall see later in which circumstances the aggregation is possible.

5. The four-layer conceptual model for the transport of non-uniform material.

For explaining all the features mentioned in the previous sections, both in equilibrium and non-equilibrium conditions, it is expedient to consider the cross section of a river subdivided in the following four layers (Fig.1):

(1) Waterstream, containing particles exclusively transported in suspension;

(2) Bottom layer, containing particles basically transported as bedload;

(3) Mixing-layer containing particles that are presently not in movement, but are liable to frequent vertical movements (entrainment to and settlement from the upper "transport" layers)

(4) Intrusion layer, containing particles liable to occasional vertical movements to and from the upper "transport" layers.

The particles below the intrusion layer are never disturbed, unless a general degradation of the bottom takes place.

The actual river bottom is represented by the instantaneous surface which separates moving particles and particles that, in any given instant of time, are at rest; a time-averaging operation over the bottom disturbances provides the average bottom surface, between transport layers (1 and 2) and storage layers (3 and 4).

Note that, by definition, the total volume concentration of sediments in the storage layers is equal to (1-n), being

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① Waterstream

$(h-a)$

instantaneous and
average bottom

② Bottom layer

a

③ Mixing layer
(PAVEMENT)

δ^P

④ Intrusion layer
(SUBPAVEMENT)

δ^S

Undisturbed
material
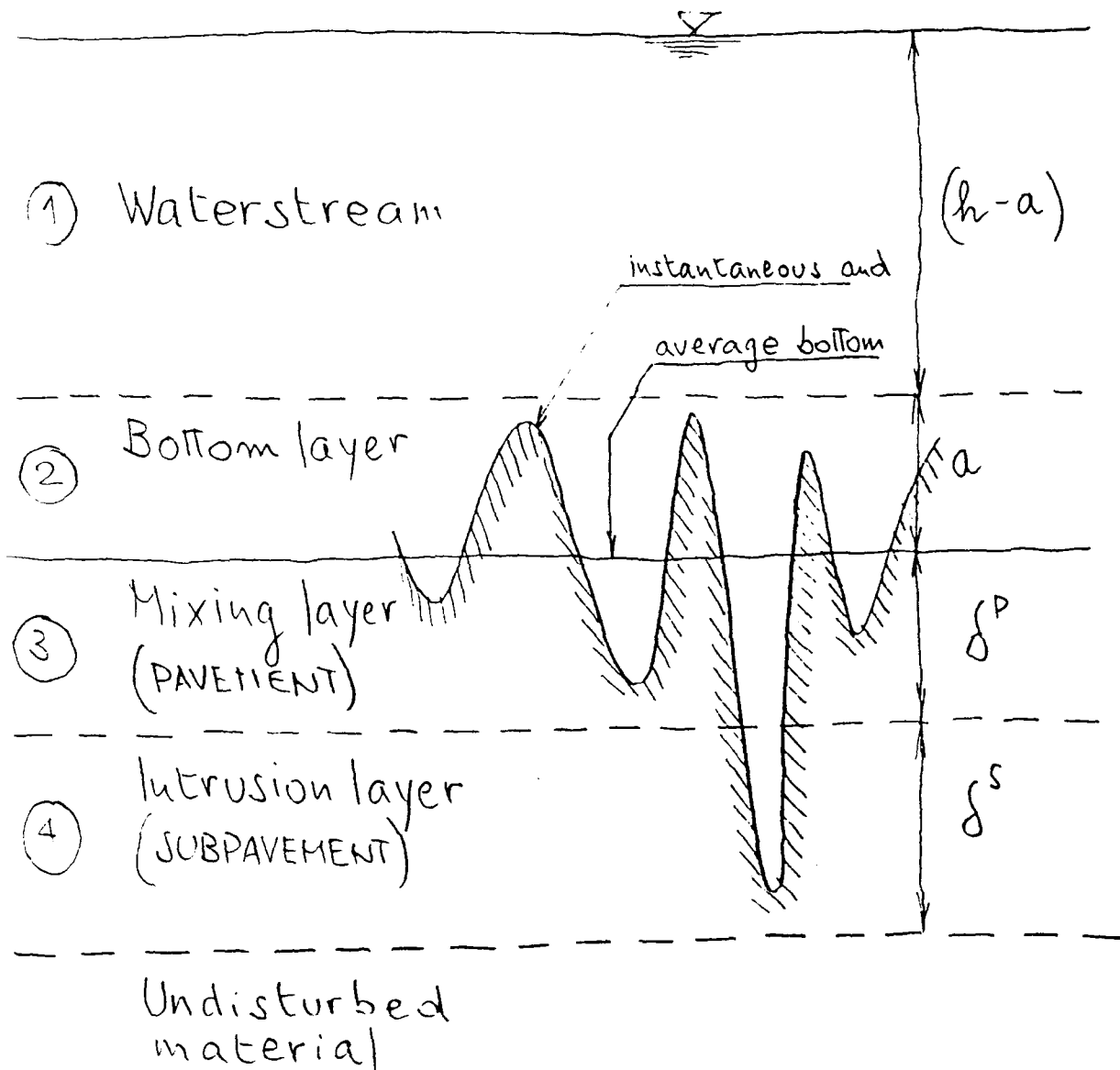
Fig. 1 - Four layer conceptual model for the transport of non-uniform material. Particles move in layers 1 and 2 (transport layers) and are temporarily at rest in layers 3 and 4 (storage layers).

n the porosity. In the transport layers, by contrast, even very close to the bottom the total concentration is a few orders of magnitude smaller (except in case of mass movements, e.g. debris-flow, that are not considered here).

The thickness of the bottom layer, a, and of the mixing layer, δ^P, are related to the average of bottom disturbances: namely to a representative height of the bedforms or to a representative grainsize for a flat bed.

The thickness δ^S of the intrusion layer, by contrast, is related to the deviation from the average of the same bottom disturbances. For a normal distribution of disturbances, the ratio δ^S/δ^P is a constant.

The four-layer model described here is quite similar to the three-layer model presented elsewhere (Armanini and Di Silvio, 1986). The fourth (intrusion) layer has been added later to explain a vertical flux below the mixing layer, observed in some experiments (Armanini and Di Silvio, 1989).

The existence of an intrusion layer was first considered by Ribberink (1987), who associated it with the less frequent deepest troughs of the dunes. As it will be seen in the next section, the same mechanism is also capable to explain the formation of a coarser mixing layer (pavement) above a finer intrusion layer (subpavement) in a gravel bed river, as shown in Fig. 3.

6. Longitudinal transport and vertical fluxes.

As said before, longitudinal movement can occur only in the transport layers. However, between the transport layers (1) and (2) and the storage layers (3) and (4), upward fluxes (pick-up) and downward fluxes (settlement) take continuously place.
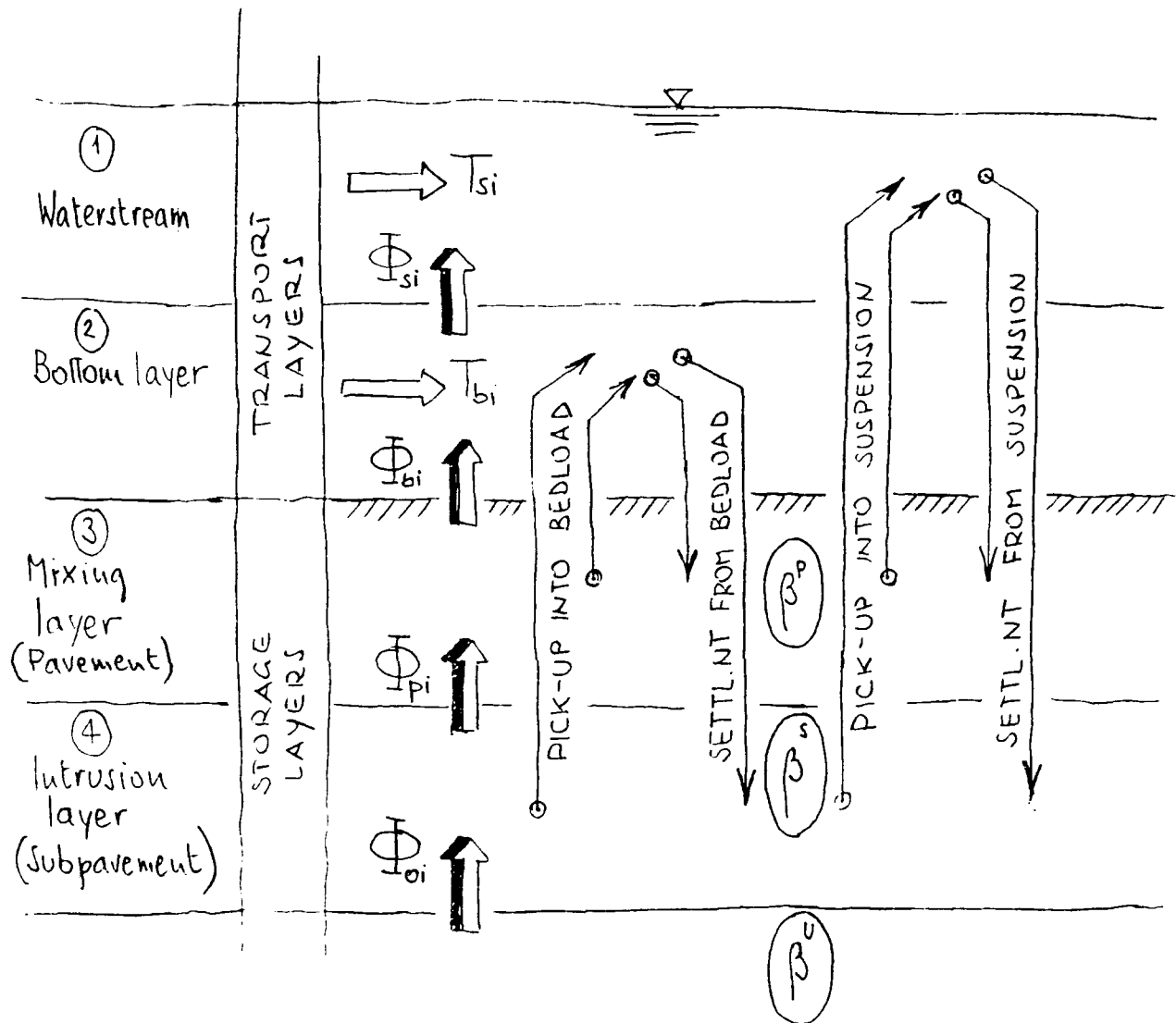
Fig. 2 — Transport and vertical fluxes of non-uniform sediments (i-th grainsize class) in non-equilibrium conditions

With reference to Fig. 2, the following balance
equations may be written for the i-th grainsize class:

$$\frac{\partial Tsi}{\partial x} + \frac{\partial C_{si}h}{\partial t} = \Phi_{si} \tag{1}$$

$$\frac{\partial Tbi}{\partial x} + \frac{\partial C_{bi}a}{\partial t} = \Phi_{bi} - \Phi_{si} \tag{2}$$

$$\frac{\partial \beta_i^P \delta^P}{\partial t} = \Phi_{pi} - \Phi_{bi} \tag{3}$$

$$\frac{\partial \beta_i^S \delta^S}{\partial t} = \Phi_{oi} - \Phi_{pi} \tag{4}$$

where Tsi and Tbi are respectively the suspended and
bedload transports; Csi and Cbi are the average
concentrations in the waterstream and in the bottom layer;
β_i^P and β_i^S are the percentages of the i-th grainsize class in

the mixing layer (pavements) and in the intrusion layer
(subpavement); $\Phi_{si}, \Phi_{bi} \Phi_{pi}$ and Φ_{oi} are the net vertical
fluxes between adjacent layers (pick-up minus settlement).
The concentrations Csi and Cbi in eqs. 1 and 2 can be
expressed in terms of Tsi and Tbi; in any case the storage
terms containing these quantities are negligible with
respect to the others.

Note that, in equilibrium conditions, all the net
vertical fluxes should be zero (pick-up is equal to
settlement). In a general situation, the net vertical fluxes
may be expressed in the following way:

$$\Phi_{si} = (\beta_i \, Tci - Ti)/L_i^* \tag{5}$$

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$$\Phi_{bi} = \Phi_{si} + \frac{\partial}{\partial x}(m_i \, \beta_i \, T_{ci}) \tag{6}$$

$$\Phi_{pi} = \left[(1-f_i) \, \beta_i^S \, T_{ci} - (1-\Sigma f_i) \, T_i\right] \frac{1}{L_i^*} \;\; +$$

$$+ \; \frac{\partial}{\partial x}\left[\frac{1-f_i}{1-\Sigma f_i} \, m_i \, \beta_i^S \, T_{ci}\right] - \frac{\partial Z}{\partial t} \, \beta_i^{S,P} \tag{7}$$

$$\Phi_{oi} = -\frac{\partial Z}{\partial t} \, \beta_i^{U,S} \tag{8}$$

where T_i is the actual total transport (bed and suspension) of the i-th grainsize class; T_{ci} is the total transport capacity of the same class, assumed as monogranular; $m_i = T_{bi}/T_{ci}$ is the percentage of T_{ci} moving as bedload; β_i^P, β_i^S, β_i^U are respectively the percentage of the i-th class in the mixing layer (pavement), in the intrusion layer (subpavement) and in the undisturbed sediment below the bottom; Z is the bottom elevation. In aggradation conditions $(\partial Z/\partial t > 0)$ it is $\beta^{S,P} \equiv \beta^P$ and $\beta^{U,S} \equiv \beta^S$, while in degradation conditions $(\partial Z/\partial t < 0)$ it is $\beta^{S,P} \equiv \beta^S$ and $\beta^{U,S} \equiv \beta^U$. The quantity L_i^*, representing the adaptation distance of the suspended transport for the i-th grainsize class, is given by

$$\frac{L_i^* \, w_i}{Uh} = \frac{a}{h} + \left(1-\frac{a}{h}\right) \exp\left[-1.5 \left(\frac{a}{h}\right)^{-1/6} \frac{w_i}{u^*}\right] \tag{9}$$
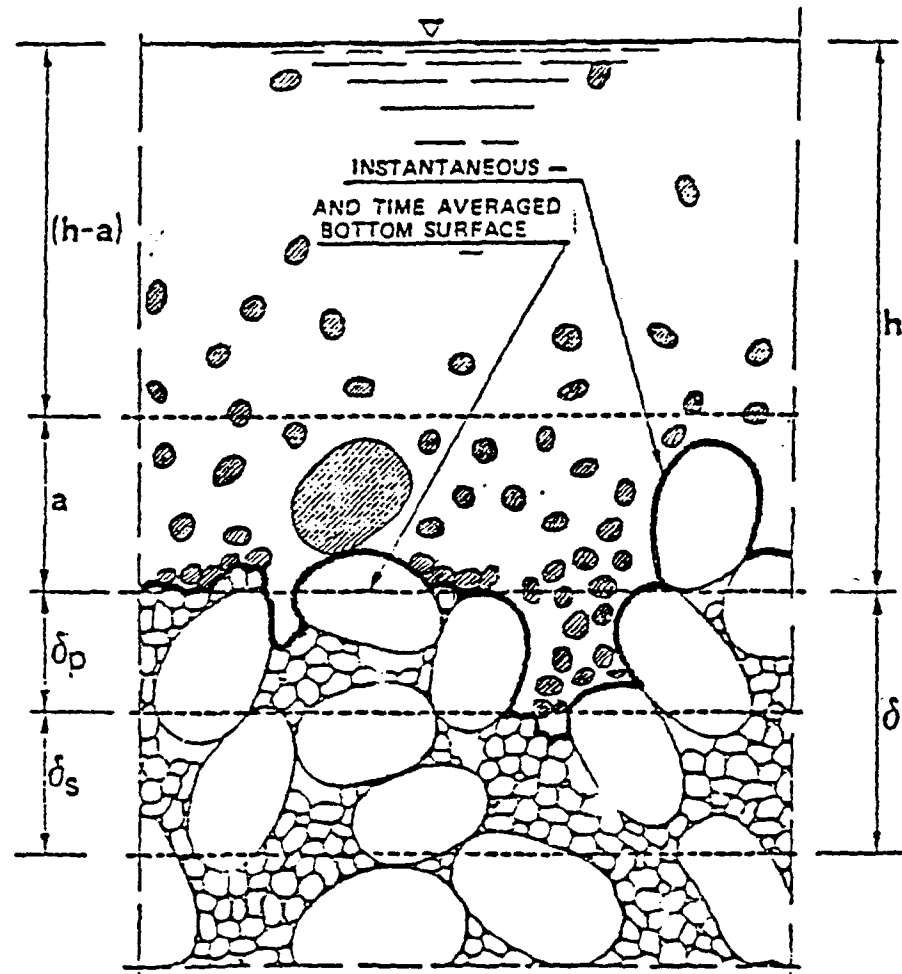
Fig. 3 — The four-layer model explains the formation of coarser pavement, δ^p, above a finer subpavement, δ^s, as a consequence of differential exchange between moving particles (dashed) and particles at rest (white).

where U is the flow velocity u* the friction velocity and and w_i the particle falling velocity for a diameter di.

The "equivalent" composition of the bottom is given by

$$\beta_i = f_i \, \beta_i^P + (1-f_i) \, \beta_i^P \qquad (10)$$

where f_i and $(1-f_i)$ respectively represent the relative importance of pavement and subpavement as far as the transport of the i-th grainsize class is concerned. The weight parameter f_i (see following section 7) is related to the different aptitude of different particle sizes transported by the stream, to exchange with the pavement rather than with the subpavement.

The total (bed and suspension) transport capacity Tci = Tbci+Tsci for a monogranular material and the ratio m_i = Tbci/Tci, for a given monogranular material of diameter di, are supposed to be known functions of hydrodynamics (for example, the van Rijn formulae). The weight parameter f_i, taking care of the bottom stratification, is also supposed to be a known function of the grainsize and hydrodynamic conditions (the definition of this function, based on experiments, is presently under way).

When appropriate initial and boundary conditions are respectively prescribed along the river (bottom composition) and at its upper end (sediment input), the solution of equations (1 to 8) provide the unknown quantities (Ti, β_i^P, β_i^S and Z) as a function of space and time. At each time-step, the (generally unsteady) waterflow along the river is to be computed according to the present values of bottom elevation and bottom roughness. For evaluating the bottom roughness in unsteady conditions, the adaptation of the bedforms may be as well taken into account (for example, by Nakagawa and Lamberti procedure).

7. Formation and evolution of pavement and subpavement.

The model described in the previous section is able to explain the existance of pavement and subpavement, as a consequence of the vertical exchange between the particles contained in these two layers and the particles transported by the waterflow. The vertical exchange is controlled by the wheight parameter f_i that appears in eqs. 5, 6, 7 and 10.

On the whole, as the large irregularities of the bottom surface are less frequent than the small ones, the exchange of the transported material is more active with the pavement than with the subpavement ($\Sigma \, f_i > 0.5$). However, f_i tends to be even larger for finer particles because of their aptitude to remain trapped in the lower layer; as a consequence, in equilibrium conditions, the pavement results to be richer of coarser particle than the subpavement. On the other hand, the composition of the transport results to be finer than the composition of the subpavement, as soon as the ratio (see eq.7):

$$\frac{Ti}{\Sigma Ti} = \frac{(1-f_i)\,\beta_i^S \, Tci}{\Sigma \, (1-f_i)\,\beta_i^S \, Tci} \tag{11}$$

is larger than β_i^S for fine diameters and smaller than β_i^S for coarse diameters.

Note that, should f_i being independent from the grainsize, the granulometry of subpavement would be equal to the one of pavement (no vertical hiding), and the mobility of any grainzise of the mixture would be the same as for uniform material. By contrast, should the product $(1-f_i)$ Tci being independent from the grainsize, the granulometry of subpavement would be equal to the one of transport (equal mobility), while the granulometry of the

pavement would be much coarser (very different vertical hiding).

The weight parameter f_i, in other words, would define the ratio $\left(\beta_i^S/\beta_i^P\right)$ in equilibrium conditions, as well as the efficiency of the pavement as a shield for sedimetns underneath; f_i is a function of grainsize diameter and flow characteristics and plays a role similar to the "hiding coefficient" mentioned in sections 2.

It is to be noted, however, that the shielding effect exerted by the pavement in equilibrium conditions may be different when net vertical fluxes are present (in principle, according to eqs. 3 and 4, each composition β_i^S and β_i^P evolves indipendently from the other). In this respect, the physical meaning of f_i seems to be more general than a simple hiding coefficient.

8. Possible simplifications of the model.

As mentioned in the introduction, the model can be substantially simplified (both for the waterflow and for the sediment transport) depending upon the space- and time-resolution required by the problem.

For morphological computations at basin scale (100-1000 km^2) river reaches of 100-500 meters may be taken as elementary space-steps (Δx) for a reasonably detailed description of the hydrographic network. Due to the irregular geometry of mountain rivers, the variation of water velocity from reach to reach is generally more relevant than the corresponding variations due to the river evolution. In these conditions, a quasi-uniform flow may be assumed in each river reach for any given water discharge. The local bottom slope may be assumed to be affected by the

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differential aggradation or degradation of adjacent reaches, or even be condidered constant in time.

Strong simplifications can also be brought in the sediment transport equations (1 to 8). As the space-step Δx is usually larger then the adaptation lenght L_i^* for all the grainsize classes, also suspended transport (besides bedload transport) can be considered immediately adapted to the local hydraulic conditions. In this hypothesis (Tsi = βi Tsci) there is no need to discriminate between the two modes of transport, so that layers (1) and (2) of Fig. 1 may be aggregated in one single layer and a total solid discharge (Ti = βi Tci) may be considered.

Another possible simplification of the model, for a mountain river subject to changes of its bottom elevation, is to consider the vertical flux Φ_{pi} between pavement and subpavement (eq.7) being negligible with respect to the other vertical fluxes. In this hypothesis, the ratio $\left(\beta_i^P/\beta_i^S\right)$ between the compositions of the two layers is supposed to be the same as in equilibrium conditions; there is no need to distinguish between layers (3) and (4), provided that the shielding effect of pavement is taken into account by a proper "hiding" coefficient.

By aggregating waterstream and bottom layer into one transport layer and by aggregating pavement and subpavement into one bottom layer, the traditional two-layer model without suspension space-lag, initially proposed by Hirano (1971), is obtained. It is to be noted, however, that these approximations are not acceptable anymore, either when the space-lag of suspended transport is comparable with the space resolution required by the model (e.g., for studying the silting of a reservoir), or when the flux between pavement and subpavement cannot be neglected (e.g. to study the intrusion of polluted sediments under the river bed).

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A different type of simplification can be introduced in the model when <u>long-term morphological changes</u> are to be simulated (e.g., to predict the effects of a new construction in the river over the period of decades). In this case, equations (1 to 8) are integrated over one or more seasonal cycles and the long-term sediment conservation equations are obtained. In the averaging operation several residue terms, due to the non-linearity of the equations, are neglected; this approsimation, however, is not critical for most of practical cases.

Examples of application of the simplified two-layer model, both for long-term and short-term simulations, are given by Di Silvio and Peviani (1989).

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NUMERICAL MODELING OF GRAVEL MOVEMENT IN CONCRETE CHANNELS

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INTRODUCTION

Corte Madera Creek drains an area of approximately 28 square miles in Marin County, California. The creek discharges into San Francisco Bay about 9 miles north of the Golden Gate. In 1972 the US Army Corps of Engineers completed three units of a proposed four-unit project on Corte Madera Creek extending from San Francisco Bay through the cities of Corte Madera, Larkspur, Kentfield, and Ross, a distance of about 4.5 miles (Figure 1). The first 2.9 miles of the project consist of an earth channel. The next mile of the project consists of a 33-ft-wide concrete channel with a stilling basin at the downstream end. The first 1,000 ft of the channel has a mild slope of 0.0007. The remainder of the concrete channel has a steep slope of 0.0038, and is designed for supercritical flow. These segments have been constructed. The final unit, Unit 4, of the project was to be an additional 3,000 ft of concrete channel. Construction of Unit 4 was delayed due to litigation, environmental concerns, and strong public opposition.

The existing channel in the Unit 4 reach has a capacity of about 3,000 cfs. Flows greater than this overtop the bank and flood adjacent businesses and residences. Additional flooding problems occur because channel capacity in the downstream reaches of the concrete channel is reduced due to the accumulation of sands and gravels. This accumulation reduces conveyance and increases the composite channel roughness.

The largest recorded flood flow on Corte Madera Creek occurred in January 1982. This flood had an estimated peak of 7,200 cfs at the Ross gage and a recurrence frequency greater than 100 years. Another flood occurred in March 1983, with an estimated peak of 3,480 cfs and a recurrence interval of about 6 years This event was the third largest flood of record. Both of these floods resulted in damages to homes and businesses adjacent to Corte Madera Creek.

In December 1983, the Corps of Engineers was requested to reinitiate the project. After extensive local coordination, engineering analysis of the data, and experience obtained from the recent floods, a new plan for Unit 4 was developed by the US Army Engineer District (USAED), Sacramento. This plan consisted of channel improvements, floodwalls, and a sediment trap.

A numerical model study was performed to evaluate the deposition pattern in the concrete channel and to determine the effectiveness of the sediment trap in reducing or eliminating the deposition problem. The effect of the
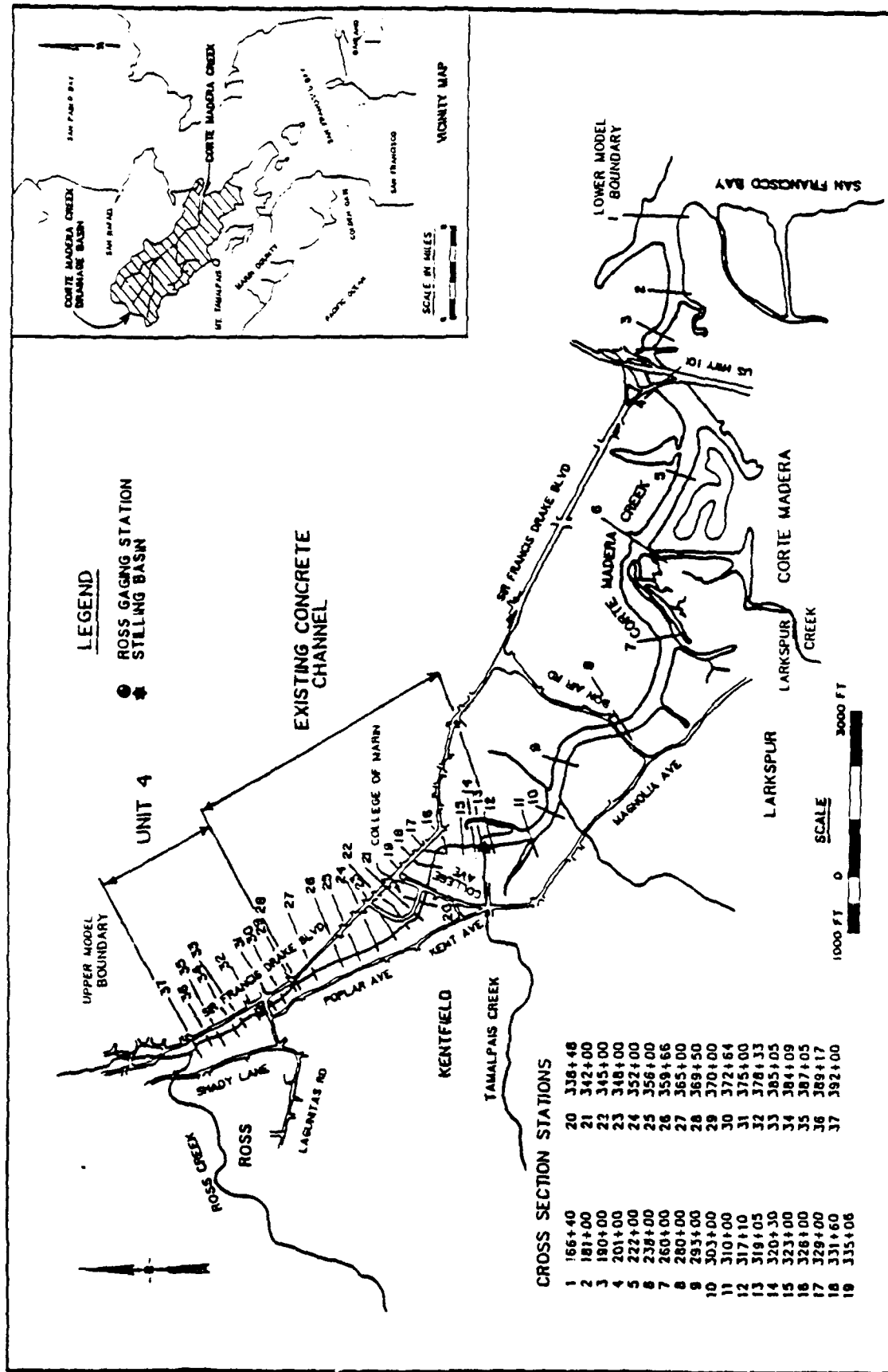
Figure 1. Cross section locations

CROSS SECTION STATIONS

1	166+40	20	338+48
2	181+00	21	342+00
3	190+00	22	345+00
4	201+00	23	348+00
5	222+00	24	352+00
6	238+00	25	356+00
7	260+00	26	359+66
8	280+00	27	365+00
9	293+00	28	369+50
10	303+00	29	370+00
11	310+00	30	372+64
12	317+10	31	375+00
13	319+05	32	378+33
14	320+30	33	385+05
15	323+00	34	384+09
16	328+00	35	387+05
17	329+00	36	389+17
18	331+60	37	392+00
19	335+08		

accumulated sediment on channel roughness was evaluated. Critical to this study was the adequacy of channel wall heights in the existing concrete channel.

THE MODEL

Description

The TABS-1 one-dimensional sedimentation program was used to develop the numerical model for this study. Development of this computer program was initiated by Mr. William Thomas at the US Army Engineer District, Little Rock, in 1967. Further development at the US Army Engineer Hydrologic Engineering Center (USAEHEC) by Mr. Thomas produced the widely used HEC-6 generalized computer program for calculating scour and deposition in rivers and reservoirs (USAEHEC 1977). Additional modification and enhancement to the basic program by Mr. Thomas at the US Army Engineer Waterways Experiment Station (WES) led to the TABS-1 program currently in use (Thomas 1980, 1982). The program produces a one-dimensional model that simulates the response of the riverbed profile to sediment inflow, bed material gradation, and hydraulic parameters. The model simulates a series of steady-state discharge events and their effects on the sediment transport capacity at cross sections and the resulting degradation or aggradation. The program calculates hydraulic parameters using a standard-step backwater method assuming subcritical flow. The program assigns critical depth for water-surface elevation if the backwater calculations indicate transitions to supercritical flow. However, for supercritical flow, hydraulic parameters for sediment transport are calculated assuming normal depth in the channel.

For numerical sedimentation models to completely simulate the behavior of a stream channel, computations would have to account for all of the basic processes of sedimentation: erosion, entrainment, transportation, deposition, and compaction of both the bed and the streambanks for the complete range of particle sizes found in nature. The state of the art has not yet advanced to such a complete simulation. The computer program used in this study, TABS-1, is a state-of-the-art program for use in mobile bed channels. It is designed to calculate aggradation and degradation of the streambed profile. When applied by experts using good engineering judgement, the TABS-1 program will provide good insight into the behavior of mobile bed channels such as Corte Madera Creek.

Particle sizes from sand to gravel are involved in Corte Madera Creek, which complicates the simulation because particle size controls the fundamental processes in river sediment behavior. The time scale of interest is from a single flood event to the life of the project. The long-term trends can be evaluated from a statistical analysis of the gage records, but a great deal of variation in water and sediment runoff occurs from one storm event to the next because of the stochastic nature of the hydrologic cycle. The approach for bridging these gaps is to formulate (a) a procedure that includes the statistical nature of the boundary conditions – the uncertainty in forecasting future hydrology and sediment yield is probably more significant than gaps in modeling the physics of the mobile boundary processes so far as the accuracy of results is concerned; and (b) a computer program that emulates the physical processes in the project reach sufficiently well to quantify how the

sedimentation processes will respond to changes in the boundary conditions and/or to changes in the project geometry or roughness.

Although the sedimentation processes are complex, procedures for describing most of them have been published. The TABS-1 computer program includes those procedures. Where gaps exist between the available procedures, TABS-1 contains logic that bridges those gaps. In summary, it is state-of-the-art technology for calculating the aggradation and degradation in mobile bed channels, and because it has given reliable results at similar projects, it is expected to give reliable answers to the questions being addressed here.

Channel Geometry

The numerical model extends from sta 166+00 near the mouth of Corte Madera Creek at San Francisco Bay to sta 392+00, which is just downstream from the confluence of Ross Creek. Reach lengths in this model were generally about 2,000 ft downstream from the stilling basin and about 300 ft upstream. Cross-section locations are shown in Figure 1.

Hydrographs

Discharge hydrographs are simulated in the numerical model by a series of steady-state events. The duration of each event is chosen such that changes in bed elevation due to deposition or scour do not significantly change the hydraulic parameters during that event. At relatively high discharges, durations need to be short: time intervals as low as 1 hour were used for Corte Madera Creek. At low discharges, the time interval may be extended. Time intervals up to 3 days were used in this study. The hydrograph used in the adjustment and circumstantiation of the study was based on historical data from the US Geological Survey's (USGS) gage on Corte Madera Creek at Ross. The effect of breakout flows were incorporated into the model.

Bed Material

The numerical model requires that an initial volume and gradation be specified for the bed material. This was accomplished in the concrete channel reach by specifying a bed material depth of zero for the initial condition. The gradation of deposited material is then calculated by the model's sorting and armoring algorithm. The bed material gradation in the natural channel, upstream from the concrete channel, was determined from the average of four samples collected in January 1985 by the Sacramento District (Figure 2). These samples were taken in the vicinity of the Ross gage. This gradation was used to determine equilibrium sediment transport capacity at the upstream boundary of the numerical model.

Channel Roughness

Hydraulic roughness is influenced by grain size or bottom roughness, bank or sidewall roughness, bed form, water depth, changes in channel shape, and changes in flow direction or distribution due to bends and confluences. In the one-dimensional numerical model these effects are accounted for by the Manning's roughness coefficient. Acceleration and deceleration of flow are accounted for with expansion and contraction coefficients. The roughness coefficient may vary significantly with discharge and time. The influence of

where

R'b − hydraulic radius of the bed attributed to grain roughness

D_{84} − particle size of which 84 percent of the bed is finer

In order for sand and gravel to influence the bottom roughness, a minimum bed thickness was required in the model. If this requirement was not met, the model assigned a D_{84} value that is representative of concrete. The minimum thickness was the larger of 4 mm or two times the grain size for which the percent coarser fraction covered the bed to a thickness of two grain diameters.

The Limerinos equation accounts primarily for the grain roughness. Additional bed roughness can be caused by the form roughness that occurs with a movable bed. Calculations using the Brownlie equation (Brownlie 1983) for upper regime flow showed an increase in the Manning's roughness coefficient of 0.010 due to form roughness. In the numerical model, the bed roughness coefficient was increased to account for form roughness if both the minimum bed thickness and the critical shear stress were exceeded. The Shields equation was used to determine critical shear stress.

$$\tau_c = \theta (\gamma_s - \gamma_w) D_{50} \qquad (3)$$

where

τ_c − critical shear stress

θ − Shields parameter

γ_s − specific weight of sediment

γ_w − specific weight of water

Various investigations have established a range for the Shields parameter between 0.03 and 0.06 when median grain diameter is used in the equation. In order to provide a continuous transition for the increase in roughness coefficient for form roughness, the following procedure was adopted. If the calculated shear stress was less than critical shear stress using a Shields parameter of 0.03, then the bed was assumed to be immobile and no adjustment was made to the Limerinos bed roughness. If the calculated shear stress was greater than the critical shear stress using a Shields parameter of 0.06, then the Limerinos bed roughness was increased by 0.010 to account for form roughness due to the movable bed. The roughness increase was linearly interpolated for conditions between these limits.

The sidewall roughness in the composite roughness equation accounted for the roughness height of the wall itself. An appropriate roughness coefficient was determined by simulating high-water marks from the January 1982 flood. The wall roughness should be somewhat higher than for finished concrete because of tubeworm and barnacle deposits. A value of 0.018 was chosen.

Initially, calculated water-surface elevations in the lower portion of
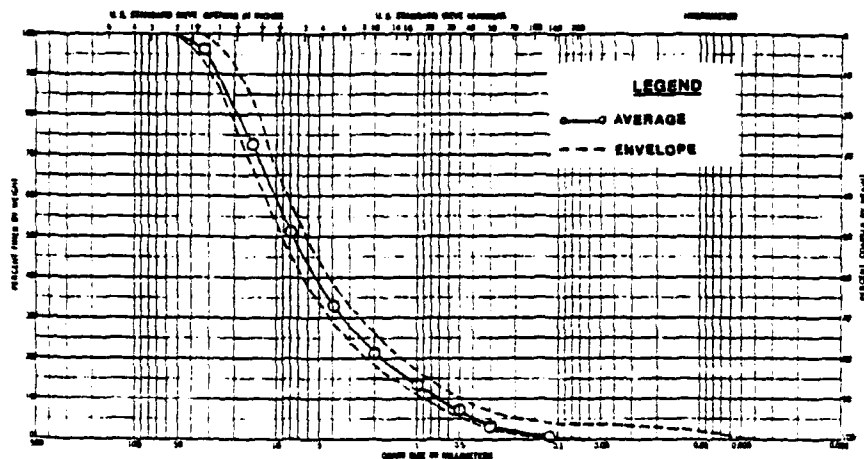
Figure 2. Bed material gradations, vicinity of Ross gage

grain or bottom roughness on hydraulic losses is known to decrease with increases in depth. Resistance due to bed forms can decline dramatically when the bed forms are washed out and replaced by a plane bed. An attempt to account for these problems was made in this study by developing an algorithm that calculated composite roughness coefficients based on roughness attributed to the bed, sidewalls, and bed movement.

Determining composite roughness in the concrete channel is complicated by the accumulation of sand and gravel on the channel bottom. A composite roughness coefficient was calculated using the following formula:

$$n = \left[\frac{P_w\, n_w^{1.5} + P_b\, n_b^{1.5}}{P_w + P_b} \right]^{2/3} \tag{1}$$

where

P_w — perimeter of the wall

n_w — roughness coefficient of the wall

P_b — perimeter of the bed

n_b — roughness of the bed

The bed grain roughness was calculated using the Limerinos equation (Limerinos 1970):

$$n_b = \frac{0.09256 (R_b')^{1/6}}{1.16 + 2.03 \log \left[\dfrac{R_b'}{D_{84}} \right]} \tag{2}$$

the concrete channel were significantly lower than the reported high-water marks for the January 1982 flood. To correct this discrepancy, the roughness was increased by 0.004 to account for losses due to channel bends in the downstream reach, below sta 342+00. With this revision, calculated water-surface elevations and high-water marks were much closer, but the calculated values were still slightly lower than high-water marks. These water-surface comparisons are shown in Figure 3. Differences in reported and calculated values for
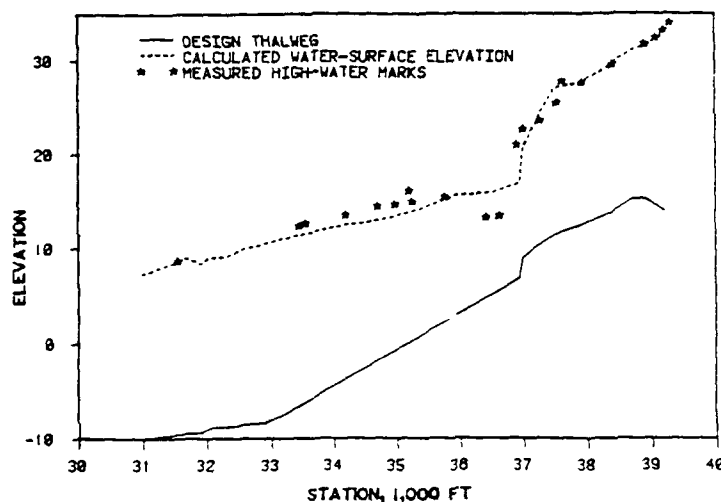
Figure 3.   Maximum water-surface elevations, 4 January 1982

the high-water marks taken along the channel may be attributed primarily to losses at bridges at College Avenue and College of Marin. The assigned roughness coefficients were within the upper range used in engineering practice for this type channel, and therefore, further increases in roughness coefficients were deemed unreasonable.

Sediment Inflow

Measured sediment inflow data for Corte Madera Creek are inadequate to determine a reliable sediment inflow rating curve for the entire range of discharges considered in this study. USGS collected suspended load samples at the Ross gage during water years 1978-1980. The samples were analyzed to determine particle size distributions. Twenty-three events were reported with water discharges varying between 47 and 1,560 cfs. The highest measured flow was well below the design flood peak of 6,900 cfs. In addition to the suspended load measurements, seven bed-load samples were collected during water year 1978 at the Ross gage. Particle size distributions of these samples were also reported. Water discharges when the samples were collected varied between 47 and 1,180 cfs. The measured data were insufficient to determine a reliable sediment inflow rating curve for the model.

An initial sediment inflow rating curve was developed based on an optical fit of measured data for water discharges less than 2,000 cfs and on equilibrium transport for higher discharges. Equilibrium transport was calculated using the SEDTRAN transport function (described in the section on transport function) and a bed material gradation determined from four samples collected in 1985 in the vicinity of the Ross gage.

The sediment inflow rating curves were adjusted during the adjustment phase of the study. In general, the concentrations of sands were increased, but the concentrations of coarse gravels were decreased. Sediment inflow within the range of sampled data was generally unchanged. Calculated transport of coarse and very coarse gravel did not agree with sampled transport; therefore, sediment inflow concentrations for these coarsest particles were based on equilibrium calculations in the numerical model. The justification for the adapted sediment inflow rating curves is based on successful simulation of measured deposition in the concrete channel as discussed in the section on "Adjustment and Circumstantiation."

Transport Function

A modification of the Laursen equation (Laursen 1958) was developed for use in this study. The Laursen function is desirable because it was developed for size class analysis and considers parameters essential to both bed and suspended sediment loads. The modified Laursen equation (labeled SEDTRAN) incorporated data for transport of gravels in addition to the sand data used to develop the original Laursen function.

The SEDTRAN function calculates the hydraulic radius due to grain roughness using the Limerinos equation. This value (instead of the depth as proposed by Laursen) is then used to calculate the grain shear stress:

$$\tau'_o = \frac{\rho V^2}{58} \left[\frac{D_{50}}{R'_b} \right]^{1/3} \tag{4}$$

where

τ'_o = grain shear stress

ρ = water density

V = water velocity

This equation is dimensionally homogeneous and can be applied with any consistent set of units.

Bed-load transport is a function of the ratio of applied to critical shear stress. In the SEDTRAN function this is express by the parameter

$$(\tau'_o/\tau_{ci}) - 1 \tag{5}$$

where τ_{ci} is the critical shear stress for the ith grain size. The critical shear stress varies with particle size, larger particles having greater critical shear stress. Paintal (1971) determined that the critical shear stress, as used in Equation 5 to determine sediment transport, also varied with applied shear stress. When the dimensionless shear stress ($\tau_*^{'}$) was less than 0.05, he found that the critical shear stress decreased significantly:

$$\tau_o^* = \frac{\tau_o}{(\gamma_s - \gamma_w)D_i}$$

(6)

where

τ_o — applied shear stress

D_i — geometric mean diameter of the ith size class

This variation in critical shear stress is accounted for in SEDTRAN by varying the Shields parameter between 0.039 and 0.020. The higher value, recommended by Laursen (1958), was used when τ_o^* was greater than 0.05. The lower limit was determined by Andrews (1983). The effect of this change is that initiation of motion for coarser particles occurs at lower shear stresses, and the transport potential of coarser particles is increased.

The SEDTRAN function uses the ratio of grain shear velocity (instead of total shear velocity) to grain fall velocity as the important parameter influencing suspended sediment transport. A functional relationship between this ratio and other parameters was determined by Laursen (1958) based on river and flume data. Due to the reformulation of Laursen's parameters, a new functional relationship was developed for SEDTRAN. The relationship is based on data from both rivers and flumes. The functional relationship and data scatter are shown in Figure 4. Flume data gathered under more controlled conditions have significantly less scatter than the river data.
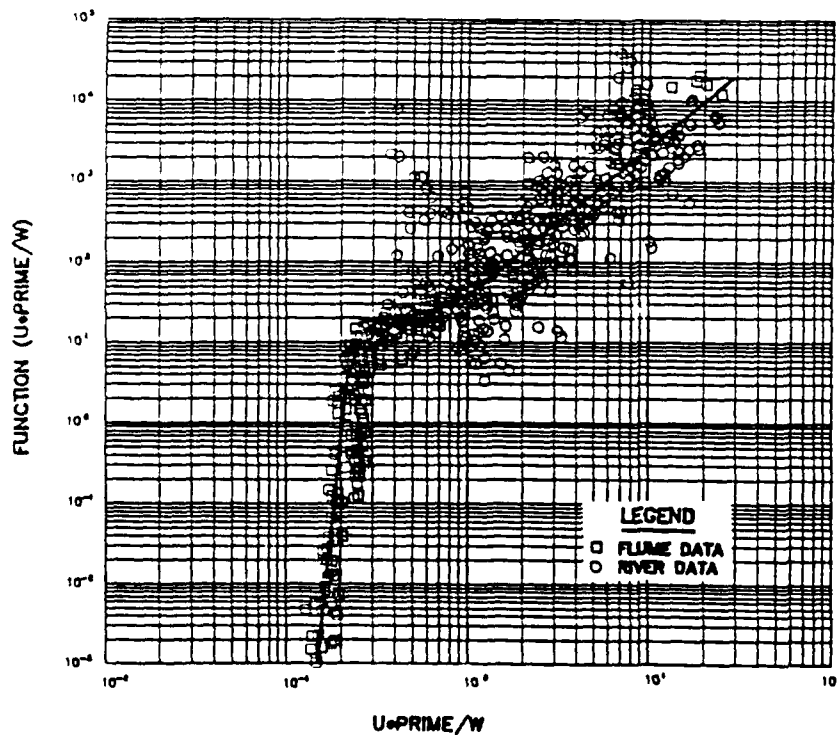
Figure 4. Function (U*Prime/W) for sediment transport function

Sediment concentration is calculated by SEDTRAN using the following formula:

$$C = 0.01\gamma_w \sum_i P_i \left[\frac{D_i}{y}\right]^{7/6} \left[\frac{\tau'_o}{\tau_{ci}} - 1\right] f\left[\frac{U'_*}{\omega_i}\right] \qquad (7)$$

where

C — concentration in weight per unit volume

P_i — fraction of grain size class in the bed

y — water depth

τ'_o — bed grain shear stress

U'_* — grain shear velocity

ω_i — fall velocity

$f\left[\dfrac{U'_*}{\omega_i}\right]$ — function defined in Figure 4

This function is considered to be a refinement to Laursen's original equation and is based on a wider range of physical data. The primary benefit is that it moves coarser gravels better than other functions. However, for the coarsest gravels (greater than 16 mm) the function still does not transport as much material as was sampled in Corte Madera Creek.

MODEL ADJUSTMENT AND CIRCUMSTANTIATION

Adjustment to Deposition Surveys

The historical hydrograph between October 1972 and July 1982 was simulated with the numerical model. This represents the time period between completion of the existing concrete channel and the first available survey of channel deposition. The survey was completed the summer after the flood of record on Corte Madera Creek. Average depths were determined from the surveyed deposition profile in the V-shaped bottom portion of the channel and an average-depth profile was developed for comparison with calculated profiles from the numerical model. Surveyed and calculated deposition profiles are compared in Figure 5. The numerical model overestimated deposition in the mild-sloped reach of the concrete channel, but reproduced both the depth and longitudinal extent of deposition in the steep-sloped portion of the channel.

The hydrograph between October 1972 and August 1984 was simulated with the numerical model. Another deposition survey was taken and bed material samples were collected in August 1984. The time period between the first
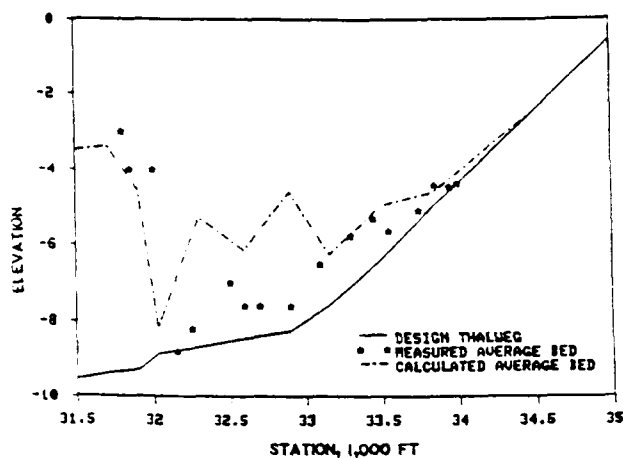
Figure 5. Aggradation in concrete
channel, October 1972 to April 1982
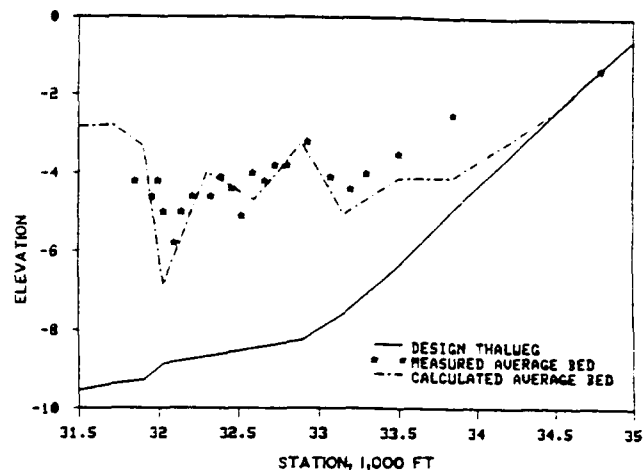
Figure 6. Aggradation in concrete
channel, October 197? to August 1984

survey, which was taken in July 1982, and August 1984 was a relatively high
runoff period. The runoff hydrograph at the Ross gage peaked at 2,690 cfs.
The calculated and surveyed deposition profiles are compared in Figure 6.
Themodel satisfactorily simulated deposition in the mild-sloped portion of the
channel, but underestimated deposition in the steep-sloped portion of the
channel. During this 2-year period, both the calculated and surveyed profiles
showed that the depth of deposition increased in the concrete channel. Bed
material samples were collected at sta 335+06, which is the College Avenue
Bridge, and at sta 341+00, which is 250 ft upstream of the College of Marin
Bridge. Calculated gradations at model cross sections in this reach were
compared to the average gradation of these two samples. The sample gradations
were found to be considerably finer than the calculated gradations. It was
determined from sensitivity studies (discussed in a subsequent section) that
an increase in sediment inflow would result in an increase in the coarseness
of the calculated gradation and that a decrease in sediment inflow would
result in a finer calculated gradation. In order to improve the comparison
between measured and calculated gradations, the sediment inflow into the model
was adjusted. Gravel inflow was decreased and inflow of fine sands increased.
The adjusted sediment inflow rating curves improved the comparison between
sampled and calculated bed material gradations, but the calculated gradation
was still coarser. Sampled and calculated bed material gradations for August
1984 are shown in Figures 7a and 7b, respectively. Further reduction in the
gravel inflow was not deemed appropriate because it would deviate too much
from the measured sediment inflow data and because it would reduce the
correlation between measured and calculated depths of deposition.

| a.   Sampled | b.   Calculated |
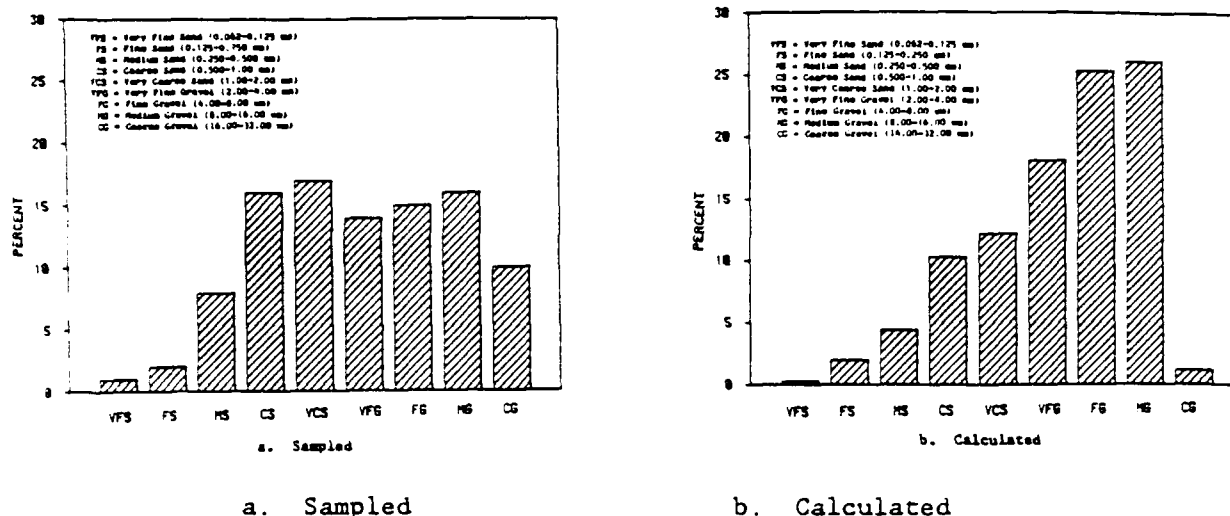
Figure 7. Bed material gradations in the vicinity
of College Avenue, August 1984

Model Circumstantiation

At this point in the study the numerical model was considered to be adequately adjusted for the prediction of general deposition patterns in Corte Madera Creek. Circumstantiation of the model was accomplished by continuing the historical simulation to January 1986, when another deposition survey was taken. The period between August 1984 and January 1986 had relatively high runoff with the largest peak being 2,600 cfs. Surveyed and calculated deposition profiles and accumulated deposition are compared in Figure 8. The January 1986 deposition profile (Figure 8a) shows little change from the August 1984 deposition profile (Figure 6). The model continued to reproduce an accurate profile downstream of the College Avenue Bridge, but underestimated deposition upstream. The model was very successful in predicting the quantity of total deposition in the channel.

A significant runoff event occurred in February 1986, when an estimated peak discharge of 4,150 cfs occurred at the Ross gage. Bed material samples were collected at three locations in the concrete channel upstream from the College of Marin Bridge in March 1986. A deposition survey in the concrete channel was taken in May 1986. Bed material samples were collected at 15 locations between sta 326+00 and 337+50 in May 1987, and three samples were taken laterally across the channel at sta 337+00 in September 1987. The period between May 1986 and May and September 1987 had relatively low runoff, with a maximum peak discharge of 2,500 cfs and only 5 days when the average daily flow exceeded 100 cfs. Due to the small amount of runoff between May 1986 and the collection of the bed material samples in 1987, it was deemed reasonable to compare these measurements with calculated gradations.

Calculated and surveyed deposition profiles and accumulated deposition for the October 1972 - May 1986 simulation are compared in Figure 9. Based on field surveys, 2,800 cu yd of material were removed from the concrete channel during the February 1986 flood. This compares with a removal of about
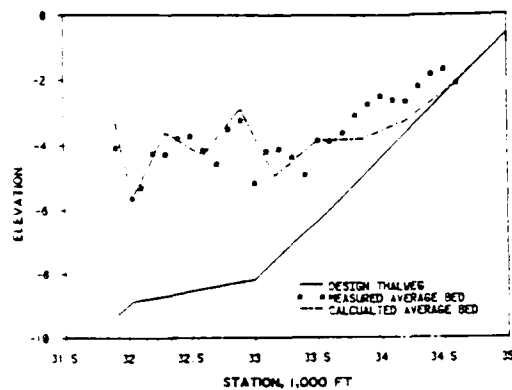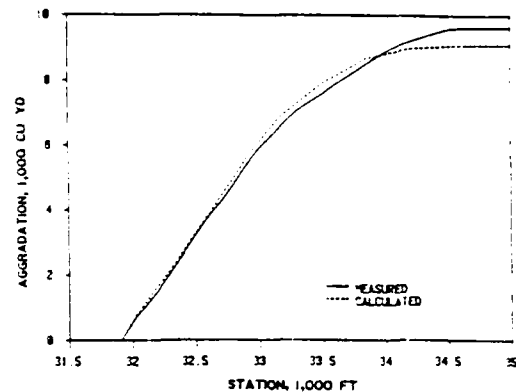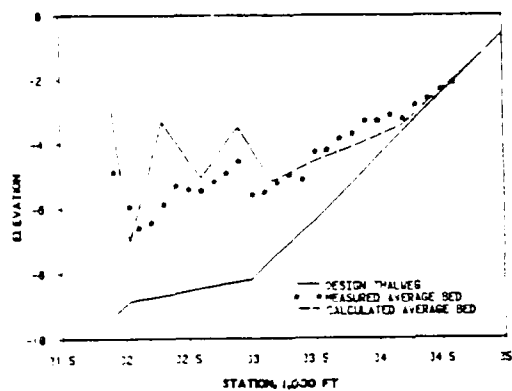
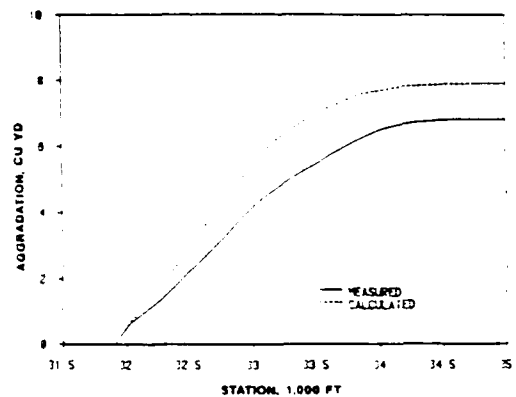a. Profile   b. Accumulated

Figure 8. Aggradation in concrete channel, October 1972 to January 1986

a. Profile   b. Accumulated

Figure 9. Aggradation in concrete channel, October 1972 to May 1986

1,200 cu yd calculated in the model. The result is consistent with the com-
parison of field surveys and calculated deposition profiles after the flood of
January 1982, which also showed more material removed in the prototype. These
results indicate that the model, using average sediment inflow rating curves,
underestimated the ability of the concrete channel deposits to degrade during
floods. This behavior in the model may be attributed to one or more of the
following factors:

> a. The sensitivity study demonstrated the importance of sediment
>   inflow concentrations on the resultant deposition. It is pos-
>   sible that sediment inflow concentrations during the two flood
>   events were significantly different from the long-term aver-
>   ages developed from the measured data. The measured sediment
>   inflow data were taken between 1978 and 1980, which were
>   fairly normal runoff years with a maximum peak discharge of
>   2,910 cfs.
>
> b. Model roughness in the concrete channel may be too high. The

high roughness values used in the model were chosen in order
to match high-water marks from the January 1982 flood. These
high roughness values reduce the velocity and thus the scour-
ing potential of the flow.

b. Model roughness in the concrete channel may be too high. The
high roughness values used in the model were chosen in order
to match high-water marks from the January 1982 flood. These
high roughness values reduce the velocity and thus the scour-
ing potential of the flow.

c. Flow breaks away from the Corte Madera channel upstream from
the concrete channel during flood flows. If the flow remain-
ing in the channel is underestimated, then the model will show
a reduced scouring potential.

Bed material gradations from samples taken in 1934-1987 are compared in
Figure 10. Lateral and longitudinal variations in bed material gradations are
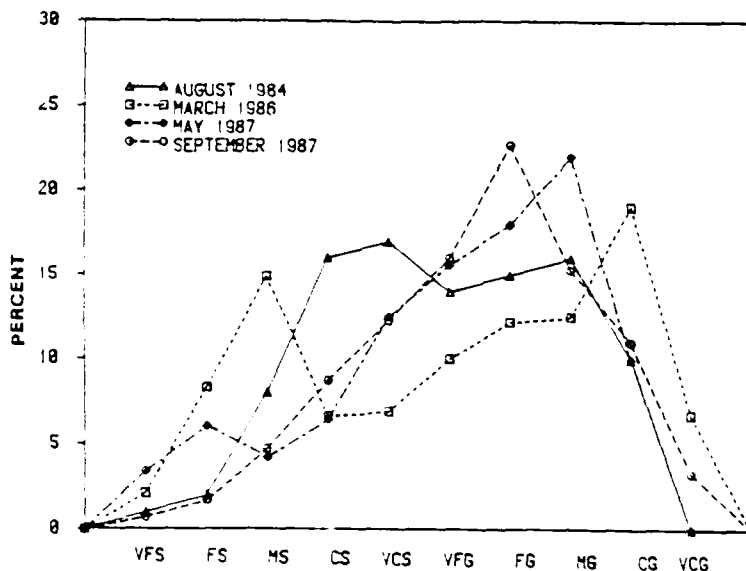
Figure 10. Sampled bed material gradations

not unusual in gravel bed streams. Therefore. average bed material gradations
determined from a large sample population will be more reliable. The March
1986 gradation is finer than the other gradations. These data came from three
samples between sta 340+50 and 321+00. One sample compared favorably with
1987 data; the other two had bimodal distributions. Ten samples between sta
326+00 and 337+50 were used to obtain the May 1987 gradation. Three samples,
varied laterally at sta 336+80, were used to obtain the September 1987 grada-
tion. Bed gradations calculated at the end of the 14-year simulation are
compared to the 1987 sample gradations in Figure 11. This comparison is much
better than the comparison of measured and calculated gradations in 1984. The
calculated bed material gradations are well within the scatter of prototype
data, and model performance can be considered reliable.

The numerical model can be used to evaluate the proposed channel
improvements for Unit 4 on Corte Madera Creek. It is recognized that the
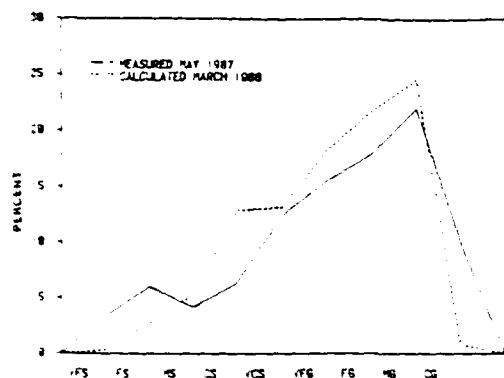
a. Between stilling basin   b. Between College Avenue
   and College of Marin bridge   and College of Marin bridges
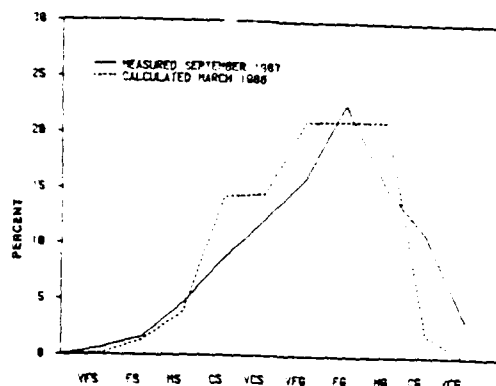
Figure 11. Bed material gradation

reliability of model predictions is somewhat limited due to the uncertainty
related to prototype sediment inflow concentration and variations in bed mate-
rial gradation data. The model was successful in simulating the longitudinal
extent and general quantity of deposition in the concrete channel. Bed mate-
rial gradations were reproduced fairly accurately. These gradations are
important because they influence the roughness of the channel bed. The model
predicted degradation during flood events. Because predicted quantities of
degradation were less than the measured quantities, the model will provide
conservative results with respect to the ability of the channel to maintain a
sediment-free bed with the addition of the sediment basin in Unit 4.

Sensitivity Study

     The sediment inflow rating curves used in the adjusted numerical model
were based on suspended and bed-load samples taken at water discharges less
than 2,000 cfs and on equilibrium transport calculations, with some adjustment
to better simulate deposition surveys and sampled bed material gradations.
Due to the critical importance of sediment inflow on deposition in the con-
crete channel, the sensitivity of the adjusted numerical model to sediment
inflow was tested. The average values for sediment inflow were increased by a
factor of 1.5 and decreased by a factor of 0.5 in the sensitivity study.
These values were still within the range of measured data. Results of a his-
torical simulation from October 1972 to August 1984 are compared in Figure 12.
As expected, the sediment inflow rating curve influences the deposition pro-
file in the concrete channel. The uncertainty and significance of sediment
inflow must be considered when interpreting study results.

Test of Design

     The proposed design for Unit 4 on Corte Madera Creek was incorporated
into the numerical model and the following design conditions were used for
testing the sedimentation effects in the proposed channel: (a) concrete chan-
nel and sediment trap maintained to design invert elevations at the start of
the flood season, (b) the 1982 flood season, with the 100-year frequency flood
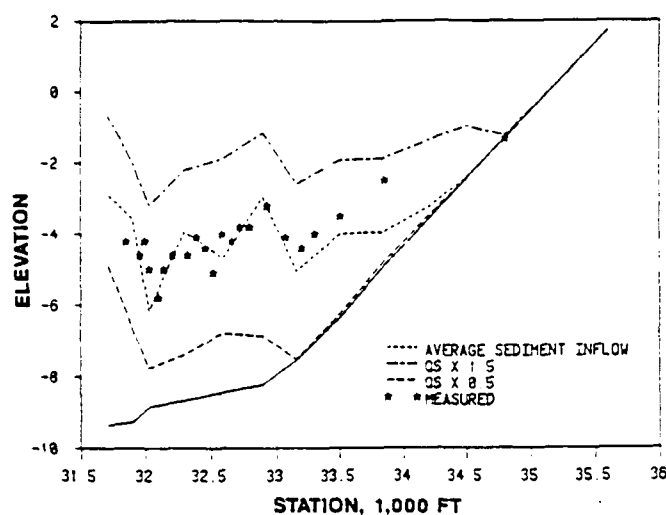replacing the 4-5 January flood, for a design hydrograph; (c) roughness

Figure 12. Sensitivity of model to sediment
inflow, measured and calculated aggradation,
August 1984

coefficients in the concrete channel determined from the January 1982 simula-
tion; and (d) sediment inflow based on long-term average inflow determined by
reproducing measured deposition in the channel.

The design channel operated satisfactorily in the numerical model test
with the design hydrograph. About 900 cu yd of sand accumulated in the con-
crete channel during the antecedent flow period. The sediment trap had accu-
mulated about 2,000 cu yd of material just before the start of the flood
hydrograph. At the peak of the flood, 3,900 cu yd were stored in the sediment
trap, 60 percent of which was sand and 40 percent gravel. Most of the sedi-
ment that had deposited in the concrete channel prior to the flood peak had
washed out by the time the discharge reached 5,000 cfs. Calculated deposition
in the concrete channel at the peak was less than 0.05 ft and can be consid-
ered to represent bed load. The D_{84} of this material was less than 2 mm, so
no increase in roughness was calculated by the model's roughness algorithm.
Maximum calculated water-surface elevations for the design flood were below
the existing wall between the stilling basin and sta 328+00, about 1 ft
higher than the existing wall at College Avenue, and about 2.4 ft higher than
the existing wall at sta 342+00. The TABS-1 model does not account for bridge
losses; therefore, design water-surface elevations will be higher. HEC-2
backwater calculations by Sacramento District indicated that about 0.7 ft of
head loss can be expected by the combined College Avenue and College of Marin
bridges. Flows subsequent to the flood peak deposited material in the con-
crete channel. Some of this material came from the sediment trap. At the end
of the flood season, 2,800 cu yd of sediment were deposited in the trap and
4,500 cu yd in the concrete channel. The channel deposits at the end of the
season were considerably coarser (D_{84} - 5 mm) than deposits from the anteced-
ent flow (D_{84} - 1 mm). The material deposited in the trap at the end of the
season was also coarser (D_{84} - 12 mm) than deposits at the peak and from the
antecedent flow (D_{84} - 6 mm). Calculated gradations of the channel and trap
deposits are shown in Figure 13. Bed changes during the passage of the test
hydrograph are shown for three concrete channel sections in Figure 14.
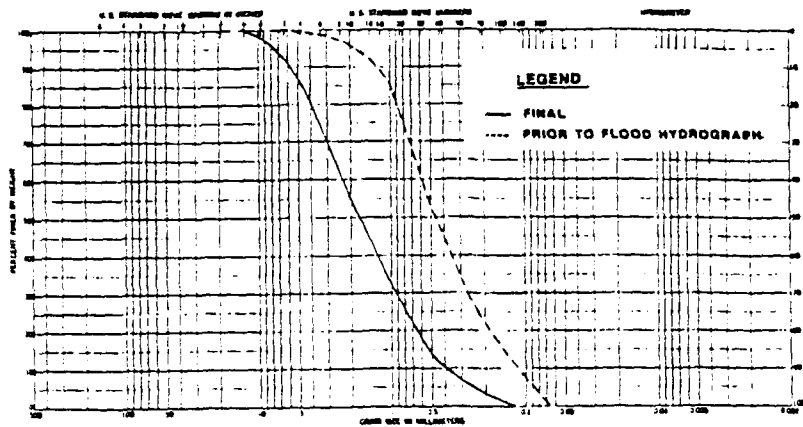
a.   Sta 329+00
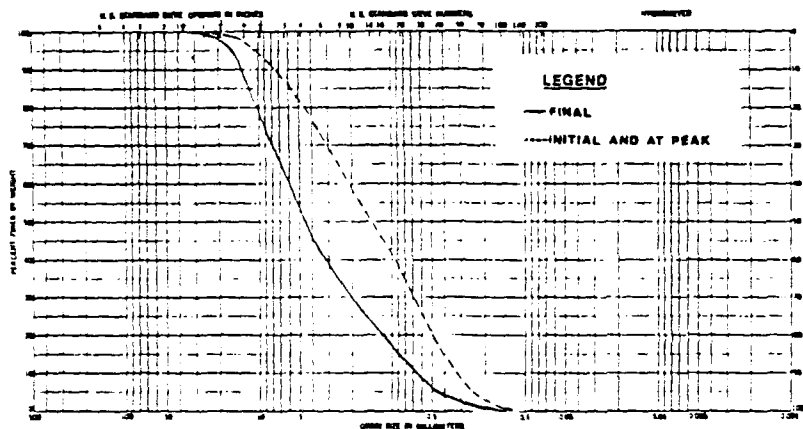
b.   Sediment trap

Figure 13.   Bed material gradations

CONCLUSIONS

Channel roughness in the concrete channel is affected by gravel
deposits. A comparison of backwater calculations and high-water marks from
the January 1982 flood demonstrated that the average Manning's roughness coef-
ficient in the concrete channel with gravel deposits was on the order of 0.030
at the flood peak. Sinuosity, grain roughness on the channel bottom, and
tubeworm and barnacle growth on the channel walls were insufficient to account
for a composite roughness of this magnitude. The additional roughness was
attributed to form roughness due to the movable gravel bed. Criteria, which
included a minimum thickness and applied shear stress, were established to
determine when adding form roughness to the composite channel roughness would
be appropriate. A conceptual framework, considered appropriate for making
comparative analyses of sedimentation in Corte Madera Creek, was developed to
emulate the complicated physical processes as much as possible, given state-
of-the-art knowledge. Using this framework, with the design flood conditions,
a roughness coefficient of 0.022 was calculated for a channel free from sedi-
ment deposits. This value increased to 0.028 when there was sufficient gravel
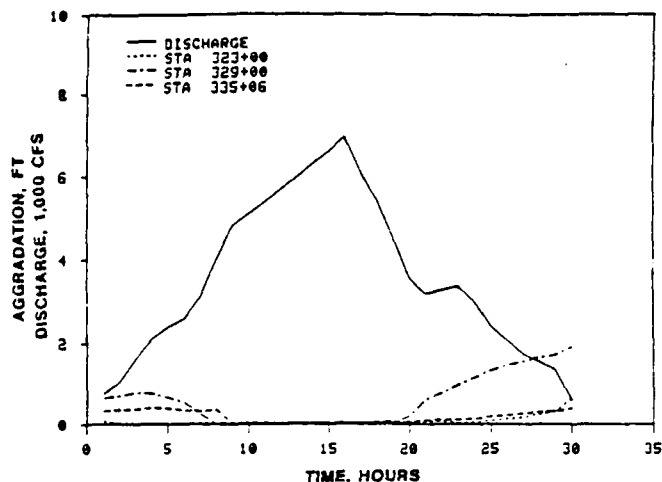
Figure 14. Aggradation in concrete channel
with passage of design flood hydrograph

in the movable bed layer, either by increased antecedent flow or deletion of
the sediment trap, to cause a calculated increase in roughness. When sediment
deposits were not removed prior to the design flood, a value of 0.030 was
calculated due to a thicker and coarser deposit.

Using average sediment inflow rating curves, the numerical model gener-
ally reproduced both measured deposition quantities and sampled gradations
over a 14-year historical period. Sensitivity tests demonstrated that sedi-
ment inflows of 1.5 and 0.5 times the average would produce deposition consid-
erably different than measured quantities. Therefore, long-term simulations
using average sediment inflow are considered fairly reliable.

Probable variations in sediment inflow during a flood event make short-
term predictions less reliable than long-term predictions. However, the sen-
sitivity tests demonstrated that increasing the average sediment inflow by
50 percent did not result in an increase in calculated roughness coefficient.
Results from the sensitivity study relieve some possible concern related to
sediment inflow uncertainty and its effect on channel roughness.

The proposed sediment trap collects both sand and gravel. The numerical
model calculated a trap storage of 3,900 cu yd, 60 percent sand and 40 percent
gravel, at the peak of the design flood; about 2,000 cu yd of this material
was deposited by flow preceding the peak. The effective storage capacity of
the trap is not the same as the volume excavated below the existing bed pro-
file, because during floods, sufficient sediment transport potential exists to
maintain some sediment movement through the trap. The numerical model results
using the design histograph demonstrated that the trap was effective in reduc-
ing channel deposition from flow preceding the peak and in reducing the
coarseness of both the bed deposit and the bed load at the flood peak. This
resulted in a reduction of about 1 ft in the computed water-surface elevation
in the concrete channel downstream from sta 342+00. Deposition in the con-
crete channel at the end of the flood season was about the same, with or
without the trap.

ACKNOWLEDGMENTS

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RIVER BED ADJUSTMENTS IN A STEEP STEP-POOL SYSTEM

(Lainbach, Upper Bavaria)

Peter Ergenzinger

Introduction

Bed adjustments by scour or fill are responsible not only for the buffered hydraulics of discharge even under very unsteady flood waves but also for the related bed load transport. Despite recent advances our knowledge about these phenomena is still somewhat restricted, especially with regard to extremely coarse-grained step-pool rivers. In the case of field studies during non-uniform unsteady floods much remains to be done. Whenever there is bed load transport, it is not sufficient to survey river sections only before and after a flood. Nor can the dynamics of roughness be determined from the average grain-size distribution of bed material. The newly developed "Tausendfuessler" can be used to measure both geometry and roughness, even during floods. First results of this approach are presented from Lainbach in Upper Bavaria.

Test site

The Lainbach catchment, the experimental site of the Geographical Institute of the University of Munich, is situated about 70 km south of Munich close to the monastery of Benediktbeuern at the northern border of the Alps. The experimental reach lies below the confluence of Schmiedlaine and Kotlaine. The reach is 150 m long and the related area of the catchment is 18.8 km² in size. In this region floods occur either during snowmelt or after thunderstorms in July or August. The bed load originates mainly from a more than 100 m thick Quaternary fill of glacial sediments in the central parts of the basin (Becht, 1989). Mass movements and related debris flows transport the material into the river system.

The steep Lainbach is called a "Wildbach" for there are many steps, rapids and pools along the river, and extreme discharges occur with related extreme rates of solid material transports. Kotlaine and Lainbach are buttressed by many small dams (Wildbachverbauung) creating an extreme fall-pool system. Only this measuring site and the Schmiedlaine still preserve the natural conditions of a typical step-pool system (Whittaker and Jaeggi, 1982). This more natural step-pool sequence at the experimental reach is depicted in fig. 1.

Grain size distribution

The grain size distribution of the river bottom material ranges from boulders with b-axis up to 2 m to sandy material in some small pools in the lee of these boulders. The inner channel is paved by cobbles, coarse pebbles are common for the bars, and boulders are restricted to steep step sites.

To match the very different scales of grain sizes different techniques must be employed. Coarse grain sieving was done according to Ibbeken (1974). The paved surfaces were investigated by the "photo-sieving" technique of Ibbeken and Schleyer (1986). In order to include the steps with the boulders a photogrammetric method was used.

At the 5 and 110 meter marks of the experimental reach (fig. 2) volume samples of more than 300 kg were sieved, resulting in the following grain size parameters (Ergenzinger, Stüve 1989):

Tab. 1: Comparison of grain-size parameters created by sieving and photo-sieving

		D_{50} (mm)	D_{84} (mm)
5 meter mark	sieving	75	135
	photo-sieving	160	280
110 meter mark	sieving	60	120
	photo-sieving	120	205

Both samples are from pool reaches. The coarseness of the photo-sieving samples are the result of pavement processes. Compared to the grain-size distribution created by sieving, the D_{84} equals the D_{50} of the pavement.

Along the step reaches the boulders must be taken into account.During floods with more than 8 m^3 s^{-1}, the boulders are submerged and then become "singular roughness elements" (Bathurst

). To describe this situation, vertical and oblique photos were taken along the whole measuring site with a special photogrammetric camera. For all boulders and cobbles with b-axis bigger than 120 mm the grain-size and the position were measured by Peter Stüve, using the Rollei picture analysis system MR 2. For this interpretation the experimental reach was differentiated into step- and pool-reaches according to fig. 1. The remaining area not covered by boulders and cobbles was filled up according to the grain-size distribution given by photo-sieving. The representative grain size parameters resulting from these combined distributions are shown in figure 2 and table 2.

Table 2: Representative grain size parameters for steps and pools
        of the experimental reach

						D_{50} (mm)	D_{84} (mm)
1. Step:	5 -	30 m:	MR 2 and p-s. at	5 m		650	1380
2. Pool:	30 -	55 m:	MR 2 and p-s. at	5 m		320	645
3. Step:	55 -	80 m:	MR 2 and p-s. at	110 m		805	1710
4. Pool:	80 -	110 m:	MR 2 and p-s. at	110 m		290	535

The steeper the step the coarser the grain size distribution. The parameter D_{84} is shifted by the boulders up to 1.7 meters. With one exception these boulders were never in motion during the last two years. The steps are the most stable reaches of the river bottom, whereas the pools changed their shape during every flood, and most of the surface material was replaced several times.

The different grain-size investigations resulted in a wide range
of so-called representative parameters. In order to decide which
one is "representative" for this Wildbach a new device, the
"Tausendfuessler", was installed at the 110 m mark of the
experimental reach, in a pool section.

The "Tausendfuessler" device and its application for measuring
river bed adjustments

By scour and fill river beds are adjusted to discharge. Since most
adjustments occurr during the most unsteady and non-uniform runoff
conditions, the observation and measurement of these changes in
geometry are invisible and difficult to measure. The
Tausendfuessler is a simple device for measurements of the
changing river cross sections and was developed and applied for
the first time at Squaw Creek in Montana (USA) in 1988. To measure
the conditions of bed and water surface at always the same spots
and along the same length intervals, a log or a tube is installed
horizontally across the river and holes are drilled vertically
every 10 cm (fig. 3). A measuring rod is driven through each hole
and the distance between the horizontal log or tube and the upper
end of the rod is measured. Depending on the stability of the
installation it is possible to locate the points again with a
precision of 0.5 to 1 cm. At Lainbach the first Tausendfuessler is
installed at the 110 meter mark and consists of three tubes. Tube
1 is 1.65 m upstream of tube 3, and very close to tube 3, on the
upward side, there is the parallel tube 2 (distance 0.15 m). The
aim of this installation was to investigate first, the
investigation of small scale variabilities along the tubes 3 and 2
and, second, the local changes in bed slope by comparing tube 1
and 3. Since late summer 1989 two measuring bridges have been
installed across a step and a pool section. Now it is no longer
necessary to wade along the Tausenfuessler during flood. But since
one total observation along one section lasts about 45 minutes the
sequences of measurements are still too long.
The flood of July 2/3 1989 peaked close to 10 $m^3 s^{-1}$ and lasted
more than 30 hours due to the rainfall conditions (fig. 4). At the
pool runoff was always subcritical with the maximum Froude number

close to 0.85. During flood, 9 measurements at tube 3 were made. The changing bed conditions are shown by the stripped area in fig. 5. The first and the last profiles of the sequence are shown separately. They stay close only at the right bar. The maximum extent of river channel changes during flood for all 9 sections defines a band of maximum scour and fill. With more than 25 cm, this band is especially thick at the inner channel and on the left bar close to the left bank. On average there are 15 cm of scour and accretion, resulting in 1.87 m² of changed river section.

In order to describe these adjustments in more detail, the river bottom was divided into the inner channel (6.5 m to 11.5 m) and the two bars (1.5 to 6.5 m and 11.5 to 13.5 m). By means of the 9 sections the adjustments in geometry can be represented in 8 steps of different time length. For each time interval the measured scour and fill and the absolute total of differences in area were determined. In order to compare the resulting figures, they were converted from m² per time interval to mm per hour by dividing them by the corresponding river width and time interval between the different measurements (fig. 6).

In accordance with the hydrograph showing peak runoff at 2 a.m. and a secondary maximum at 4 p.m. next day there was intensive erosion at the very beginning of flood on both parts of the section. Afterwards the development of the adjustments was more differentiated. The inner channel was first filled, then just after peak flow slightly eroded, but maximum erosion occurred at the end after the second flood wave. By contrast, the bar was intensively eroded just after peakflow, but stabilized afterwards, and fill was dominant after the second peak during recession. The discussion of these reactions and adjustments will be postponed until more is known about the related changes in hydraulics and roughness.

The k3 concept and the determination of changing roughness conditions.

Measuring the depth of the river bottom every 10 cm by the Tausendfuessler not only reveals the changing geometry of the river bed, but results can also be interpreted for the changing roughness conditions. Since it is impossible to obtain bed surface samples under flood conditions a more indirect technique must be used to describe the changing roughness conditions. Roughness creates turbulence mainly due to vertical differences of the river bottom. Furthermore, in the case of coarse material river bottoms the common distinction between grain and form roughness is very difficult to apply. The same coarse particle can create form roughness under low flow conditions and grain roughness under high flow conditions. This is especially true for boulders. To overcome these artificially defined differences the concept of k3 roughness was created: k3 roughness is defined as maximum difference in height across three neighbouring points at 10 cm intervals. The aim is to define a representative number which reflects the turbulence created by vertical differences across one third of a meter. This length was chosen because the impact of larger cobbles can be measured on both sides of the obstacle over this distance. In order to describe the k3 situation of a certain section of the river bed, the sweeping maximum differences of height across 3 decimeters were determined (fig. 7).

At Lainbach the averages of these sweeping maximum differences were determined for the above defined two parts of the cross section. The aim is to obtain the different roughnesses during a flood both on the bars and in the inner channel. The k3 values changed between 28 and 45 mm. In the inner channel low k3 numbers occurred at the very beginning, just before peak, and before the secondary peak of the flood wave. These low k3 numbers were always due to a small but very effective layer of pebbles infilling the gaps between the pavement cobbles. Rather rough conditions existed during peak runoff and during the recession after the secondary peak. The bar developed similarly except that at the very beginning and during the secondary peak flow the bar had a higher k3 number than the inner channel. Again the highest k3 number

occurred during recession. The k3 numbers differ from the D $_{50}$ of the grain-size distribution and amount to only about one tenth of the measured b-axis. Nevertheless the k3 numbers change by a factor close to two during a single flood wave.

The changing roughness conditions can be described by the data of relative roughness (Rr = d / k3 with d = average water depth). Since grain size has been replaced by the more dynamic k3 the resulting datas of relative roughness change quite dramatically. Especially in the inner channel the phases with reduced k3 numbers are phases with rather high values for relative roughness. Due to greater water depth in the inner channel the values of relative roughness are always higher here in comparison with the bar.

The good correspondence of relative roughness with Darcy-Weissbach roughness is rather astonishing (f_D = (8 g d S) : v^2 with S = slope). By transferring f_D to $1/f_D^{1\,2}$ (fig. 8) a similar distribution to that of relative roughness is obtained . The dynamics of roughness changes of different parts of a river are easily described if the related water depths and the differences in height of the river bottom are known. This is easier to measure than for instance. the slope of the different parts of the section. The "Tausendfuessler" approach is useful not only to measure the changes in geometry but also to determine the related changes in roughness.

The dynamics of the river and bed adjustments

During flood velocity profiles were measured by current meter across the "Tausendfuessler" section. The resulting data were used to determine stream power and mobility number. The mobility number reflects the big changes in geometry and roughness at the beginning and at the end of the flood far better than stream power (fig. 9). The mobility number (Fr = v^2 : (g k3) does not include slope as a parameter, and grain size is replaced by k3 in order to include more of the changing conditions of the river bottom.

The relationship between roughness and mobility number is straightforward (fig. 10). There is some bias between the reactions on bars and inner channel. The gradient of Darcy-Weissbach roughness increases more rapidly on bars than in the inner channel. On the bottom of the inner channel big changes in roughness at the start of flood are not reflected by high mobility numbers. These are effects of a temporary input of bed load which does not correspond with the local dynamics of runoff.

The changes of mobility numbers for the inner channel and bar and their relationship with the changing geometry of the "Tausendfuessler" section can be studied by comparing the figures x and y. If the heavy erosion at the very beginning of flood is excluded the important changes occur during the phases when the mobility number is higher than 2. The very rapid changes of erosion and fill during the rising stage of the hydrograph is not reflected by similarly changing mobility numbers. During the periods of maximum gradient of the mobility numbers there was accretion in the inner channel at the rising limb and erosion at the falling limb. The same tendency is true for the bar but here the high gradient mobility numbers with erosion occurred just after peak flow of the flood. If the amount of change in geometry is compared with the related mobility numbers (fig. 11) the tendency is quite clear: excluding the first and the last observations at the very start and during the last phases of recession there is growing change with growing mobility numbers. But the extent of change can vary and there is no rule relating it to scour or accretion. At the pool below the "Tausendfuessler" there are very important changes in pool geometry at the beginning and towards the end of flood. This results in excessive amounts of erosion related to a rather low mobility of below 2 and rather small amounts of accretion during recession when the mobility number is still close to 13. These biases are related to phases when the impact of bed load created by erosion of the river bed just above the site is especially important.

Results

The new "Tausendfuessler" device measures the changing conditions both of geometry and of roughness during a flood wave. The dynamics of river bed adjustments are quite complicated, for there are many possibilities of negative or positive feed-back. In step-pool systems like the Lainbach the pools are very sensitive areas which react quickly to runoff changes. In contrast the neighbouring step areas are stable and no big change occurs during low or average floods. This is due to the remarkable difference in grain size between the two different parts of the river long profile.

The dynamics of roughness changes of different parts of a section during a flood can be studied with the help of the k3 concept. The study of grain-size distribution of the superficial material is replaced by the determination of differences in height above the local level. Bottom height is measured every 10 cm, and by determining the maximum height difference between 3 neighbouring measurements it is possible to obtain the moving differences and the related averages. What is striking that even during the rising limb of the flood there are phases with low k3 numbers created by incoming pebble material. The k3 numbers are used to replace grain size for the calculation of relative roughness and mobility numbers.

The extent of change of the cross section due to scour and/or fill was determined by the comparison of different "Tausendfuessler" measurements. There are extreme reactions again at the very start of flood when the mobility numbers are rather low. For the flood the reaction measured as change in mm per hour shows quite a good relationship with the mobility numbers, but there is no change to determine the phases of erosion or accretion by means of the mobility number .

As long as the input of bed load material from the river above the measuring site is unknown, the resulting reaction of the river bed during the unsteady conditions of a flood cannot be precisely determined. A very important factor here is the availability of

rather coarse material sufficient to reduce scour and to change the local geometry of the river bed. Since availability is very important for step-pool rivers more studies are needed focusing on the sources of coarse material and their distribution along the river.

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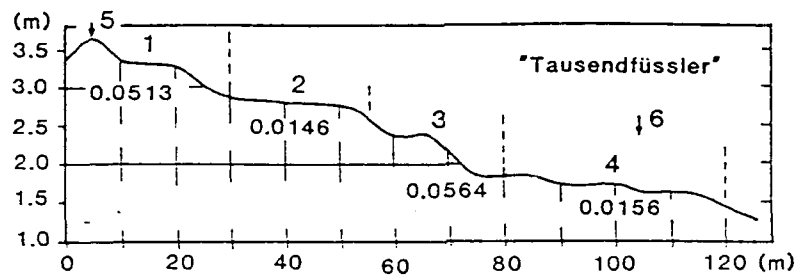
Figures:

fig. 1: Longitudinal profile and the average slopes of the steps
         and pools of the experimental reach
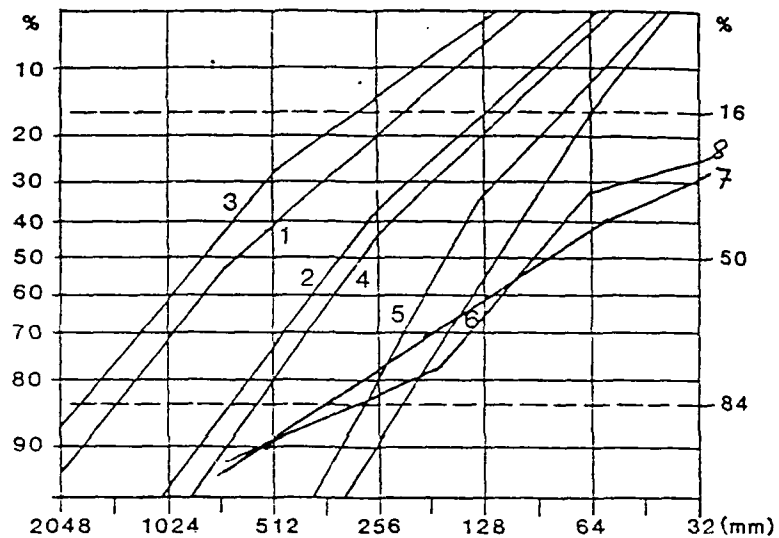
fig. 2: Grain-size distribution for distinctive parts of the
         experimental reach
         1 and 3: grain-size distribution for step reaches
         2 and 4: grain-size distribution for pool reaches
         5 and 6: grain-size distribution according to photo-
                  sieving at 5 and 110 m
         7 and 8: grain-size distribution according to sieving
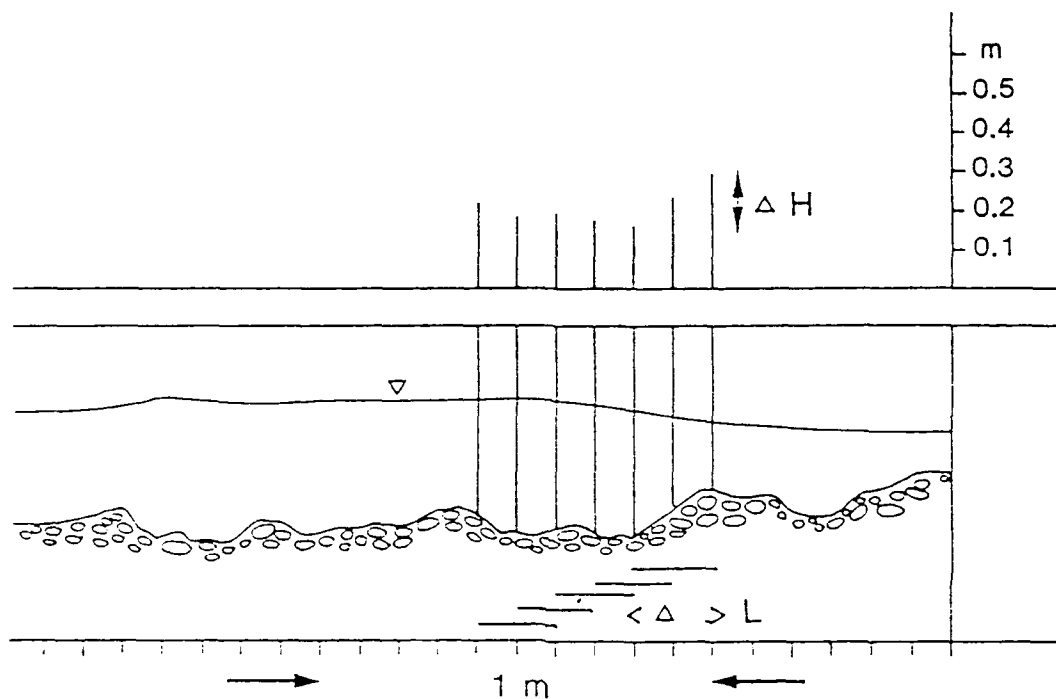   at 5 and 110 m

fig. 3: Tausendfuessler device: Differences of height of the
         river bottom are measured by rods at 10 cm intervals.
         The sweeping maximum differences of height across 3
         decimeters are representative numbers for roughness
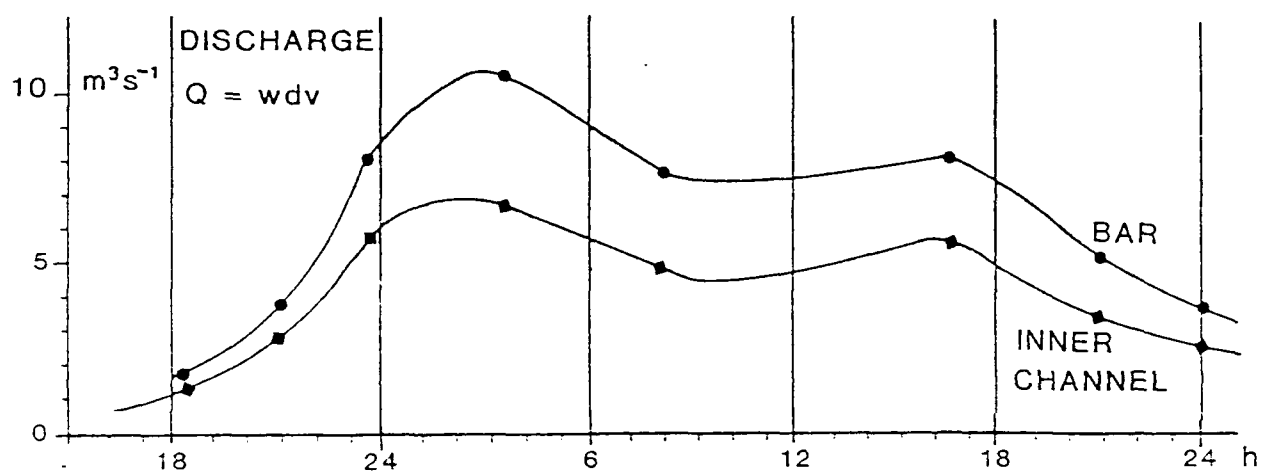   height

fig. 4: Hydrograph of flood at Lainbach 2/3.7.1989

fig. 5: Maximum extent of 9 Tausendfuessler section measured
         during the flood in comparison with the section before
         and after flood

fig. 6: River bed adjustments during flood. Comparison of the
         total number of river bed changes during flood with
         the resulting amounts of scour and accretion of the bars
         and the inner channel

fig. 7: An example for the sweeping values of maximal differences
         across 3 decimeters

fig. 8: Changes of roughness of inner channel and the bars during
         flood. a.: k3 roughness
                b.: Relative roughness
                c.: Darcy-Weissbach roughness

Fig. 9: Runoff dynamics during flood of 2/3.7.1989

                a.: Hydrograph
                b.: Stream power
                c.: Mobility number

Fig. 10: Relationship between mobility number and Darcy-Weissbach
         friction

Fig. 11: Relationship between mobility number and the total number
         of river bed changes

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FLOOD AT LAINBACH 2/3.7.1989

fig. 4: Hydrograph of flood at Lainbach 2/3.7.1989
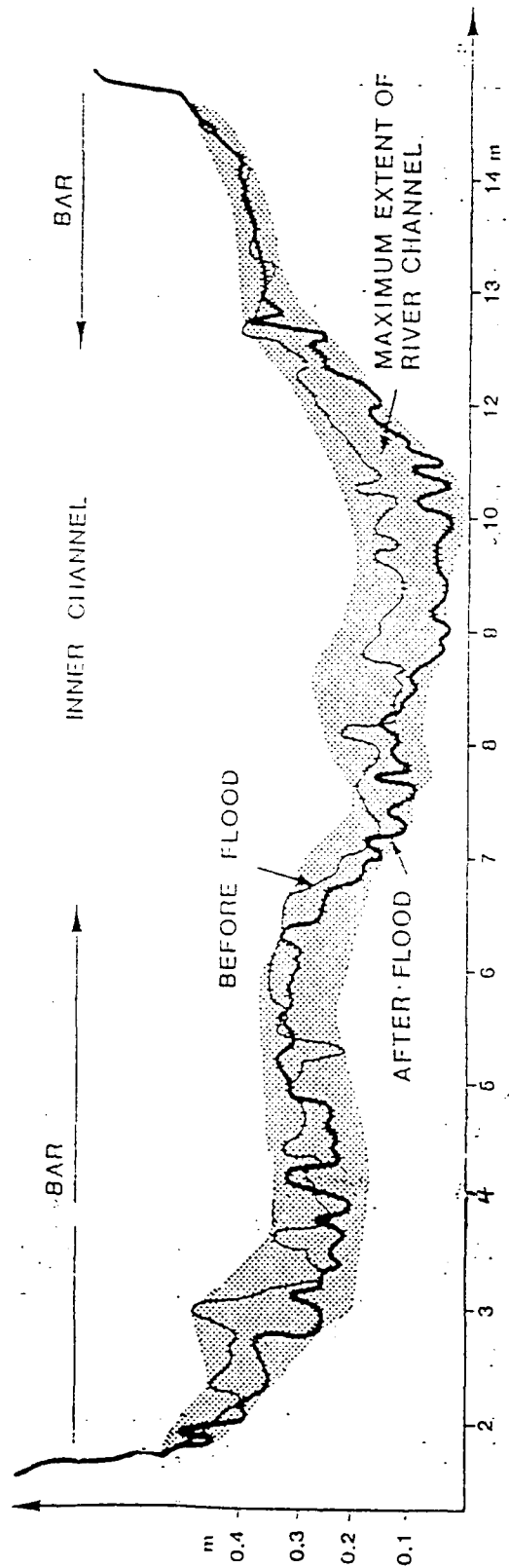
fig. 5: Maximum extent of 9 Tausendfuessler section measured during the flood in comparison with the section before and after flood

fig. 6: River bed adjustments during flood. Comparison of the
        total number of river bed changes during flood with
        the resulting amounts of scour and accretion of the bars
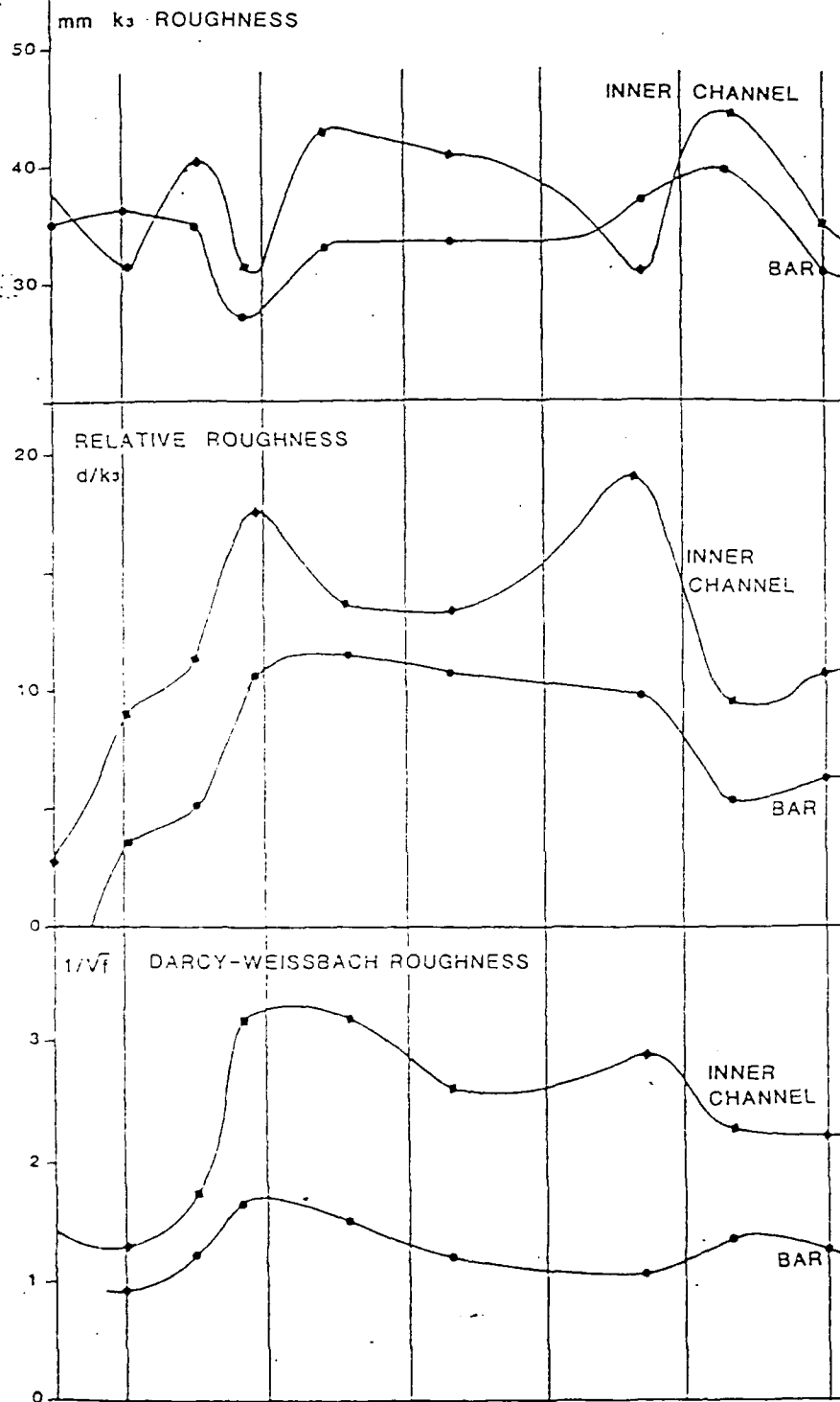   and the inner channel

MOVING AVERAGE OF MAXIMAL DIFFERENCES ACROSS 3 DECIMETRES
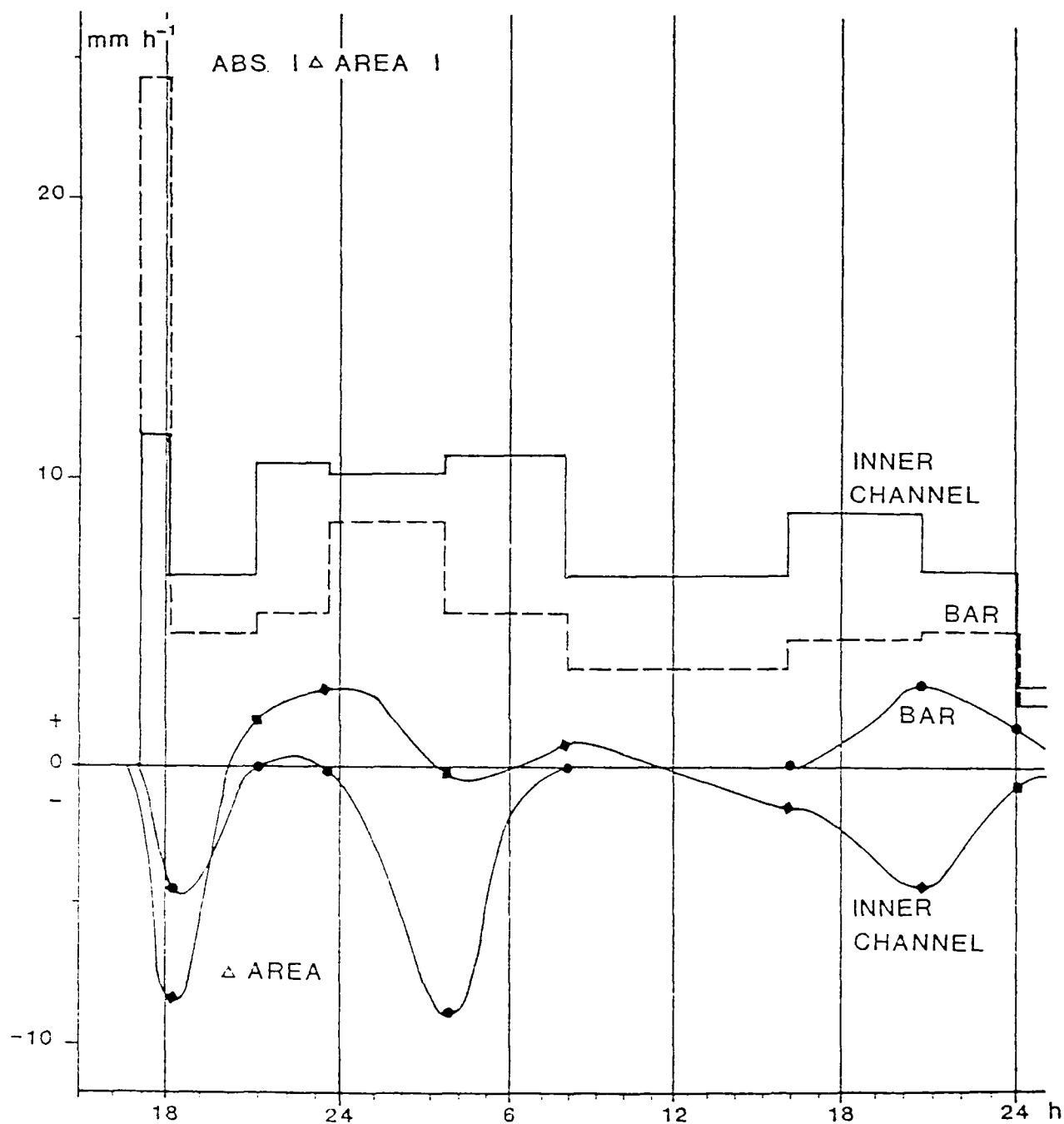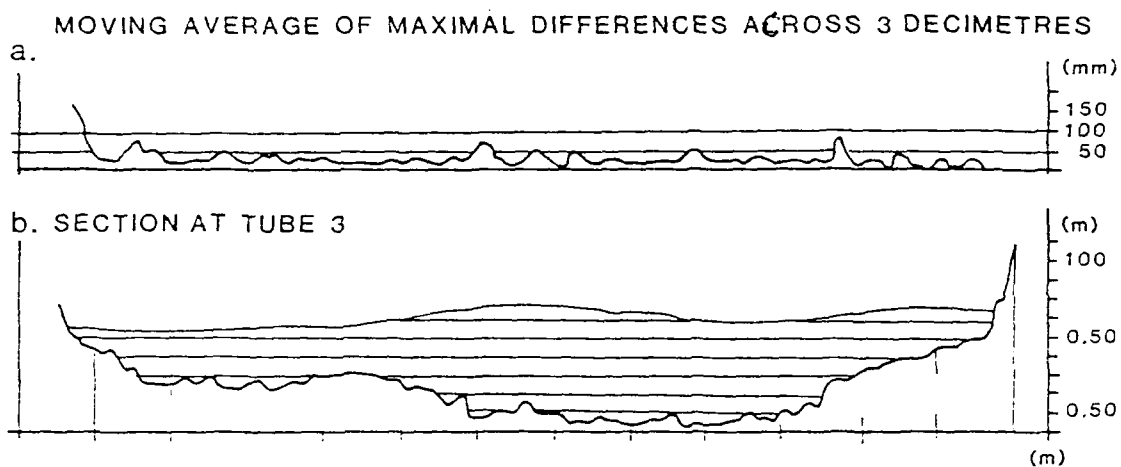
fig. 7: An example for the sweeping values of maximal differences
across 3 decimeters

FLOOD AT LAINBACH 2./3.7.1989

mm k₃ ROUGHNESS

[Figure: three stacked plots of roughness versus time during the flood at Lainbach]

Top plot — vertical axis labeled mm k₃ ROUGHNESS with values 50, 40, 30. Curves labeled INNER CHANNEL and BAR.

Middle plot — RELATIVE ROUGHNESS, d/k₃, vertical axis 0, 10, 20. Curves labeled INNER CHANNEL and BAR.

Bottom plot — 1/√f DARCY-WEISSBACH ROUGHNESS, vertical axis 0, 1, 2, 3. Curves labeled INNER CHANNEL and BAR.

3: Changes of roughness of inner channel and the bars during
   flood. a.: k3 roughness
          b.: Relative roughness
          c.: Darcy-Weissbach roughness

FLOOD AT LAINBACH 2/3.7.1989
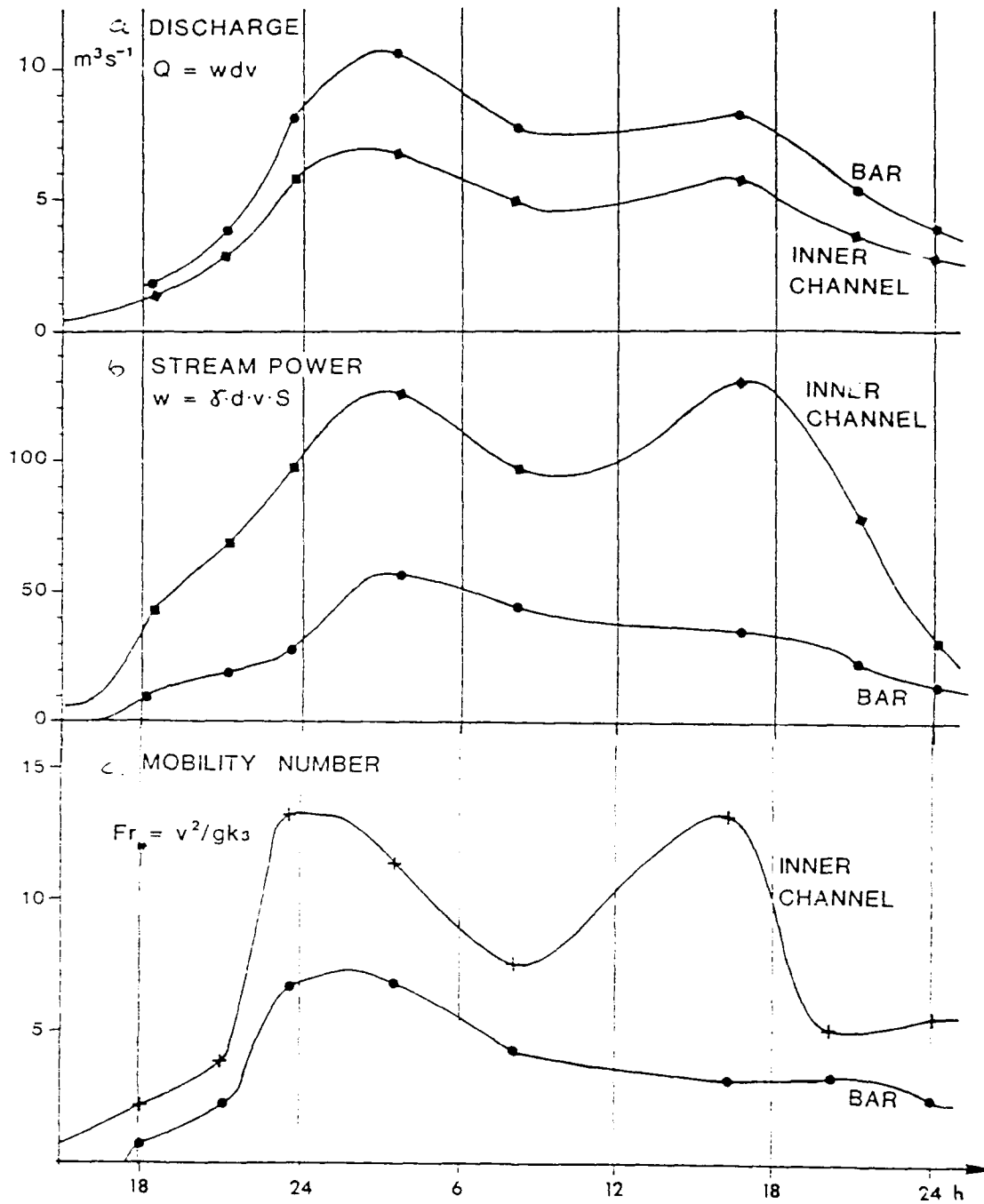
Fig. 9: Runoff dynamics during flood of 2/3.7.1989

a.: Hydrograph

b.: Stream power
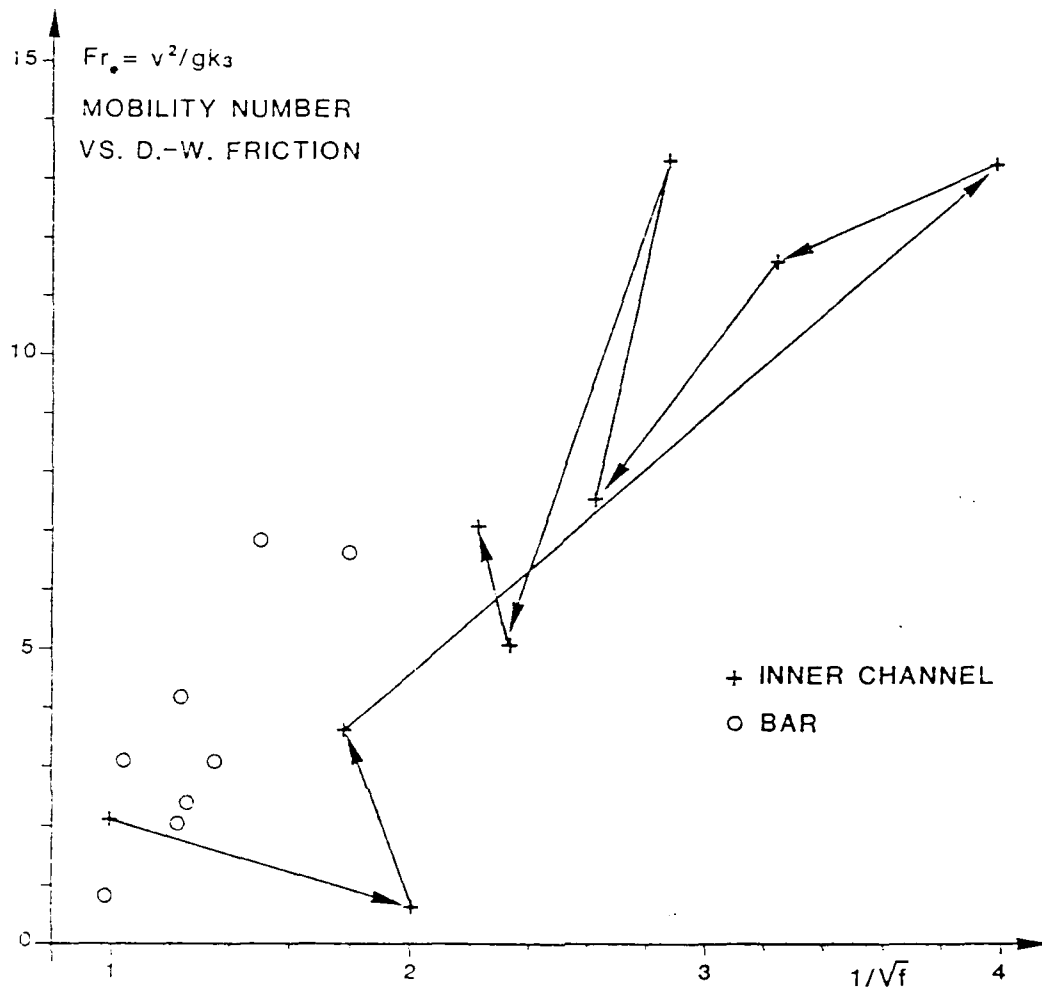
c.: Mobility number

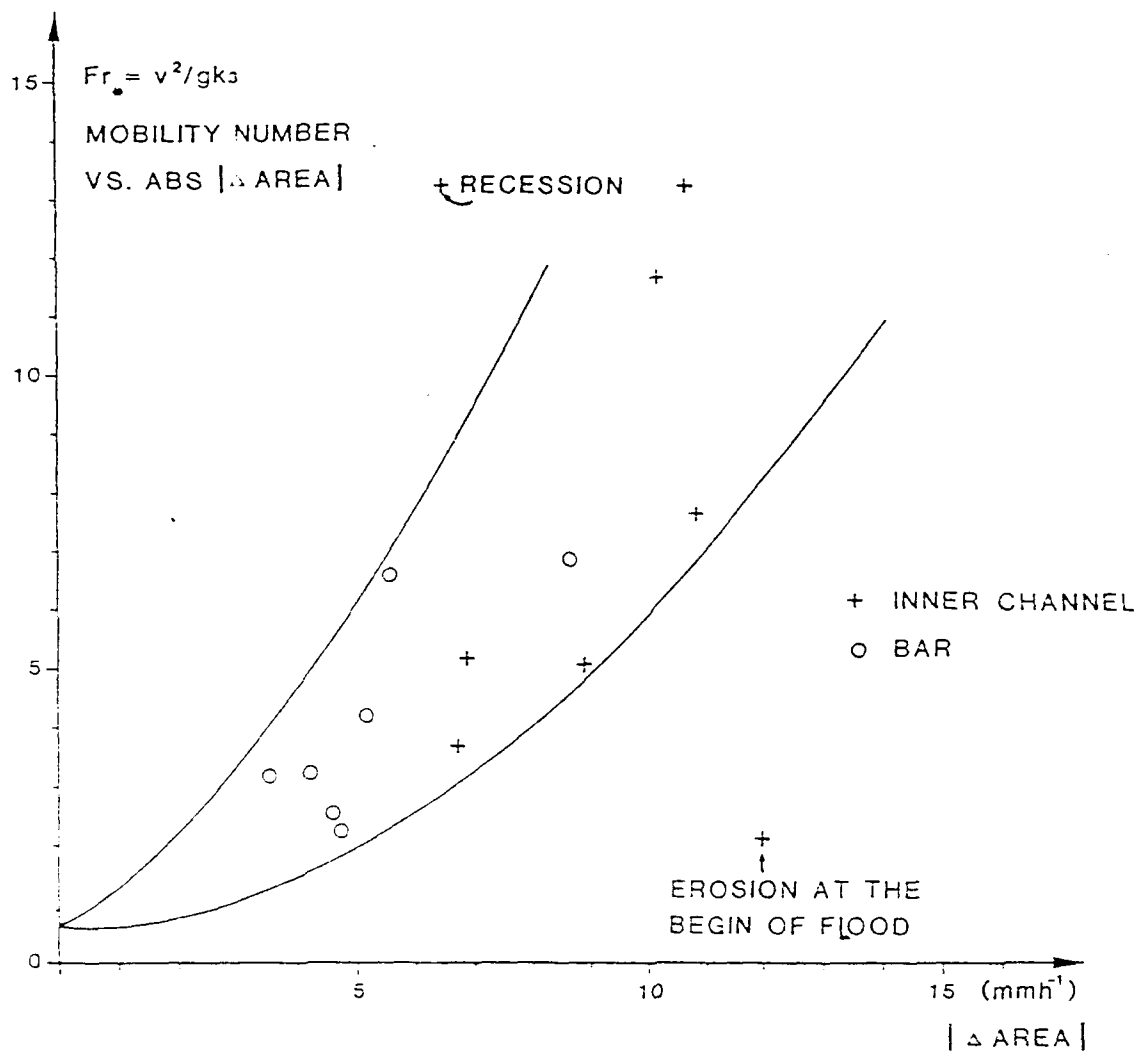Fig. 10: Relationship between mobility number and Darcy-Weissbach friction

Fig. 11: Relationship between mobility number and the total number
of river bed changes

C. FRIZ, G.GIADA, V. VILLI

Long term sediment budget in an alpine catchment

INTRODUCTION

Since 1983, hydrological and geomorphological observations of
the Missiaga basin situated in the Belluna Dolomites, on the
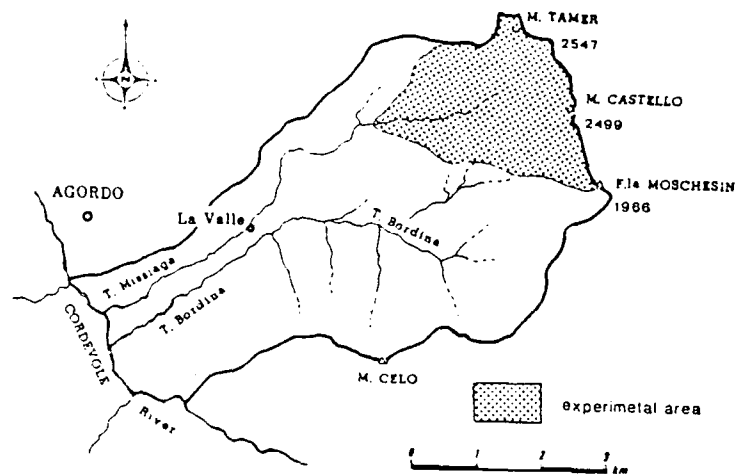hydrographic left of the Cordevole River have been effected
(Fig. 1).

Fig. 1 Location of the study site

The resalts obtained up to new, above all for those regarding the transport of solid material, have been encouraging, in that it has been possible to define the relationships, even thoug only empirically, between the quantity of material transported (m3) and the significant hydrologic parameters, such as, for example the peak discharge and the quick flow volumes.

These results are, howerer, limited because of the short observation period wich does not allow for making temporal extrapolations and the physiographic nature of the basin situated, as it is, in the mountainous dolomite region.

Nevertheless, being aware of the limits of their validity has meant that, at the same time as data was being aquired about the volume of material removed from the basin on each during hydrological events, research into the indirect quantitative evalutation of the erosion was also begun.

This problem is rather complex, given that all case must be dealt with individually as function of the morphological evolution of the ridge and the possibility of being able to reproduce such modification graphically (PELLEGRINI, 1984).

From an operational poit of view in fact, the indirect quantitative evalutation of the erosion is resolved by calculating the volume between the real topographical surface and an artifically constructed reference surface relating to a predeterminated moment in the morphological evolution of the area being studied.

Therfore, placing the reference surface within a certain time period, is of primary importance, and that must, of necessity, lead to taking great care when making assessments of a geomorphological nature.

METODOLOGY

In the case of the Missiaga-Bordina basin it was held that an overall indirect estimate of erosion could not but begin from the immediate postglacial period. During that period in fact, after the widespread settling phenomena and the receding of the Cordevole glacier, the erosion processes began wich have

resulted in the present day morphological configuration of the basin.
  This choice was also conditioned by the fact that is was

unnecessary to specify the nature or define the erosion processes wich the basin had been submitted to, given that these occured probably under constantly changing climatic and enviromental condition.

Thus, after giving a general outline of time period within wich the volume of material that was removed was estimated, it was decided to make use of C14 analysis of tree trunks in an upright position and still present on the banks of the Cordevole river, in order to obtain absolute chronological dating for greater precision. The remaine of the tree trunks were found in a tract of the Cordevole river-bed where an alluvial fan formed by material orininating from the enormous mass of detrirus in the Missiaga-Bordina valley, barred the course of the Cordevole, giving rise to the Agordo Lake.

On the basis of the above mentioned absolute dating, it can be said that such event occured about 5900 B.C., therfore around the superior preboreal period (VENZO S., 1977).

An analysis of the bibliographic data together with the survey data has meant that a reasonably accurate reconstruction of the topographic nature of the Missiaga-Bordina basin has been possible for the end of the Wurmian period (Fig. 2)

Substantial modifications took place in the basin during the immediate postglacial period as a results of the grandiose phenomena wich occured as well as the large quantity of detritus transported and accumulating downstream.

Howerer, it remains a fact that, during the late wurmian period the depression was gradually filled by a large mass of detritus, the volume of wich is difficult to quantify.

It was howerer, possible to determine the quantity of detritus which was removed from the accumulation and transported downstream to form the already mentioned alluvial fan. From the field survey, sufficient objective data were obtained to reconstruct, with a high degree of precision, the topography of the basin during the phase of maximum accumulation, such as to realise the chart reproduced in figur 3, wich was the taken as the original reference surface for the successive calculation of the volumes removed.

On the basis of the morphological elements described above, the quantification of the volume between this reconstructed topographic surface and the actual present day (Fig. 4) was carried out by means of the computerized system DIGCON. This system first realizes the discretization of the area under examination by superimposing a reguarly sized gride, in wich the altitude of each node is found by interpiolation starting from the digitate level curves.

The grid consist of the intersection of all level curves

with two groups of profiles wich are at right angles to each
other in a N-S and E-W dirction. The interdistance between
the profiles obviously define the basib mesh size on the gris
itself.
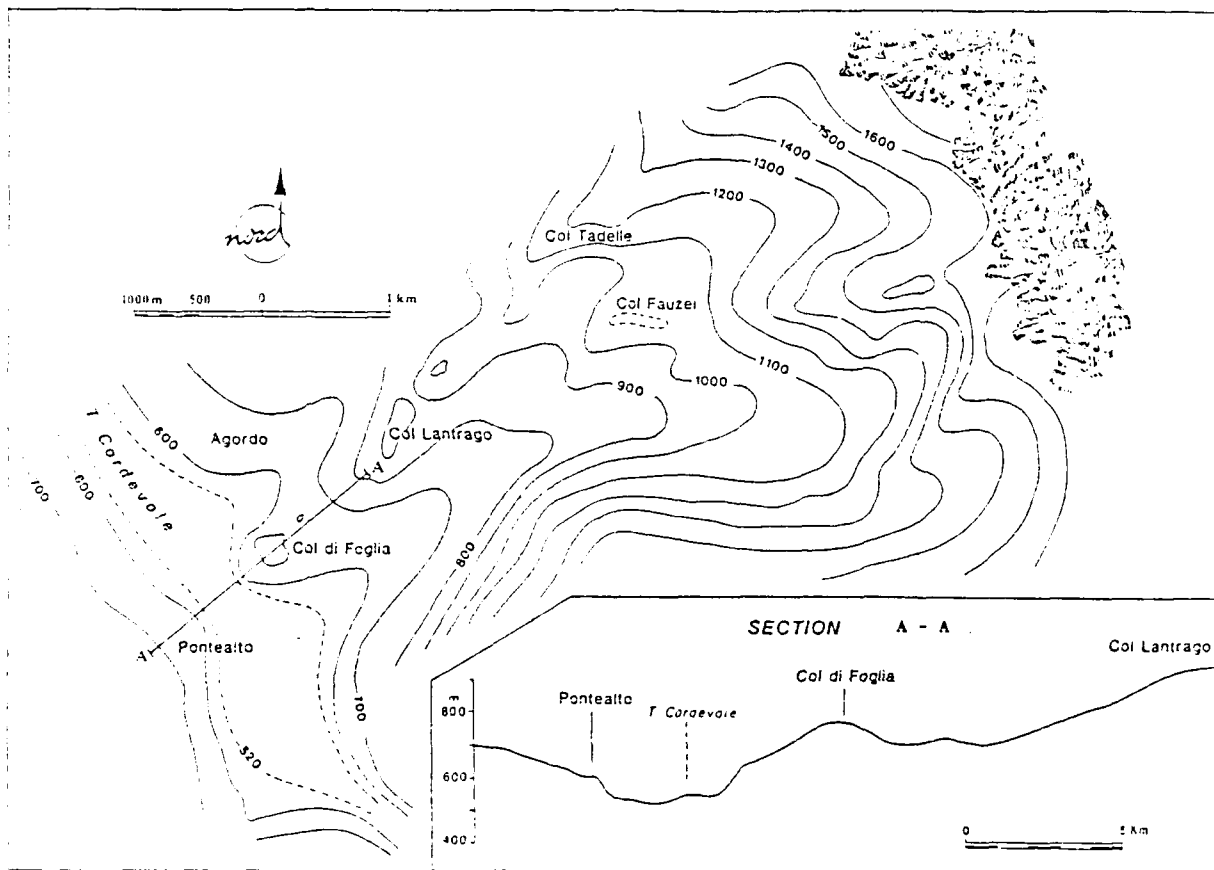The altitude in each node, as already mentioned, is deter-

Fig. 2 The Missiaga-Bordina valley at the end of Wurmian period

mined by linear interpolation separately along the two righy
angled profiles that characterize it. Whenever the calculated

values do not coincided (which can happen when in the two
right angled direction, the slope and/or distances between
the intersections are different) then the weighted average
of the determining values are calculated, the inverse distan-
ce of the nearest intersections in the two right angled
directions being used as a weighting factor.
The sizes of mesh, or rather the interdistances between the
profiles, bearing in mind the peculiarity on the
morphological element which must be show and the actual
topographical situation was chosen as 100m.
Finally, by comparing the actual and postglacial digital
model, it was possible to estimate, for every mash, the
volume of the parallepipedon whic had as its base the mesh,
and the arithmetic mean of the four different values of the
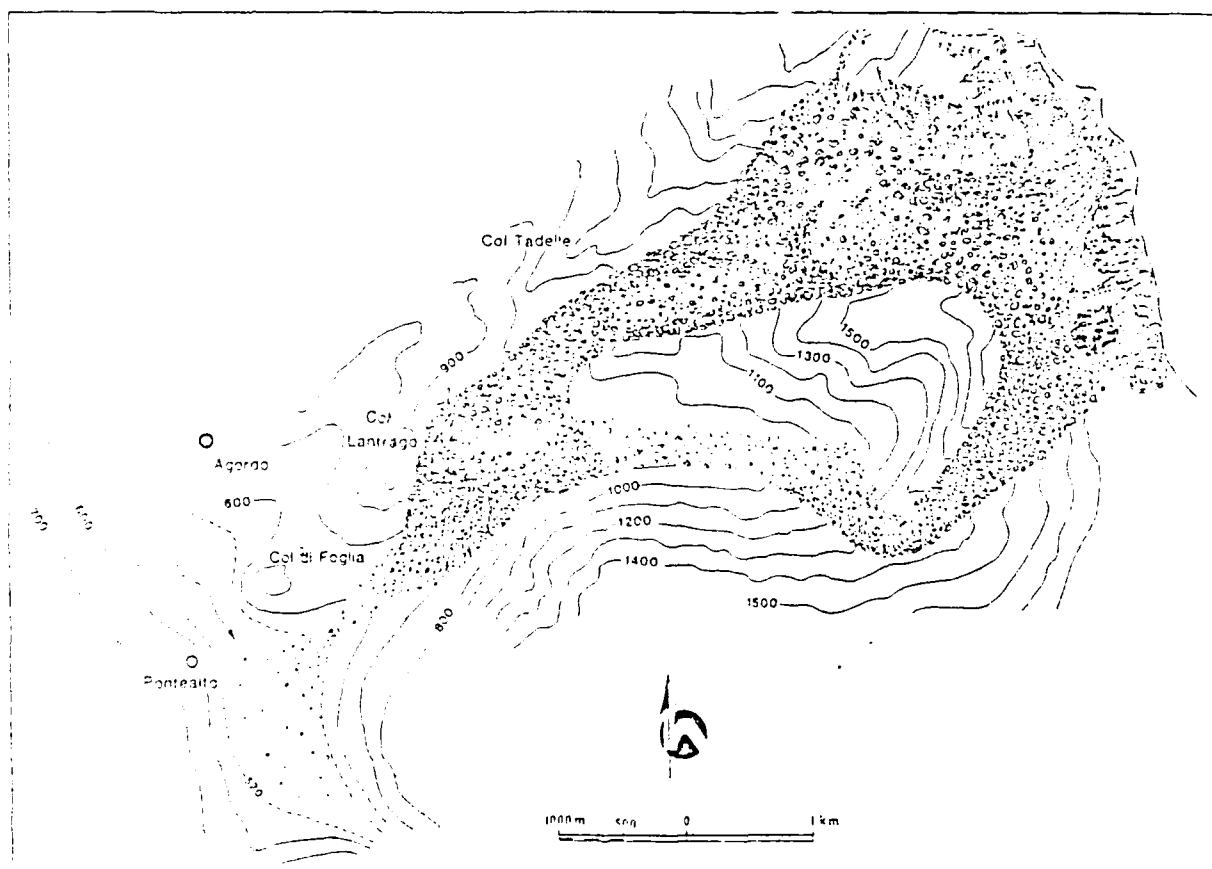altitudes at the nodes as its height.

Fig. 3   The Missiaga-Bordina valley quite filled (about 8000 years ago)
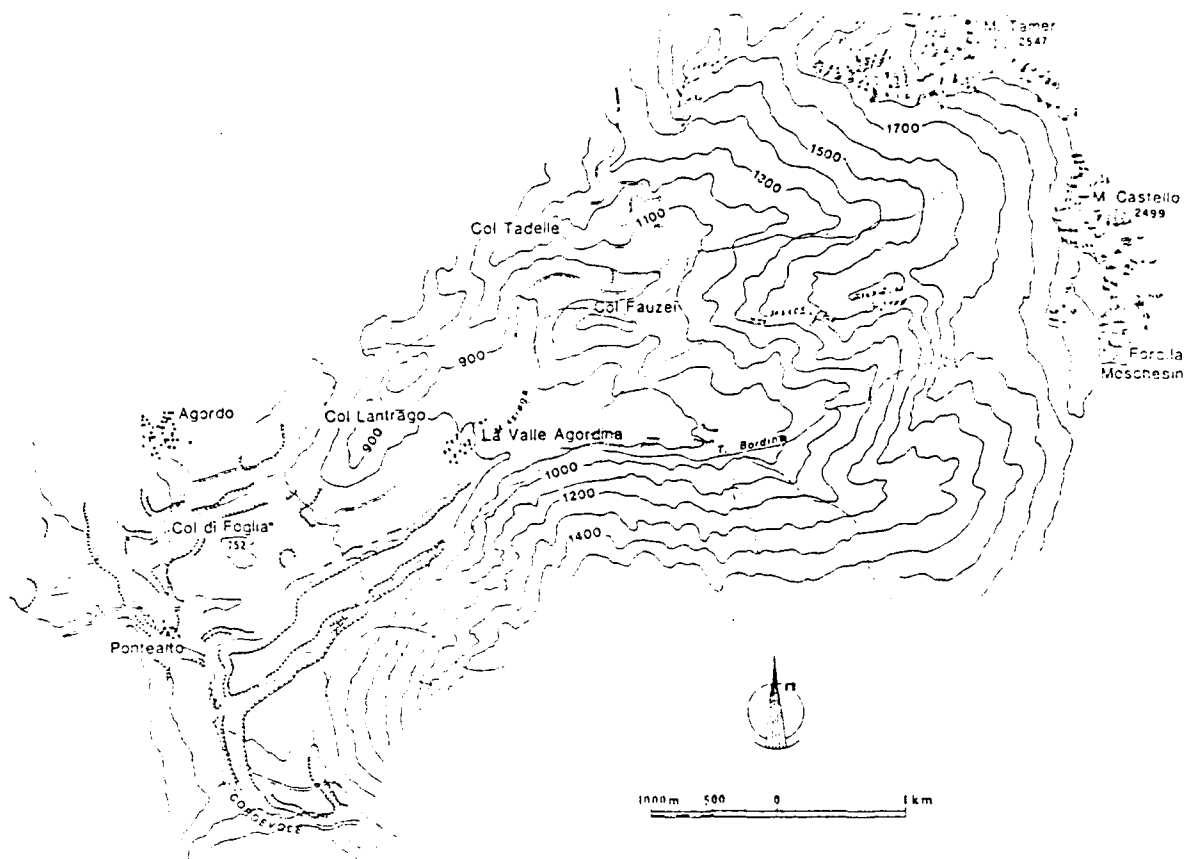
Fig. 4 Actual topography of the Missiaga-Bordina valley

RESULTS

The algebreic sum of the single volumes then gave the overall
results of 1.1 milliard m3
It should be kept in mind that:
- the Missiaga-Bordina basin that was filled up covers an
  area of 9.08 sq.km;
- on the basis of the chronological data furnished by the C14
  analysis, the erosion of the detritus could have begun about
  8000 years ago, which means that the average annual unitary
  loss is 15.143 m3/sq.km, a value which demonstrates unequi-
  vocably the force of the erosion processes in wearing away
  the detritus whic had accumulated in the Missiaga-Bordina
  valley.

It is not possible to compare this average value with the
data relating to the transport of solid material during the
accumulation of this in the upper part of the Missiaga basin,
given that it can not be assumed that the condition which
brought this about were homogeneous.
Howerer that under present geomorphologicla ad climatic con-
ditions, the transport of solid material into the basin under
examination amounts to about 120 m3/sq.km/yr.
This is an unitary value that is congruent with many of data
furnishsd by CATI (1981, see table) deduced from the filling
up of natural and artificial alpine lakes (Fig. 5).

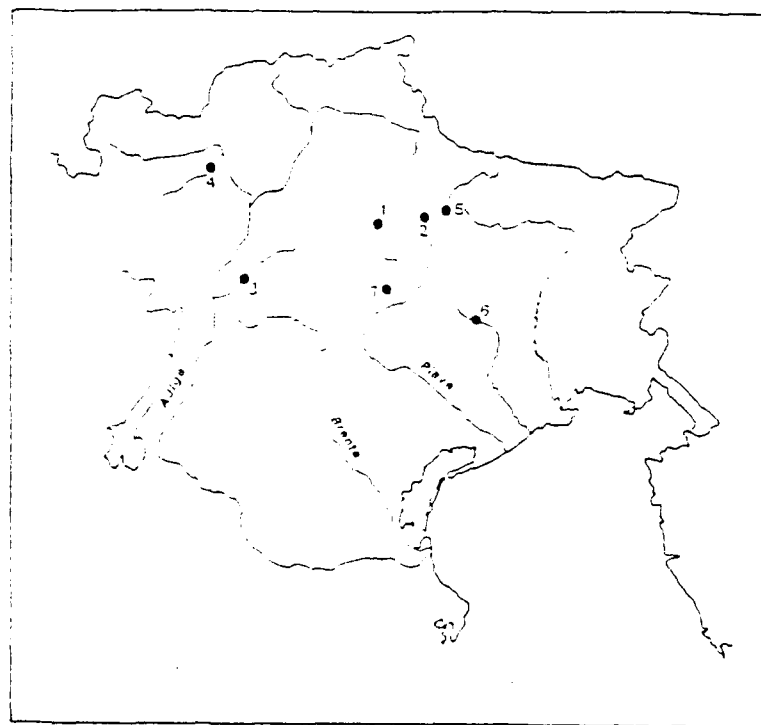| Basin	lake	Drainage basin	Filling in m3k2/yr
1 Cordevole	Alleghe	246	171 (1933-1943)
2 Boite Piave	Vodo	353	118 (1958-1976)
3 Avisio (Adige)	Stramentizzo	740	130 (1951-1970)
4 Valsura (adige)	Alborello	190	149 (1953-1970)
5 Piave	Pieve Cad.	690	775 (1946-1974)
6 Cellina	Barcis	388	875 (1955-1972)
7 Canale	Mis	109	4000 (1963-1971)

Fig. 5 Location of the mentionaled lakes

With respect to the others, the average value of the transport of solid material seen in the Missiaga basin, is decidedly modest. While keeping in mind the temporal dissimilarity og the observation period, the figure relating to the Mis basin may seem surprising. Howerer, the geomorphological situation of this basin, characterized by a narrow valley sole and sides covered by detritus is, from certain point of view, similar to the morphological situation of the Missiaga-Bordina basin in the past, thus lending it credibility.

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C. FRIZ, L. GIADA, V. VILLI

LA VALUTAZIONE QUANTITATIVA DELL'EROSIONE IN UN BACINO IDROGRAFICIO:ASPETTI METODOLOGICI E OPERATIVI

PREMESSA

Dal 1983 sono in corso osservazioni idrologiche e geomorfologiche nel bacino attrezzato sdel rio Missiaga situato nello dolomiti bellunesi, in sinistra idrografica del Cordevole (fig.).

I risultati finora conseguiti, soprattutto per quel che riguarda il trasporto solido, sono stati incoraggianti, essendosi potuto definire delle relazioni, sia pur empiriche tra quantita' del materiale trasportato (in m³ e parametri idrologici significativi come ad esempio le portate al colmo di piena ed i volumi del quick flow.

Il limite pero' di questi risulatati e' costituito dalla brevita' del periodo di osservazione che non consente estrapolazioni temporali e dalla tipicita' fisiografica del bacino propria dell'area montuosa dolomitica.

La consapevolezza di tali limiti di validita' tuttavia, ha fatto si che di pari passo con la acquisizione di dati relativi ai volumi di materiale di volta in volta asportato dal bacino da singoli eventi di piena si iniziasse una ricerca avente per obbiettivo la valutazione quantitativa indiretta

1

dell'erosione.

Tale argomento presenta una certa complessita' dovendo essere trattato caso per caso in funzione della evoluzione morfologica subita dal rilievo e della possibilita'di poterla tradurre graficamente (PELLEGRINI, 1984).

Dal punto di vista operativo infatti la valutazione quantitativa indiretta dell'erosione si risolve calcolo del volume compreso tra la superficie topografica attuale ed una superficie di riferimento ricostruita relativa ad un momento prefissato dell'evoluzione morfologica dell'area in studio.

La collocazione temporale dunque della superficie di riferimento rappresenta un elemento di primaria importanza che deve necessariamente scaturire da attente considerazioni di carattere geomorfologico.

VALUTAZIONE QUANTITATIVA INDIRETTA DELL'EROSIONE NEL BACINO MISSIAGA-BORDINA

Nel caso del bacino Missiaga-Bordina si e' ritenuto che la stima indiretta glabale dell'erosione non potesse decorrere che dal tardoglaciale.

In quel periodo infatti dopo i diffusi fenomeni di collasso seguiti al ritiro del ghiacciaio del Cordevole, sono iniziati i processi erosivi che hanno portato all'attuale configurazione morfologica del bacino. Inoltre, dal punto di vista operativo tale scelta consente di poter prescindere dal dover specificare la natura e definire l' intensita' dei processi erosivi che hanno operato nel bacino, essendosi

2

verificati in condizioni climatiche ed ambientali verosimil-
mente mutevoli.

Per definire nel modo piu' rigoroso l'intervallo di tempo
entro il quale realizzare il bilancio globale dell' erosione,
del bacino Missiaga-Bordina, ci si e' avvalsi tuttavia anche
delle datazioni cronologiche assolute fornite da analisi C^{14}
su resti di tronchi d'albero in posizione di crescita ancora
presenti nel greto del torrente Cordevole. Questi resti af-
fiorano infatti in un tratto d'alveo dove, uan conoide costi-
tuita da materiali provenienti dall'enorme ammasso detritico
stabilitosi nella valle Missiaga-Bordina, ha sbarrato il
corso del Cordevole, dando origine al lago di Agordo.

Tali avvenimenti sulla base delle citate datazioni assolute
si sarebbero verficati 5900 anni a.C., quindi all'incirca
nel preboreale superiore e prima dello stadio di Daun (VENZO
S,1977).

La valle Missiaga Bordina nel Postglaciale

L'analisi congiunta dei dati bibliografici e di campagna ha
permesso di ricostruire con buon grado di verosimiglianza i
carattteri topografici del bacino del Missiaga-Bordina al
termine del wurmiano.

Le modificazioni morfologiche nel bacino nell'immediato
postglaciale sono state sostanziali in seguito ai grandiosi
fenomeni verificatisi e all'ingente trasporto ed accumulo dei
dteriti cosi' originatisi verso valle.

Sta di fatto che la depressione valliva tardowurmiana e'

3

stata via via invasa da una massa abnorme di materiali detri-
tici il cui volume risulta difficilmente quantificabile.

Cio' che invece e' stato possibile determinare e' la quanti-
ta' di detrito rimosso dall'accumulo e trasportata verso
valle. Sempre dai rilevamenti di campagna sono emersi dati
oggettivi per ricostruire con sufficente precisione la topo-
grafia del bacino in fase di massimo accumulo, cosi' da
consentire la realizzazione della planimetria di fig. che e'
stata cosi' assunta come superficie di riferimento origina-
ria per il successivo calcolo dei volumi asportati.

Calcolo del volume eroso

La quantificazione del volume compreso tra questa superficie
topografica l'attuale (fig.) e' stato effettuato con una
procedura informatica denominata DIGCON. Tale procedura rea-
lizza anzitutto una discretizzazione della zona esaminata.
Tutto cio' con la sovraimposizione ad essa di una griglia a
maglie regolari, nella quale la quota di ogni nodo e' ricava-
ta mediante un procedimento di interpolazione a partire dalle
curve di livello digitate.

Piu' in particolare detta griglia e' costituita dall'inter-
sezione dell'insieme delle curve di livello con due famiglie
di paralleli ad i ... graficati ... ...
interdistanze tra i profili definiscono ovviamente le dimen-
sioni della maglia elementare.

La quota in ogni nodo, come si e' gia' accennato viene

4

determinare per interpolazione lineare ... ...
due profili ortogonali che la individuano. Qualora i valori
calcolati non coincidano (puo' verificarsi quando nelle due
direzioni ortogonali le pendenze e/o le distanze tra le
intersezioni sono diverse) si calcola la media ponderata dei
valori determinati, usando come peso l'inverso della distanza
dalle intersezioni piu vicine nelle due direzioni ortogona-
li.

Le dimensioni della maglia , ovvero l' interdistanza tra i
profili, tenuto conto della particolarita' dell' elemento
morfologico da rappresentare e' stata scelta di 100 m.

Il confronto tra i modelli digitali attuale e postglaciale
infine ha consentito di stimare in corrispondenza ad ogni
maglia il volume del parallelopipedo avente per base la
maglia e per altezza la media aritmetica dei quattro valori
di differenza di quota nodali.

La somma algebrica dei singoli volumi ha fornito infine il
risultato complessivo risultato di 1,1 miliardi di m^3.

Tenendo conto che:

- l'area del bacino Missiaga Bordina interessata dal riempi-
  mento e' di 9.08 km^2

- che l'avvio della fase erosiva sul corpo del riempimento,
  sulla scorta dei dati forniti dalle analisi cronologiche
  C^{14} puo' essere fatto risalire a circa 8000 anni fa, ri-
  sulta' una perdita unitaria media annua di 15,143 m^3/km^2,
  valore che evidenzia in modo inequivocabile l'energia con

cui i processi erosivi hanno smantellato l'accumulo detritico
formatosi nella valle Missiaga-Bordina Un confronto tra detto
valore medio ed i dati inerenti il trasporto solido in corso
di acquisizione nella parte superiore del bacino Missiaga non
e' possibile, mancando il presupposto della omogeneita' delle
condizioni che li hanno determinati.

Le valutazioni volumetriche effettuate dal 1984 indicano in
ogni modo che nelle attuali condizioni geomorfologiche e
climatiche il trasporto solido nel bacino in studio ammonta
a circa 120 m^3/km^2/anno. Si tratta di un valore unitario
che e' del tutto congruente con molti dei dati forniti da
CATI (1981, vedi tabella) dedotti dall'interrimento di laghi
naturali ed artificiali alpini.

Bacino	Nome del lago	Bacino km^2	Interrimento m^3/km^2/anno
Cordevole	Alleghe	246	171
Boite(Piave)	Vodo di Cadore	353	118
Avisio(Adige)	Stramentizzo	710	130
Valsura(Adige)	Alborelo	190	149
Piave	Pieve Cadore	690	775
Cellina	Barcis	388	875
Canale	Mis	109	4000

Rispetto ad altri invece il valore medio del trasporto
solido riscontrato nel Rio Missiaga appare decisamente mode-
sto. Pur tenendo conto della inomogeneita' temporale dei
periodi di osservazione, del tutto sorprendente puo' appa-
rire il dato relativo al bacino del Mis. La situazione geomo-
rfologica di questo bacino tuttavia,caratterizzata da un
fondovalle stretto e versanti ricoperti da detriti richiama,

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per certi aspetti, situazioni morfologiche pregresse del
bacino Missiaga-Bordina, rendendo cossi' credibile il dato
fornito.

CAUSES AND NATURE OF RIVER PLANFORM CHANGE

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Introduction

Research on changes in planform of gravel-bed rivers over the past 20 years has done much to elucidate their dynamic nature and characteristics. Even so, the assumption still seems to be prevalent that, given stability of conditions, meandering channels, after an initial development phase would show an equilibrium of form as measured by standard meander parameters. Various models have been produced (e.g. Brice 1974, Hooke and Harvey 1983, Keller 1972, Kondrat'yev 1968) showing evolutionary sequences in meander bends, even so, where trends in characteristics such as sinuosity are detected over long periods of time, the channel system is suggested to be in a transient state (Hickin 1983). Although conceptual frameworks (eg. Schumm 1979, Brunsden and Thornes 1989) incorporate such transience and the existence of intrinsic thresholds seems to be accepted, there is still a tendency to search for triggers to phases of change. That the most active meanders may show evolution of characteristics is not taken as a basic premise.

This paper aims to examine the evidence for the extent of autogenesis in meander change by discussing the hypotheses on causes of meander changes and the expected responses from each, examining how the evidence helps to substantiate or eliminate any of these hypotheses and discussing whether it is indeed possible to distinguish one cause from another.

Possible causes of change

Four major explanations of change can be identified and these can be divided into two major groups of causes which may be termed allogenic and autogenic (Lewin, 1977). Under the allogenic causes are human activities, particularly landuse and climate. Under autogenic are intrinsic thresholds and chaos theory. On the basis of theoretical models, hardware models and experiments, and from field studies, the expected responses can be identified though, of course, these are rarely very specific because of the problems of indeterminancy and number of variables in the system, as highlighted by Hey (1978). Schumm (1969) long ago identified the likely direction of change of major channel variables in response to changes in discharge or sediment load.

Anthropogenic changes:
A wide range of human activities can affect stream channels, both directly and indirectly, and can alter either the discharge regime or the sediment load or a combination. Even the alteration of discharge may be complex and several flow parameters may be needed to characterise the change in magnitude, frequency and occurrence of flows. The combination of water and sediment alterations may be even more complex and may be

difficult to predict yet the relative balance and timing of discharge and sediment alterations will have a crucial effect on the channel response. Nevertheless, general tendencies in behaviour can be recognised. For any specific alteration, it is expected that the main impacts occur downstream though some effects can then work back upstream eg. a steepening or lowering of the channel in the form of a headcut. This can lead to a complex response over time for any specific location. The impact will vary with the spatial distribution and extent of the human activity eg. a landuse alteration may be widespread and relatively synchronous or it may be more patchy. Urbanisation will tend to be localised.

Climatic change:
Similarly, the effects of precipitation and temperature can lead to complex combined effects on runoff regime, especially through the intermediary of vegetation. The effects on the channel would be expected to be widespread in the catchment unless different parts of a large catchment have rather different climatic regimes.

Both of these types of allogenic alteration lead to modifications in the basic force element of the system ie. the discharge. In terms of response, the resistance factor of the varying channel characteristics along the length of the stream must then be taken into account. Obviously, change in planform will only take place if the banks are erodible and less resistant than the bed. It can be hypothesised that, given a change such as urbanisation in a catchment and assuming that it has altered the discharge regime downstream, the channel should adjust. The different types and scales of adjustment may then provide an indication of the sensitivity of various channel sections to such alterations. A declining effect on runoff regime with distance from the urbanised area and as its proportion of the catchment alters may be expected and thus a similar response in terms of channel change. Departures from this, eg. where individual reaches show markedly different responses, may then be attributed to local site factors placing constraints on the types of adjustment possible. Basically, the response from climatic change would be expected to be much more widespread and more synchronous than that to landuse and land practice alterations (cf Cooke and Reeves, 1976).

Temporal coincidence of activities is no proof of a cause and effect link since there may be a lag in response, the change may bring the system nearer the threshold for response but further slight alteration may push it over. However, if an alteration such as landuse or climate was the basic cause of change then some associated phasing of the channel response may be expected. Differences in temporal pattern might also be expected with much greater synchroneity in the case of climatic causes. However, the rate of propagation of effects and the problems of complex response need to be considered.

Intrinsic thresholds:
If channel changes are in response to intrinsic thresholds in the channel system then some association of zones and types of change with the spatial distribution of physical characteristics should be identifiable. The situation is complicated, however, by changes in those physical characteristics over the longer-term either because of the channel shifting and thus, for example, encountering different materials, or as a result of the channel behaviour itself eg. deposition altering slope in the longterm. Again the

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situation is complicated because of the number of variables which may alter but certain key ones have been identified from experiments and case studies. The crossing of these thresholds can themselves lead to spatial propagation of the effects and complex response. However, given good enough historical information and rapid enough changes it should be possible to identify the spatial and temporal patterns of change and elucidate the dynamics.

Chaos theory:
This suggests that the behaviour of systems is markedly non-linear and very sensitive to initial conditions. It is therefore difficult to relate the behaviour of a system to physical determinants but if the behaviour over time is analysed then certain patterns and consistency of behaviour become apparent.

Evidence

Historical data at a number of different dates over a reasonably long historical period for a number of river channels may allow us to elucidate some of these patterns. The spatial distribution of changes needs to be examined at a number of different scales and the temporal sequence for each reach needs to be analysed. A number of methods have been used to provide suitable evidence.

1. To identify the distribution of channel changes over wide areas and to investigate whether there is any kind of regional or general spatial pattern, the courses of channels of rivers have been compared at two dates, c.1870 and c.1970. This has been done for a sample of streams in lowland England and for most streams draining from upland England. Changes already identified in the literature have also been collated. Only those changes greater than one channel width and therefore well above any possible map inaccuracy and only those which are not directly due to channelisation of the stream have been included.

2. The changes along particular streams and in particular reaches, especially the most active reaches have been compiled in greater detail, first by comparison of O.S. maps at two dates at 1/10560 then by compilation of maps at 1/2500 for several dates.

3. To test the relationship between changes and possible physical determinants, physical characteristics have been measured through stable and unstable reaches on certain rivers. From previous work it is hypothesised that the changes are not related in a simple manner to conventional factors and therefore complex combinations and patterns of variation have been examined.

4. It is recognised that physical determinants at the reach scale such as bank materials may not be the only determinants and that precise quantification of all characteristics may not be possible or suitable so a classification of various catchment, reach and site characteristics has also been carried out. The aim of this work is to identify constraints on channel change as much as causes of instability.

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5. The fifth type of procedure is to produce time plots and phase diagrams of the channel behaviour to examine whether patterns are present which are not otherwise apparent and which it may not be possible to tie in specifically to physical determinants.

Distribution of changes

National scale:
Figure 1 is a map of the planform changes found from map comparisons, as described above. This shows some interesting distributions. First, it can be seen that almost all the changes are in zones fringing the uplands. In the reconnaissance survey it was soon found that most channels in the lowlands were stable or had been channelised. Some may have changed slightly but shifts of channel location were below the channel width threshold for significant change. This is not to say there are no significant changes at all in the lowlands.

The reaches which have shown marked change over the past 100 years are mostly in the so-called 'piedmont' zone (Newson, 1982). The channels in these reaches have certain characteristics: they are downstream of the bedrock reaches of the uplands and are flowing in valleys with a floodplain formed from alluvium; the material tends to be highly erodible because it is still relatively coarse, with a high gravel and sand component, compared with greater clay content farther downstream; the gradients are sufficiently steep for stream power to be high but the streams are large enough to have developed a regularity of form as portrayed by meanders and pools and riffles.

However, active lateral movement does not occur on all streams in the piedmont zone and the distributions should be examined in relation to the hypothesised causes of change. If the changes were due to climatic change alone then changes might be expected everywhere or at least widespread in a region. The fact that the changes are much more localised but tending to be in a certain zone implies that there are constraints on change and that thresholds are operating.

If the changes are a response to human activities then one would expect a close relationship in spatial pattern. Very many of the basins have been affected in some way by landuse change, urbanisation, reservoirs or water abstraction. Analysis of the type of change might then show if the response was indeed in the direction expected from such changes and this will be examined below. The situation is complicated with multiple human modifications in many basins. Examination of the timing of changes may help to see if there is any association. Turning the argument round though, many case studies have shown the types of changes to be expected from specific basin and channel alterations. Perhaps we should rather ask why have certain channels not changed in planform when they have been affected by major human activities such as urbanisation? Again this points to the operation of strong thresholds and may allow us to identify reaches vulnerable to planform adjustment and other reaches which will adjust by other variables. This has important implications for channel management and impact prediction.

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Basin scale:

Various patterns of distribution of planform alteration can be seen along the channels and these relate to degree of activity (rate of change) and type of change. On some rivers there are long reaches that are stable then long reaches which are markedly unstable (Fig. 2). One of the striking aspects is that there is little transition between the two, with completely stable reaches adjacent to very unstable reaches. Again this points to the operation of thresholds but that once the thresholds are crossed channel pattern then alters very rapidly.

The second type of distribution is where reaches, including the whole path round bends, are affected but there is a series of short unstable reaches separated by stable ones. This implies some specific site controls and measurement of characteristics down though such sections should allow those controls to be identified. The system must be near the threshold for instability. The River Culm in Devon exhibits such a distribution (Hooke, 1977).

The third type of distribution is where whole reaches are not changing but rather erosion is confined to meander apices and change overall is very limited. Rates of erosion in such reaches are usually low. Such sites may allow the identification of critical force-resistance relations in specific terms.

Types of change

An important aspect of the distribution of changes is the differentiation of types of change. It has already been indicated that certain theoretical frameworks predict distinctive planform responses depending on the direction of alteration of runoff and sediment load. A major indicator of channel response and one which is easy to measure without encountering the difficulties of meander definition is sinuosity. Increase in peak discharge is expected to decrease sinuosity because of a tendency to increase meander wavelength and therefore straighten the channel. Decrease in discharge would have the converse effect.

If change is in response to a climatic modification then it might be expected that whole rivers and even whole regions would show the same general tendency or direction of type of change. Of course the effects of a climatic change do depend on the runoff response which varies with the lithological, soil, topographic and landuse characteristics of the area but the change should not be in different directions within a region. If a whole catchment is tending to show a particular type of change but this differs from that of neighbouring catchments then this would imply that the impetus to change is confined to that catchment. This points to landuse change as being a significant factor.

Where different reaches of the same channel are behaving in contrasting manner then other explanations need to be found. It is possible that the different reaches are in fact related to some different landuse changes along the channel eg. afforestation in the uplands, then urbanisation farther down, but many examples are found where reaches close to each other in the same type of catchment are exhibiting different types of change. This implies that the changes may be part of the same phenomenon, simply at

different stages or that the period of evidence simply captures a particular phase of activity but that trends would be different if a longer period and more evidence was available.

58 rivers in England have been classified according to whether they show a net increase, net decrease or no change in sinuosity over the past 100 years. Of the 58 rivers, 22 shown an increase, 23 a decrease and 13 no change in sinuosity. Little regional differentiation is present except a rather greater proportion increasing or not changing in sinuosity in the NE and E Pennines and more decreasing in sinuosity in the Welsh Borderlands.

To understand the changes in more detail the types of change need to be classified. Changes on individual bends have also been classified according to the typology in Figure 3. Analysis of historical sequences from large-scale maps and aerial photographs on several rivers, but particularly the Dane (Hooke and Harvey 1983) has revealed a common sequence of meander development and this is presented as a graphical model in Figure 4 (Hooke 1989).

Examples of the different types of distribution can be found. For instance, taking five active meandering rivers in SE Devon and classifying the type of movement on bends, it is found that some whole rivers are behaving differently from others. The Creedy is predominantly migrating; the Otter, Axe and Yarty are growing and the Culm is mixed in type. In NW England an example of apparently contrasting behaviour can be found by comparing the neighbouring rivers of the Bollin and the Dane. According to Mosley (1975) a reach of the river Bollin downstream of Wilmslow has shown a marked decrease in sinuosity since about 1935 with numerous cutoffs. A 14km reach of the River Dane has in the past 150 years shown an increase in channel length of 10% with growth being an important component of change on individual bends.

Temporal distribution

If the historical evidence shows that change began at a particular time then this implies that sometime earlier (how much depending on the lags in the system) there had either been an alteration of the inputs to the channel system or that an intrinsic threshold had been crossed. The latter idea may be particularly difficult to test since there are problems in reconstructing conditions, eg. local channel slope or bank materials, at that time. Even the former hypothesis poses problems of investigation since it is notoriously difficult, though not impossible, to reproduce land-use patterns on a catchment basis, let alone land-use practices. There are cases where channel change seems to have begun at a particular period eg. on the Bollin (Mosley, 1975), Skirden Beck, Lancashire (Thompson, 1984). In the case of the Bollin, Mosley investigated a number of possible causes including rainfall change and concluded the changes were probably due to urbanisation and possibly agricultural drainage due to the spatial and temporal coincidence of patterns. Further evidence on changes elsewhere on the Bollin and alternative explanations of the changes will be put forward later.

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What is remarkable when many of the channel changes in Britain are examined, particularly those in the least modified catchments is that channel development seems to have been taking place from the date of the earliest evidence. In some cases this is 1840 and several channels in different parts of the country show evidence of a much lower sinuosity channel at that time. That this is not an artefact of the maps or due to channel straightening is confirmed by field evidence of the courses. This general trend of increasing planform development is substantiated by the evidence of the floodplains and terraces. In several of these valleys the floodplain is of restricted width and does not occupy the whole valley floor. It is bordered by alluvial terraces but these are highly erodible. Almost all the floodplain can be dated to channel movement within the period of map evidence and the field evidence strongly supports the idea that these are first generation floodplains. The floodplains are bounded by distinct and fresh looking terraces. In the case of the Dane the base of the depositional sequence has been dated to 4500-1200BP (Hooke et al 1990) but this is succeeded by a major aggradational phase forming the upper part of the terrace deposit. The origins of this are still unclear and there is some evidence which points towards medieval forest clearance and increase in agriculture as being the major cause. However, the increasing evidence that several valleys may be showing the same sequences and the increasing number of dates showing a high degree of coincidence (Richards, 1987, Macklin, 1989) may point towards a greater influence of climate change than has hitherto been invoked in Britain. Further work on the sequences in different parts of Britain is vitally needed and the work such as that of Macklin (pers. comm.) on the Little Ice Age is of crucial importance.

Environmental and physical characteristics

The aim of this part of the research has been to investigate whether planform changes and subdivision of type seem to be associated with particular conditions at the basin, valley and reach scale. If it is not possible to find common factors it may at least be possible to identify constraints, factors that prevent planform change (eg. bedrock gorges).

The problem with much analysis of physical controls on channel behaviour is that it often ignores the environmental factors because of their scale and also ignores many potentially important influences because they are so difficult to quantify. A classificatory approach has therefore been taken and the analysis is based on presence or absence of particular features and their combinations. Forms were devised, based on the work of Kellerhals et al (1976) and the categories included are listed in Table 1a. For each category or characteristic a choice of codes was available eg. terraces (Table 1b). Where there may be a mixture of characteristics up to three choices were allowed given in order of proportion eg. valley vegetation (Table 1c). These forms were completed in the field for 30 different reaches of actively shifting channel on 11 rivers. Of course, this only gives an indication of present conditions.

Techniques for analysis of this type of data are still fairly limited but certain associations became apparent. Channels have been classified into stable, slightly unstable, moderately unstable and highly unstable with respectively 8, 7, 4 and 11 in those classes. A strong association appears between degree of instability and type of pattern. Stable

channels tend to be slightly sinuous or irregular in pattern whereas all highly unstable channels are regular meanders or tortuous. No stable but well developed meanders were found and likewise the less developed patterns did not seem to be altering rapidly. This is not as obvious as it might seem and it raises the question of how low sinuosity and irregular patterns make the transition to unstable tortuous patterns. Another association which emerged clearly and is not as obvious as it might seem is that there is a strong positive relationship between degree of lateral instability and number and continuity of terraces. If lateral erosion is so rapid then it might be expected that terraces are not acting as a constraint on lateral development. What it does imply is that these reaches

already shown that the most complete sections of terraces are likely to be in these piedmont zones due to burial and height range downstream. However, the association between recent channel change and the longer-term development merits further work. The classification also shows that stable reaches occur not just where valleys are restricted but in wide valleys as well. Highly unstable reaches are mostly in wide valleys but they also have a number of constrictions. Changes are strongly associated with the most erodible bank materials as might be expected but again the classes are not exclusive. Composite banks of silt overlying either gravel or cobbles is very common on the highly unstable reaches but do not occur on any of the stable channels. Again, as would be expected, stable channels have good or very strong bank vegetation but highly unstable channels have no banks in these categories of cover. However, vegetation can be a response rather than a cause.

Reach scale:
Preliminary analysis of the distribution of channel changes had indicated that there was not a simple relationship to controls conventionally advocated, eg. slope, bank materials. The aim was to investigate whether more complex and subtle combination of conditions were acting as thresholds to change. Various physical characteristics have been measured in the field down through stable and actively changing sections. Conventional physics demands that erosion is dependent on force on the bank at a point exceeding

has been to look at absolute force of resistance eg. power or % silt-clay. However, the force-resistance threshold may be crossed by combinations of factors varying spatially in a very complex way.

Measurements have been made on sections of the Rivers Teme, Arrow and Lugg in the Welsh Borderlands but clear patterns have not yet emerged. One of the reasons for lack of association in characteristics is that the change is a phenomenon acting over time whereas these factors are measured spatially at one point in time. It can therefore be suggested in the model that threshold conditions for change may have been reached at some location within the reach sometime in the past but the changes have since been propagated both upstream and downstream. Constraints then act at a lower level to prevent change and restrict the active reaches spatially. However, the abruptness of the changing reaches and their patterns of distribution imply that channels are very sensitive to alterations in conditions.

8

One of the remarkable aspects emerging from the work on sequences of change in meanders is in fact the consistency of change in different environments and the high degree of systematic change in situations where materials, flow patterns, etc. are varying. However, it has been shown that it is remarkably difficult to relate the channel behaviour to physical determinants.

Much of the historical evidence shows that once certain thresholds, which remain difficult to identify, are crossed the channel instability then takes off. The channel becomes increasingly unstable and there seems to be a positive feedback effect. This can be seen both at the reach scale and the individual bend scale. The model of meander development described earlier shows a take-off point when rapid growth suddenly occurs. This appears to have some relationship to form (Hooke and Harvey 1983), as depicted by path length and curvature.

Analysis of the evidence of time sequences of meander changes shows that apparently different types of behaviour may be part of an autogenic sequence. This is particularly well illustrated by a reach of the River Bollin (Fig. 5). At present in a reach of this active and freely meandering river, upstream of Wilmslow, cutoffs are occurring and if the changes over the past 10-20 years are measured then a net decrease in sinuosity is seen. However, if the longer-term sequence from 1840 is compiled then it is seen that there has been a gradual increase in sinuosity until recently. It is also seen that the rate of change has increased. Cutoffs have taken place in recent years as the bends have reached the stage of development such that they intersect. The greater the rate of erosion then the sooner this point is reached. The neighbouring catchment of the River Dane which had apparently contrasting behaviour over the period 1840-1980 is now showing a similar trend with cutoffs occurring. Once a cutoff occurs then this makes room for meander development to take place again and this will go on until another cutoff occurs. Therefore once this phase is reached sinuosity may shown an oscillating pattern over time. This is supported by evidence from the Welsh Borderland rivers. The River Teme is the most active channel and has an apparently chaotic sequence of changes but if sinuosity over time is tabulated then an oscillatory tendency is seen. That the crucial threshold may be related to sinuosity is seen by the association between activity and sinuosity (Table 2). This also emerged from the coding classification discussed earlier on a wider range of channels where there was a close association between degree of instability and type of channel pattern. It also emerged from that analysis that all the highly unstable reaches in that sample showed a net change in sinuosity and that, although increases in sinuosity appeared in all classes of instability, decreases in sinuosity are mostly in the highly unstable class. This indicates the trend for cutoffs, high instability and tortuous patterns to be the end of a sequence.

Work by Hooke (1987) was an intuitive attempt to analyse morphology relations, and time paths in phase space though not placed explicitly within the chaos and dynamical theory framework. Indeed the data there could be reinterpreted as supporting chaos theory in that similar trajectories are seen but they differ according to the starting point and physical conditions in the bend. In Figure 6 there is a suggestion of an attractor.

The accelerating activity of the channel as sinuosity increases may provide an explanation for the increasing rate of erosion found on Devon rivers (Hooke, 1977, Hooke and Kain, 1982). Although a wide variety of possible causes was investigated none seemed to provide an adequate explanation. It now appears that it may simply be part of the inherent trend of meanders.

Although it is hard to move away from the physical determinacy inculcated in us by conventional physics, explanations based much more on the recognition of non-linear behaviour, positive feedback and of systematic behaviour in apparently different circumstances may offer an explanation of some channel changes. It is already well established that the relationship between rate of erosion and channel curvature is markedly non-linear (Nanson & Hickin, 1983, Hooke, 1987). Ultimately we should still be able to relate these changes to what is happening in terms of force and resistance at any point in the bend but this may be so difficult to measure and model over the 3-d form of a bend let alone a series of bends that the ideas of chaos theory provide a viable alternative approach.

That this type of explanation may be adequate in itself is supported by some computer simulations of meander development. For example, Ferguson's (1984) model is based simply on a non-linear relation between rate of erosion and curvature, with no process basis directly, and yet it replicates some of the changes seen in nature.

Conclusions

The importance of an historical perspective and analysis of time paths in understanding and explaining river behaviour has been demonstrated. There is a need for analysis at different spatial scales and for both an environmental and process-based approach. Undoubtedly human activities have had an impact on river channels but the possibility of profound influence of climate within the historical period even in Britain should not be rejected. Superimposed on this is the strong evidence for autogenesis in meander development which leads to particular sequences of form and phases of activity even if other causes provide the impetus. Physical conditions provide at least some limits on planform development. The actual distribution of changes is a complex mosaic of these different patterns but the degree of autogenesis may be greater than is often recognised.

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Table 1

Variables in classification database of reach
and channel characteristics

		Fieldname	Variable description
River data			
	1	RIVER	name
	2	AREA	region in England
	3	PERCENT	% of reach length shifted course
	4	TERRAIN	type of terrain
Valley reach			
	5	VALLEYVEG	vegetation type
	6	VALLEYTREE	type of trees
	7	VALLEYUSE	valley landuse
	8	GEOLOGY	type of deposits
	9	BEDROCK	type of bedrock
Valley wall			
	10	WALLVEG	wall vegetation
	11	WALLTREE	type of trees
Terraces			
	12	TERRACES	extent, none-continuous
	13	LEVELS	how many terrace levels
Channel-valley relation			
	14	VALLEYTYPE	classification of type
	15	NOVALLEY	fan etc.
	16	UNDERFIT	yes/no
	17	CONSTRICT	lateral constriction
	18	VERTICAL	upgrading/degrading
	19	LATERAL	relationship to valley walls/ high terraces
Valley flat			
	20	VALLEYFLAT	presence of
	21	EXTENT	width
	22	FLATVEG	vegetation
	23	FLATTREE	tree type
	24	FLATUSE	valley flat land use

13

Table 1 (Continued)

Channel		Fieldname	Variable description
	25	PATTERN	channel pattern
	26	ISLANDS	presence of +type
	27	BRAIDING	braiding index
	28	FLOW	type of flow
	29	BARS	type
	30	BELTWIDTH	from map data
	31	WAVELENGTH	from map data
	32	SINUOSITY	from map data
	33	OBSTRUCT	natural destruction's in channel
Lateral activity			
	34	DEGREE	degree of obstruction
	35	ACTIVITY	lateral activity
	36	STABILITY	lateral stability
	37	BENDS	changes on bends
	38	SINCHANGE	changes in sinuosity
Channel banks and bed			
	39	ALLBANK	alluvial bank material
	40	NONBANK	non-alluvial bank material
	41	HEIGHT	of banks
	42	VEGBANK	vegetation cover
	43	BED	present of bedrock in channel bed
Bed rock			
	44	BASEROCK	predominant bed material
	45	BASETYPE	rocktype at channel base
	46	ERODE	erodibility

Table 1b	Table 1c
Coding for Terraces	Coding for Valley Vegetation

Terraces

Vegetation:

Terrace presence:

		0	0	0	not applicable
0	none	1	1	1	almost none
1	indefinite	2	2	2	grass
2	fragmentary	3	3	3	shrubs
3	continuous	4	4	4	individual trees
		5	5	5	sparsely forested (0-25%)
		6	6	6	moderately forested (25-75%)
		7	7	7	heavily forested (75-100%)
		8	8	8	swamp or marsh
		9	9	9	crops

Table 2

Sinuosity of Stable and Changing River Reaches
on Two Rivers

River	Degree of Change	Sinuosity of Reaches
Teme	Stable	1.15, 1.2, 1.0, 1.18
	Migrating	1.2, 1.19, 1.22. 1.07, 1.06
	Extensive change	1.4, 1.42
	'Chaotic' change	2.04
Arrow	Stable	1.1, 1.18, 1.2
	Isolated change	1.41
	Extensive change	1.53

Figure Captions

Figure 1   Survey of river planform changes in England and Wales.

Figure 2   Example of stable and unstable reaches, River Clywedog, Clywed, N. Wales.

Figure 3   Types of meander change (Hooke, 1984).

Figure 4   Model of meander change (Hooke, 1989).

Figure 5   Sequences of changes on the River Bollin, near Wilmslow, Cheshire, N.W. England.

Figure 6   Rates of movement plotted against sinuosity showing sequences of movement for individual bends (Hooke, 1987).

Migration

Confined
migration

Growth

Lobing

Double heading

New bends

Retraction

Cut off

Complex changes, island formation,
abandonment and small irregular
movements

Stable bends, no change

Fig. 3

Initiation

Migration

Growth

Double-
heading

Riffle

Pool

Erosion

Primary
flow

Fig. 4

1850

1871

1907

1970

1989

0   100m

Fig. 5

Stable
Migration
Growth
Retraction

Maximum rate of movement in yr -1

Sinuosity

Fig. 1

Rivers studied in the authors'
O.S. 6″ survey

Changes identified from authors'
O.S. 6″ survey

Changes studied in the literature

100 km

0   50 m

Fig. 2

HISTORIC FLOODS AND VERTICAL ACCRETION
OF FINE-GRAINED ALLUVIUM IN THE LOWER TYNE VALLEY,
NORTH EAST ENGLAND

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Abstract

Historical documentation of floods together with sedimentological and trace metal analysis of fine-grained flood sediments have been used to construct a flood record for the River Tyne a gravel-bed river in northern England over the last 100 years. Upsection changes of trace metal concentrations in vertically accreted alluvium have been related to lead and zinc ore production in the Tyne basin and used to date flood events. Exceptionally high (by British standards) vertical accretion rates (2.4 cm a^{-1}) in the lower Tyne Valley allow individual flood events to be recognised and dated. However, significant differences between documentary and sedimentary-based estimates of the frequency of inundation emerge when these two records are compared. This is shown to be the result of bias in archive sources toward recording large overbank events and the censoring of the alluvial stratigraphic record, as the result of variations in sediment supply and channel entrenchment, to reflect progressively less frequent and larger floods.

INTRODUCTION

The paradox of flood prediction is that, as the need for accuracy increases (e.g. where sensitive areas need to be protected against flooding of a high recurrence interval) the database provided by hydrometric recordings of river discharge becomes proportionally smaller in relation to that interval. Considering the relatively recent introduction of scientific hydrometry, hydrologists in most regions of the world have turned to documented and monumented historical records (McEwen, 1987) as well as rainfall-runoff models (Anderson and Burt, 1985) to help improve prediction. Arguably the most appropriate, and certainly the most generally available means for extending the flood series beyond the period of instrumentation, however, is the stratigraphic record of alluvial deposits in river valleys (Costa, 1978). The inherent advantages of using alluvial sediments for investigating flood histories in river basins are essentially two fold: first they are not limited by climatic regime or location. and second provided that sediment units can be dated. they can be used over any time period. Palaeoflood hydrology (cf. Stedinger and Baker, 1987), utilising procedures from geobotany, sedimentology, stratigraphy and geomorphology, is particularly useful for estimating the magnitudes and frequencies of rare and large floods, which is often very difficult using conventional flood frequency analysis.

For more than a decade North American geomorphologists have been successfully employing alluvial stratigraphic analysis in flood reconstruction over Holocene and historical timescales both in semi-arid (Webb et al., 1988) and humid (Costa, 1974; Knox, 1987a) regions of the USA. In Europe, however, although extensive use has been made of documentary sources to extend the flood record beyond the period of gauging (e.g. Decamps et al., 1989), geomorphological investigations of floods have by and large focused on the impact of single "extraordinary" events which usually attract attention by virtue of spectacular floodplain erosion and deposition (Acreman, 1983; Brown, 1983; Carling, 1986; Harvey, 1986; McEwen and Werritty, 1988, Newson and Macklin, 1990). To our knowledge no investigation has used the alluvial record to reconstruct historic flood histories in a major British or, for that matter, continental European watershed. In the study reported here historical reports of floods together with sedimentological analysis of textural discontinuities in fine-grained. vertically accreted alluvium are used to construct a flood record of the River Tyne in northern England over the last 100 years.

STUDY AREA

Flood histories

Following a major flood on the 26th August 1986 bank erosion at Low Prudhoe (NZ 088637) (Fig. 1) in the lower Tyne Valley, 15 km west of Newcastle upon Tyne, revealed extensive sections in the floodplain of the River Tyne. Floodplain sediments were found to be composed of 2-3m of finely-laminated sands and silty sands overlying 2m of sandy gravels. Preliminary granulometric analysis of prominent coarser sand layers in the upper find-grained alluvial unit showed them to have similar physical properties to sands deposited overbank on the Tyne floodplain during the August 1986 flood (Macklin and Dowsett, 1989). This suggested that sediments at this site could provide a valuable record of recent major floods in the lower Tyne Valley.

The River Tyne is a gravel-bed river with mean and maximum recorded discharges of 44 m^3S^{-1} and 1497 m^3S^{-1} measured at Bywell (NZ 038617), 5 km west of Low Prudhoe. Although the first gauging station in the catchment was completed as early as 1939 (Barrasford NY 920733 in the North Tyne), systematic instrumented records of river flow are only available after 1950. Although unusually good exposure of alluvium at Low Prudhoe initially prompted sedimentological and

1

stratigraphic based studies of recent floods in the lower Tyne Valley, floodstones on the Rectory steps at Ovingham (Fig. 2) marking the heights of the great floods of 1771 (the largest on record) and 1815, together with good archive records (newspapers, local books and journals, meteorological publications) of major floods in the Tyne catchment since 1699 (Archer, unpublished), make Low Prudhoe an excellent site at which to compare documentary and sedimentary flood records.

River channel change and vertical accretion processes

Over the last 130 years or so, the thalweg of the Tyne at Low Prudhoe has moved very little (Fig 2a), but since the early 1950's its channel has narrowed appreciably most notably in a 500m reach downstream of Ovingham Bridge, within which the study section is located. Channel narrowing has been primarily the result of incision which elevated former lateral gravel bars above the level of the low flow channel and enabled them to be colonised by vegetation. These sites subsequently became the forms of fine sediment deposition resulting in the infilling of site channels and attachment of bars to the former river bank. Rates of incision and bed erosion increased considerably in the late 1950's and 1960's following gravel extraction from the river bed, and is still continuing today necessitating regular upgrading of the footings of Ovingham Bridge.

The well-bedded sands and silty sands exposed in the river bank at Low Prudhoe are believed to have been formed by vertical accretion. This deposit has a similar surface morphology (asymmetric in cross section with its highest point located adjacent to the channel, see Fig. 2b) and stratigraphy (see below) to shrub and herb covered alluvial benches located <u>within</u> the present Tyne channel which accrete fine sediment to the level of the contemporary floodplain. These are not levees or overbank splays but resemble alluvial benches described by Page and Nanson (1982) and Taylor and Woodyer (1978) in several Australian and Canadian rivers and defined as, "actively accreting flat-topped bodies of sediment occurring along the banks of a stream channel" (Woodyer et al., 1979). Initial sedimentation at the study section was therefore under sub-bank full flow conditions with fine sediment deposition confined within the channel banks. Later sediments, as the result of progressive vertical accretion, would have tended to be deposited under flow conditions closer to bankfull and also by overbank floods. Thus a considerably proportion of fine-grained alluvium at Low Prudhoe represents the vertical component on within-channel sedimentation and is of a somewhat different origin to "overbank fines" (usually of silt and clay size and deposited some distance from the channel) more frequently described on British floodplains (eg Lambert and Walling, 1987).

Geology and metal mining history

The Tyne upstream of Low Prudhoe has a drainage area of 2198 km^2 and is fed by two major tributaries (Fig. 1); the River South Tyne (drainage area 800 km^2) that drains the northern flank of the Northern Pennines, and the River North Tyne (drainage area 1118 km^2) whose catchment comprises the Newcastle Fells and southern part of the Cheviot Hills. Both sub-catchments are primary underlain by sedimentary rocks (sandstones, limestones and shales) of Carboniferous age though small outcrops of igneous rocks (Whin Sill) are found in the lower reaches of the North and South Tyne Valleys and also in the headwaters of the River North Tyne (Cheviot granite).

The predominance of sandstone rock types within the Tyne catchment, both in pre-Quaternary lithologies and Pleistocene glacial deposits, results in a large sand component in Tyne flood sediments (Macklin and Dowsett, 1989; Newson, 1989). A high rate of "run-out" of fine-grained sediment especially from the Northern Pennine Uplands during major floods supplies abundant material for overbank

deposition on floodplains in the lower parts of the basin (Macklin and Dowsett, 1989). This enables some fine-grained alluvial stratigraphic units to be traced considerable distances through the Tyne drainage system, and allows flood histories in headwater and lowland river reaches to be compared.

The headwater tributaries of the South Tyne, most notably the Rivers Allen, Black Burn and Nent (Fig. 1), drain the northern part of the North Pennine Orefield, once the most productive metalliferous region in the UK (Dunham, 1944). A smaller area of mineralisation is also found between Haydon Bridge and Hexham (Fig. 1), close to the confluence of the North and South Tyne. There are no significant metal-bearing deposits in the North Tyne basin (Smith, 1923). The principal base metal minerals are galena (PbS) and sphalerite (ZnS) and both contain significant concentrations of trace metals including cadmium, copper and silver (Young et al., 1987).

In common with many other river systems in Britain draining metal mining areas (Lewin and Macklin, 1987), effluent from the extraction and processing of lead and zinc ores resulted in the release of large quantities of fine-grained, metal-rich waste into the River Tyne. Although sediment-borne metals caused serious contamination of riparian agricultural land downstream of mining areas, and gross pollution of many water courses, they do however provide a distinctive source of "labelled" sediment. Furthermore, as the development and history of metal mining in the Tyne basin is known in some detail from the end of the seventeenth century (Raistrick and Jennings, 1965), heavy metals are important stratigraphic markers, especially within sequences of vertically accreted fine-grained alluvium (Klimek and Zawilinska, 1985; Knox, 1987b; Lewin and Macklin, 1987 and Macklin, 1985) such as those exposed at Low Prudhoe.

FIELD SAMPLING AND LABORATORY ANALYSES OF FLOOD SEDIMENTS

Fine-grained alluvial sediments in the upper part of the river bank section at Low Prudhoe were sampled using Kubiena tins. In the laboratory the 2.4m long monolith was cleaned, drawn and sub-samples (135 in total) taken from every distinct layer for metal analyses and organic matter determinations. Additional material for grain size analysis was collected from the coarser, more prominent sand units (13 samples). Sediment colour was described using Munsell notation.

All samples were air dried, passed through a 2 mm sieve and heavy metal contents were estimated from the amounts of metals brought into solution by digesting sediment samples in nitric acid. Metal (lead, silver and zinc) concentrations were determined using atomic absorption spectrophotometry (Philips SP2900 with computer) involving direct aspiration of the aqueous solution into an air-acetylene flame. The organic content of sediments was taken as the loss of weight on ignition (L.O.I.) at 450 $^\circ$C. Particle size distributions were obtained using standard dry sieving methods with sieves graded at 0.25 $\emptyset$ intervals. Mean grain size, sorting, skewers and kurtosis were determined graphically following Folk and Ward (1957).

SEDIMENTARY PROPERTIES OF FLOOD DEPOSITS AND DATING FLOOD EVENTS AT LOW PRUDHOE

Sedimentary sequence and grain size characteristics of flood sediments

Flood events in fine-grained alluvial deposits at Low Prudhoe are represented by layers of generally flat beaded sands and silty sands (Fig. 3), which on the basis of grain size (following Niggli, 1935 and Pettijohn, 1975) and colour can be assigned to one of three sedimentary categories. Grain size curves for these flood sediment types are shown in Fig. 4.

1   Medium-fine sand (type 1 flood unit): yellow brown in colour (10 YR 5 4), mean grain size between 2.1 and 2.7 $\emptyset$, mean L.O.I. 2.4% and with generally less than 5% silt and clay.

2   Fine-very fine sand (type 2 flood unit): dark yellowish brown in colour (10 YR 3 4), mean grain size between 2.7 and 3.3 $\emptyset$, mean L.O.I. 3.3% and with between 5-10% silt and clay.

3   Silty fine- very fine sand (type 3 flood unit): dark brown in colour (7.5 YR 3 4), mean grain size 3.3 $\emptyset$, mean L.O.I. 3.7% and with more than 10% silt and clay.

Following Knox (1987a), textural discontinuities formed by layers of medium-fine (type 1 flood unit) and fine- very fine sand (type 2 flood unit) that reverse the overall "fining upward" sequence are interpreted as the deposits of large floods. Finer-grained silty sands (type 3 flood unit) are believed to represent lower magnitude floods or sediment deposited on the failing stage of a large flood event. Where coarser type 1 or 2 flood deposits overlie finer-grained type 3 flood units boundaries are generally sharp and sometimes marked by angular unconformities (eg 80 and 120 cm), probably produced by scour at the initial stages of a flood (cf. Jahns, 1947). In contrast, where finer flood sediment (type 3 flood unit) overlie type 1 or 2 flood deposits (eg 100 and 170 cm) bounding surfaces are more usually gradational, fining-upwards over 1-2 cm. In such cases finer sediment is believed to have been deposited on the falling limb of a flood hydrograph producing a flood "couplet" similar to flood units identified by Mansfield (1938), Farrell (1987), Klimek (1974) and Knox (1987a). On the other hand, sharp, non-gradational transitions from type 1 or 2 to type 3 flood units are the product of one or, more probably, several lower magnitude events. Higher percentages of organic material towards the top of many silty fine- very fine sand units suggest incipient soil development and slower accumulation rates. Complex grading is also evident within several of the sediment units (eg 222-205, 187-185, 135-128, 118-102, 60-48, 42-32 cm) believed to have been deposited by larger floods. This is likely to have resulted from changing flow conditions during a multi-peaked hydrograph (Taylor and Woodyer (1978), varying sources of sediment or periodic upstream bank collapse during a flood event (Osterkamp and Costa, 1987). Mudcracks (eg 118 cm), scour hollows (eg 183-180, 42-40 cm) and other pre-existing surface irregularities are also useful for differentiating between flood units and using these, together with other sedimentological criteria defined above, 25 large (identified by reversals in the normal fining-upward sequence) flood events would appear to be represented in the alluvial profile at Low Prudhoe. Vertical variations in the mean size and summary movement statistics of the thickest and best developed major flood units are plotted in Fig. 4. In general, flood units become finer, relatively better sorted and more platykurtic upsection. Moment statistics also pick out major textural discontinuities, corresponding with significant changes in flood unit thickness, at 185 cm, 120 cm and 35 cm. In the lower unit (230-185 cm) grain size decreases upsection with flood units becoming more poorly sorted and platykurtic. Between 185 and 120 cm this trend continues although flood units are generally more leptokurtic. In the upper sequence (120-35 cm) coarser, well sorted, leptokurtic flood sediments at the base are followed by progressively finer, less-well sorted and platykurtic flood deposits. Although graphical moment statistics confirm the basic facies architecture and stratigraphy of the Low Prudhoe section, they are less useful for identifying the upper and lower boundaries of flood units which can only be delineated by very careful logging of the profile. Establishing precisely the number, thickness and sedimentary characteristics of major flood units is of paramount importance for reconstructing the history of flooding at a site.

Trace metal dating of flood sediments

Vertical changes in sediment trace metal concentration at Low Prudhoe have been related to metal ore production in the Tyne basin and used to provide age estimates for flood events (Figs. 5 and 6). Water used in ore-processing, heavily laden with fine-grained metal-rich sediment, was usually discharged <u>directly</u> into the nearest stream. Thus although major lead and zinc mines in the Tyne catchment lay some distance from Low Prudhoe, the fine-grained nature of the mining waste enabled it to be transported rapidly downstream (particularly during major floods) thereby minimising the time interval (probably ranging from a few weeks to less than a year) between the release of metals from upstream mining operations and their incorporation in vertically accreted alluvium downstream within the lower Tyne Valley.

As has been shown in mining age alluvium elsewhere in the Tyne basin (Macklin, 1986, Macklin and Lewin, 1989; Macklin and Smith, 1990), especially high zinc levels and zinc lead ratios in sediments between 196-118 cm mark the peak of zinc extraction in the Allendale and Alston moor ore fields between 1897 and 1915 (Fig. 6). With a sharp reduction in zinc concentrations above 120 cm following the rapid decline of zinc mining in the region after 1915 (Dunham, 1944). Comparatively low lead levels (<1000 mg kg^{-1}) below 90 cm indicate sedimentation at Low Prudhoe post-dates the main phase of lead production in the Tyne basin that ended in 1880, confirming age estimates provided by sediment zinc concentrations. Above 90 cm lead concentrations rise in response to the revival of lead mining in the 1920's and 1930's. Peaks in lead production in 1927, 1933 and 1937 are reflected by increased lead concentrations in flood sediments and decreased zinc/lead ratios at 56, 28 and 24 cm. Most lead veins in the Tyne basin are argentiferous (Dunham, 1948) and, as one would expect, changes in silver concentration generally follow those of lead. To summarise, trace metal dating of fine-grained alluvial sediments at Low Prudhoe, somewhat unexpectedly, indicate that the major part of the sequence was deposited over ra comparatively short period of time (c.50 years) between 1890 and 1937.

FLOOD FREQUENCIES AT LOW PRUDHOE 1890 - 1990: A COMPARISON OF THE DOCUMENTARY AND SEDIMENTARY FLOOD RECORD IN THE LOWER TYNE VALLEY

On the basis of dating control provided by sediment heavy metal analyses, and in the knowledge that archive sources tend to record larger spatially extensive floods (Archer, 1987) flood units evident in the Low Prudhoe section were assigned as far as possible to documented floods since 1890 in the Tyne Valley (Table 1, Fig. 5). Not all flood units, however, could be related to recorded flood events and in these cases chemostratigraphic dating control provided general bounded time limits. In Fig. 7 floods documented in the Tyne basin between 1890 and 1989 are plotted with the sedimentary flood record (where present) over the same period and compared below on a decade by decade basis.

1890-1899

The sedimentary sequence for this period contains many more flood events than are documented in archive records (Fig. 7 and Table 1). This can be explained by a small height difference between the depositional surface and river bed at this time resulting in more frequent inundation by sub-bankfull flows. Although the accretionary surface at Low Prudhoe was probably flooded several times a year (as indicated by the large number of flood units dated to the 1890's), individual flood units are relatively thin (2-3 cm) and suggest deposition by relatively low magnitude events. Indeed, very few of these floods were sufficiently large to merit being

recorded by contemporary commentators. Thicker, coarser layers, however, probably correspond to higher magnitude floods (eg those of 1891 and 1892).

1900-1919

Sedimentary and documentary based flood records are more closely matched than in the previous decade. After 1900 (c.182 cm) the thickness of individual flood units increases significantly (>10 cm) as do zinc concentrations. Both reflect the input of large quantities of metal-rich fines into the South Tyne system during the peak period of zinc mining, 1898-1915. Lower zinc levels between 160-140 cm probably mark a temporary fall in zinc production between 1906 and 1910. Though this coincides with, and may well hav-- --en accentuated by, the North Tyne floods of 1907 and 1911 that would have introduced fine sediment with low metal concentrations into the River Tyne.

1920-1929

Six large floods are evident in the part of the Low Prudhoe profile dated to the 1920's (c.100-30 cm), one less than recorded in archive sources. Flood units are generally coarser (medium-fine sand) and thicker than those of earlier decades but low metal concentrations indicate that metal mining was not the source of this sediment. A marked reduction in the supply of mining-related fine material, following the closure of many mines at the end of World War I, appears to have been compensated by fine sediment generated from bank erosion (noted by commentators) during a series of very large floods in 1923, 1924 and 1926. Flood units with higher zinc concentrations and zinc/lead ratios (eg 60-50 cm) were probably derived from reworking of earlier metal contaminated alluvium located some distance upstream.

1930-1949

Although the documentary record for this period shows the frequency of large floods remained high (particularly during the early to mid 1930's) only two distinct flood units, dated to 1933 and 1947, can be recognised. Thus by the early 1930's the alluvial surface at Low Prudhoe appears to have accreted to a level above the channel thalweg that could only be over-topped by floods of a much higher magnitude and reduced frequency than before.

Post 1949

Since 1947 only two exceptional floods have inundated the Low Prudhoe section, those of 1955 (the largest flood in the Tyne since 1771) and August 1986 (the largest flood since 1955). River bed incision (up to 2m, Fig. 2b), which probably began sometime in the 1940's, transformed the alluvial bench at Low Prudhoe (a site of rapid vertical accretion between 1890 and 1937) into a low terrace. Many reaches of the Rivers South Tyne and Tyne, similar to the Tyne at Low Prudhoe, have incised since World War II (Macklin and Lewin, 1989; Macklin and Smith, 1990) in response to gravel extraction from channels (gravel was dug from the bed of the Tyne immediately upstream of Ovingham Bridge in the 1950's and 1960's) and bed erosion following the 1947 and 1955 floods. Widespread channel degradation also occurred during the August 1986 flood (Newson and Macklin, 1990), although a thin (2-3 mm) and patchy veneer of fine sand was deposited overbank at Low Prudhoe during this event. Four years on bioturbation processes have mixed this deposit with earlier sediment, to the extent that it is now very difficult to recognise as a distinct flood unit both in its original thickness and distribution.

In summary, the documentary flood record of the River Tyne between 1890 and 1949 shows an increase in flood frequency between 1890-1909 and 1920-1939 with

relatively few floods in the decades 1910-9 and 1940-9. Flooding over this period (1890-1949) follows changing hydrometeorological conditions in north east England with rainfall maxima recorded in the region during the late nineteenth century and again between 1920-1939 (Harris, 1985).

Most of the flood units deposited before 1900 appear to have resulted from sub-bankfull flows many of which will not have been reported by local commentators or newspapers. The similar number of floods between 1900 and 1929 evident in both the stratigraphic and documented flood record, however, suggests type 1 and 2 flood units were deposited during overbank events that inundated both the depositional bench at Low Prudhoe and the Tyne Valley flow. Since 1930 an increase in the relative height of this surface above the river bed (resulting initially from sediment accretion and later by channel incision) has effected a "censoring" of the alluvial stratigraphic record to reflect progressively less frequent and larger floods. Today continuing river bed incision has enlarged the Tyne channel at Low Prudhoe to a size that can accommodate floodwaters and sediment in all but the very largest floods.

We believe the disparity between the stratigraphic and documentary flood records at Low Prudhoe over the last 100 years reflects the latter's bias towards recording large overbank floods. Archive sources will therefore inevitably under-represent the actual number of flood events recorded in the stratigraphic record, especially when a depositional surface is of low elevation and subject to inundation by low magnitude floods. Deposition of fine-grained flood sediment within the Tyne channel (which in this particular study forms the basis of the paleoflood record) though reflecting in a systematic way changes in flood flow magnitude and frequency (controlled primarily by climate) has also been strongly influenced by variations in sediment supply (associated with upstream bank erosion rates and, in the case of the Tyne, input of mining waste) and entrenchment of the channel over the period of investigation. It is therefore imperative that sediment availability and the vertical tendency of a channel reach are evaluated, and quantified, before stratigraphic evidence at a site is used to extend the flood series in a river basin.

VERTICAL ACCRETION RATES AT LOW PRUDHOE 1890-1990

Rates of vertical accretion at Low Prudhoe from the end of the nineteenth century up to the present have been calculated by comparing the thickness of flood sediment accumulated between dated stratigraphic horizons (Fig. 7). There is a fairly close inverse relationship between organic matter in vertically accreted sediments and sedimentation ra s (Table 2), similar to that found by Knox (1987b) in Upper Mississippi floodplain sequences. Alluvial surfaces that are characterised by slow rates of sedimentation tend to accumulate organic matter in an incipient soil A horizon, whereas surfaces experiencing rapid alluviation normally accumulate relatively small amounts of organic matter (cf. Knox, 1987b).

Between 1890 and 1937 vertical accretion rates at Low Prudhoe were high averaging 4.8 cm a^{-1}. Since the 1930's, however, average yearly sedimentation rates have decreased by an order of magnitude to less than 0.4 cm a^{-1} at the present day (Fig. 7). Nevertheless, rates of recent fine-grained vertical accretion in the lower Tyne Valley are unusually high for British rivers (Table 3) and match rates recorded in some of the major river systems of Europe, North America and Australia reviewed by Bridge and Leeder (1979). The large sand load of the Tyne, derived from Carboniferous sandstone lithologies in its catchment and augmented by fine grained metal mining waste, appear to be the major causative factors. However, closer examination of the few river basins in Britain in which longer-term fine-grained sedimentation rates have been estimated show most are underlain by shale or clay lithologies where alluvium has been deposited overbank rather than within-channel as at Low Prudhoe. This, together with divergent catchment land-use histories, may

explain the generally low floodplain sedimentation rates recorded in the silt and clay suspended load rivers of lowland southern England (eg lower Severn, Avon, Ripple Brook. Culm) compared with much higher rates found to be more typical of river systems developed on sandstone lithologies in northern Britain (eg Swale and Tyne). Indeed, high vertical accretion rates would appear to be essential for the preservation of flood-related bedding structures and units in the alluvial stratigraphic record, whereas these features would tend to be destroyed by bioturbation and pedogenic processes at floodplain sites with lower rates of deposition.

CONCLUSIONS

Our studies of fine sediment deposition in the lower Tyne Valley represent the most detailed investigation of longer-term vertical accretion yet undertaken in a major British river. From a methodological viewpoint we have confirmed the utility of trace metals for dating fine-grained alluvial sequences in large mineralised catchments. While the identification of systematic differences between sedimentary and documentary based estimates of flood frequency in the lower Tyne highlights the important contribution fluvial geomorphologists can make to developing and refining historical flood information. Hitherto hydrologists have questioned lengthy extensions of flood data (much beyond a hundred years) because of possible climatic changes introducing non-stationarity into data series. The Tyne study reveals another source of non-stationarity; changes in river bed and sedimentation levels that may alter the flood-carrying capacity of a channel, and therefore the frequency of inundation and the flood hazard at a site. This is an important second-order effect when considering the non-stationarity introduced by climatic change and is of paramount importance in locating protection works. Significant recent channel bed level changes in the lower Tyne also to a large extent invalidates using floodstones to accurately calculate discharges and return periods of extreme floods. This constraint to historical flood reconstruction may well apply to many other northern British rivers. If, as predicted, the next 100 years is a period of profound climatic change it may well be that river engineers will need to know considerably more about the spatial and temporal patterns of channel capacity change as well as their controls.

Finally, our investigations highlight at least two lines of future research with respect to deposition of fine-grained alluvium and floods on British and European floodplains. Firstly, the recognition of significant _vertical_ accretion of _fine_ sediment _within_ river channels suggest that fine sediment members of historic and Holocene meandering rivers that are generally described as overbank deposits may need to be re-interpreted. The higher sand component in within-channel fines. together with regular textural reversals, may be useful criteria by which they can be distinguished from floodplain sediments deposited solely by overbank flood events. Secondly, with the development of luminescence sediment dating techniques (Smith et al., 1990), sedimentological-based estimates for flood frequency at a site using sequences of vertically accreted alluvium (as outlined in this paper) could be employed over longer Holocene timescales. O.S.L. dating of flood units in sand members of late and middle Holocene alluvial fills in the lower Tyne Valley is now in progress to assess the potential of this approach.

ACKNOWLEDGEMENTS

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YEAR	DAY & MONTH	YEAR	DAY & MONTH	YEAR	DAY & MONTH
1891	24 AUGUST	1930	26 JULY	1960	10 OCTOBER
1892	2 SEPTEMBER	1931	7 NOVEMBER	1962	3 APRIL
1892	3 NOVEMBER	1933	1 FEBRUARY	1962	27 AUGUST
1895	3 AUGUST	1934	17 MARCH	1962	11 SEPTEMBER
1900	6 OCTOBER	1934	14 APRIL	1963	21 NOVEMBER
1900	27 OCTOBER	1936	14 DECEMBER	1964	8 DECEMBER
(5) 1903	9 OCTOBER	1938	8 OCTOBER	1965	30 JULY
1904	24 NOVEMBER	1939	9 JANUARY	(6) 1967	17 OCTOBER
(10) 1906	30 MAY	1939	1 DECEMBER	1968	23 MARCH
1907	9 JUNE	1941	17 FEBRUARY	1968	13 SEPTEMBER
1911	14 MAY	(3) 1947	22 APRIL	1975	30 AUGUST
1912	- JUNE	1948	20 AUGUST	1976	21 JANUARY
1919	1 NOVEMBER	1953	12 NOVEMBER	1977	26 JANUARY
(9) 1923	13 NOVEMBER	1954	3 APRIL	1978	28 DECEMBER
(4) 1924	30 DECEMBER	(7) 1954	18 OCTOBER	1979	29 MARCH
(8) 1926	20 SEPTEMBER	1954	2 DECEMBER	1979	26 NOVEMBER
1927	23 JULY	(1) 1955	10 JANUARY	1981	3 MARCH
1927	21 SEPTEMBER	1956	17 FEBRUARY	(2) 1986	26 AUGUST
1929	10 AUGUST	1956	28 AUGUST	1990	4 FEBRUARY
1929	16 NOVEMBER	1958	12 FEBRUARY		

Table 1 Documented floods in the River Tyne 1891 - 1990 as recorded in local newspapers, books, journals and meteorological publications

The ten largest floods and their probable ranks are also indicated

DEPTH (cm)	DATE	SEDIMENTATION RATE (cm a-1)	MEAN LOI (%)
— 0	1990		
		0.3	4.2
—10	1950		
		0.8	5.5
—18	1940		
		1.2	5.3
— 30	1930		
		7.0	2.9
— 100	1920		
		5.0	3.1
—150	1910		
		3.0	3.3
—180	1900		
		5.0	3.1
—230	1890		

Table 2   Rates of fine-grained vertical accretion at Low Prudhoe 1890-1990

Table 3 Fine-grained vertical accretion rates in selected British river basins

AUTHOR	RIVER BASIN	CATCHMENT AREA (km²)	SEDIMENTATION RATE (cm a⁻¹)	TIME SCALE OF DEPOSITION (yr BP)
BROWN, 1987	SEVERN, UK	10000	0.14	0-10000
THIS STUDY	TYNE	2198	2.37	0 - 97
SHOTTON, 1978	AVON	1870	0.50	0 -3000
BRENNAN AND MACKLIN, UNPUBLISHED	SWALE	550	0.53	0 - 130
MACKLIN, 1985	AXE	31	0.54	0 - 312
BROWN AND BARBER, 1985	RIPPLE BROOK	19	0.05	0 -2500
BROWN, 1983	STOUR	620	10.20	MAXIMUM DEPTH OF OVERBANK DEPOSITION IN 1979 FLOOD
MACKLIN AND NEWSON, UNPUBLISHED	SWALE	550	13.00	MAXIMUM DEPTH OF OVERBANK DEPOSITION IN 1986 FLOOD
LAMBERT AND OVERBANK WALLING, 1987	CULM	276	0.05	AVERAGE SEDIMENTATION 1983-1984

catchment area header is CATCHMENT AREA (km²) and sedimentation rate header is SEDIMENTATION RATE (cm a⁻¹).

Lead Mines Showing
Output in Tons.
1700-1938

- 20 50.000
- 50-100.000
- >100.000

Zn Mines that Produced Zinc

☐ Study Site

N

TYNE
BASIN

10 km

NEWCASTLE

R. Tyne

R. Derwent

Consett

Derwent
Resr

Low Prudhoe

Devil's Water

HEXHAM

R. East Allen

R. West Allen

Zn

Zn Zn

Zn

Zn

R.
Alston Nent

Zn

Black Burn

R. South Tyne

Haydon Bridge

Haltwhistle

Bellingham

R. North Tyne

Kielder Resr

R. Rede

Catcleugh Resr

Otterburn

A

1860   1895   1926   1951   1978   N

OTTER BURN   WHITTLE BURN   OVINGHAM RECTORY

Prudhoe Station

RIVER TYNE

B   B'

Section

0   0.5   1 km

B

B   1771   B'
1815
LOW PRUDHOE   Present River Level   OVINGHAM RECTORY FLOOD STONES

A   1771   A'
1815
RAILWAY

SECTION

5
E
0
0   20
m

V.E. x 4

Channel and Gravel Bars

Tributaries

Cross Sections

Fords

Road Bridge

Railway

Cumulative % Coarser

Grain Size (Ø)

99.99 99.00 90.00 50.00 10.00 1.00 0.10

1.00 1.50 2.00 2.50 3.00 3.50 4.00 (4.00)

Medium–Fine Sand (110 cm)

Fine–V. Fine Sand (170 cm)

Silty Fine–Very Fine Sand (98 cm)

KURTOSIS(ø)   05 075 10 125 15

SKEWNESS(ø)   -025 0 025 05

SORTING(ø)   05 06 07 08 09

MEAN SIZE(ø)   20 25 30 35

MAJOR FLOOD UNITS WITH PROBABLE DATES
DATES BASED ON METAL CONCENTRATION AND RATIOS

1947
1937 1933
1933
1929
1927 1927
1926
1924
1923
1919
1915
1911 1910
1907
1903
1900 1900
1895
1892
1891
1890 1890

DEPTH (CM)

0 10 20 30 40 50 60 70 80 90 100 110 120 130 140 150 160 170 180 190 200 210 220 230 240 250

Gravel

Medium–fine sand

Fine–very fine sand

Silty fine–very fine sand

	Gravel		Fine–very fine sand
	Medium–fine sand		Silty fine–very fine sand

Documented floods in the Tyne Valley

Sedimentary flood unit at Low Prudhoe

Zn Concentrates

Pb Concentrates

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EFFECT OF ENGINEERING SOLUTIONS ON SEDIMENT TRANSPORT

Abstract

The scope of engineering works on rivers are flood capacity, navigability and water use. Channel stability is a necessary condition for these purposes. The concept of a static stable channel or a dynamic equilibrium between sediment supply and channel geometry is usually the base of river training methods. Very often an overcapacity of sediment transport was induced by regulation works, which was necessary for a certain time to create the desired channel shape. Long term evolution often resulted in substantual erosion, what called for additional drop structures. At lower slopes an apparent equilibrium is possible, what means that non-equilibrium conditions will induce geometrical changes only over an extremely long period. On steeper slopes, buffers (e.g. between sills) usually reduce the effects of long term changes.

1. Natural Gravel Bed Rivers

Natural alluvial rivers with gravel bed do form their channel in their own deposits or in ancient fluvial or glacial deposits. In the first case, they are usually aggrading (see fig. 1). Erosion is more common in the upper reaches of such rivers, allthough often limited by rock outcrops. Fig. 2 shows an example of a degrading reach, which lies upstream of a gorge which is incised in a rock barreer, which had orginally formed a lake. This incision is followed by a degradation process in the lake deposits.

The width of these rivers normally adjusts to a value which corresponds nearly to conditions of beginning of motion for a low frequency flood discharge (Parker, 1976).

Natural gravel bed rivers are usually braided, as shwon in Fig. 1 and 2. Smaller rivers with coarser bed material tend to meander. Straight reaches are found in boulder bed channels.

Therefore, natural gravel bed rivers are not stable, but shift their channels as a consequence of their planform, and are aggrading or degrading. Except for steep mountain gullies, these processes can be expected to be slow because of the large channel width and the associated low bed shear stresses.

Fig. 1 Rakaia River (New Zealand) near its mouth, in an aggrading reach

Fig. 2 Rakaia River (New Zealand) in its middle reach, where it is incising in old lake deposits

2. Static and Dynamic Equilibriums - Scope of river Engineering Work

Land reclamation, flood protection, navigation, and use of water power all need stable channels. This stability must be warranted by a dynamic equilibrium between sediment transport capacity and sediment supply (Zanke, 1972). Meyer-Peter and Lichtenhahn (1963) claim the search for such an equilibrium to be the main objective of river engineering works.

If their is no or negligible sediment supply to a river reach, equilibrium there can be only a static one. The discharge should always remain smaller than the one responsible for beginning of motion. In gravel bed rivers, critical conditions of an armour layer become dominant (Guenter, 1971).

In general, river training works tend to produce narrow and straight channels in order to increase transport capacity. In eroding channels the bed resistance is increased or drop structures are inserted to reduce the channel slope and thus transport capacity (e.g. see Ashida 1987).

It is interesting to note that there is almost no discusssion about equilibrium in sand bed channels. Janssen et al. (1979) consider fixation of a channel against the natural migration associated with planforms as the only stabilisation scope. It seems that owing to the very low slopes a major difference between transport capacity and sediment supply does not necessarily create major problems. In trained gravel bed channels with slopes of the order of 0.1 to 1.0 %, such differences can lead to rapid fill or scour as a result of change in slope, what in turn leads to a loss of flood capacity or collapse of bank protections. Channel narrowing reduces substantually the sediment storage capacity; so that aggradation and degradation processes will be substantually faster in the trained then in the natural channel.

3. Effect of a Channel Geometry Change on Sediment Transport Capacity

3.1 Theoretical background

All the particles which are relevant for the formation and alteration of a gravel bed channel move as bedload. Even the finer bedload particles may not be involved in this process. Considering bed load transport capacity of a channel, care has to be given to the representative grain sizes.
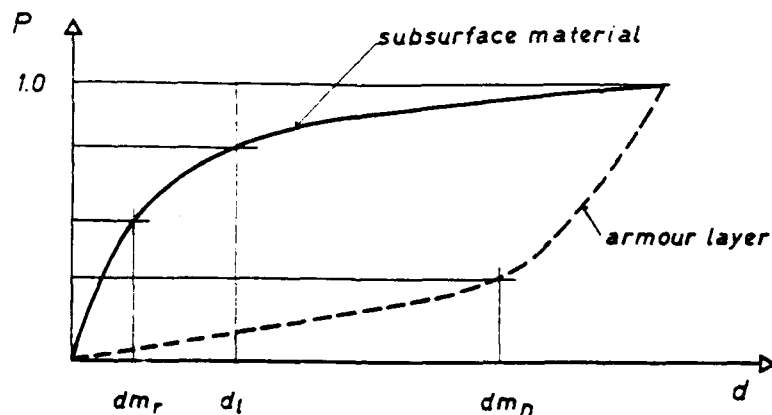
Fig. 3 Grain size distribution of gravel bed, and of the armour layer. Mean grain size d_{mr} of moving bedload and mean grain size of armour layer dm_D.

Fig. 3 shows a typical grain size distribution of a gravel bed and of possible armour layer. Nummerical simulation has shown that a reduced mean grain size of the subsurface material and the mean grain size of the armour layer can be used as the characteristic grain sizes (Hunziker and Jaeggi, 1988).

The reduced mean grain size is supposed to represent the moving bedload material and is defined by

$$d_{mr} = \frac{\sum\limits_{n} \Delta p_i \cdot d_i}{\sum\limits_{n} \Delta p_i} \tag{1}$$

where d_i is the mean grain size of a fraction. Δp_i the proportion of this fraction in the mixture measured by weight, and n the number of fractions finer than a limiting diameter d_L. d_L is normally characterized by a sudden change of the gradient in the grain size distribution curve.

The mean grain size of the armour layer corresponds to the usual definition

$$d_{mD} = \sum \Delta p_{Di} \cdot d_i \tag{2}$$

which this time extends over all the fractions. Δp_{Di} represents the proportion of a fraction in the armour layer and can be calculated by Gessler's (1965) procedure of others. d_{mD} is commonly approximated by d_{90}.

Sediment transport theory allows to define a sediment transport capacity for given channel cross section, slope, grain-size and discharge. A generalized version of a transport formula for gravel bed rivers, applicable over a wide range of slopes (Smart and Jaeggi, 1984), is

$$Q_{TC} = \frac{\alpha}{(\rho_s/\rho_w - 1)} \beta Q J^{1.6} (1 - \frac{1}{\eta}) \tag{3}$$

where Q_{TC} [m³/s] is the bedload transport capacity over the full width, ρ_s [kg/m³] and ρ_w [kg/m³] the densities of sediment and water, Q the water discharge, and J the slope. α is a coefficient determined experimentally to be $4 \le \alpha \le 4.8$, it may be considered as a variable in a somewhat wider range to account for the problem of fitting a single equation to numerous transport experiments. β is an efficiency factor. $\beta = 1$ corresponds to ideal condition in a laboratory flume with negligible wall drag. In straight single channels with plane bed wall drag reduces β to 0.7÷0.9. In strongly meandering or braided channels β may decrease to be 0.1÷0.3, but will definitely not reduce to zero. Finally

$$\eta = \frac{\Theta}{\Theta_{cr}} = \frac{h_m J}{(\rho_s/\rho_w - 1) d_{mr} \Theta_{cr}} \tag{4}$$

with $\Theta_{cr} \cong 0.05$ and h_m being a representative flow depth in the bedload carrying parts of the channel.

Armouring is considered by a value of

$$\eta_D = \left[\frac{d_{mD}}{d_m} \right]^{0.67} \tag{5}$$

For $\eta \leq 1$ there is no motion, for $1 \leq \eta \leq \eta_D$ material of size d supplied from upstream can be moved downstream, and for $\eta \geq \eta_D$ bed erosion is possible and thus supply of sediment from the bed in the absence of upstream supply.

If $Q(t)$ is a duration curve of a river, then

$$W = \int_0^T Q\ dt \qquad (6)$$

is the total water yield over a period T, usually taken as one year. Using eq. (3) over the whole discharge range, then one gets bedload yield over the same period T:

$$GF = \beta \cdot J^{1.6}\ \frac{\alpha}{(\rho_s/\rho_w - 1)}\ \int_0^T Q\ (1 - \frac{1}{\eta})\,dT = A \int_0^T Q\ (1 - \frac{1}{\eta})\ dT \qquad (7)$$

Fig. 4 shows that the integral part corresponds to an "effective part" of the water yield over the period T for sediment transport. The reduction is due to the beginning of motion conditions expressed by $1/\eta$. Potential erosion yield only is given for the period T_D for which η_D is exceeded. This is charaterized by F_1 and thus the corresponding minimum transport capacity by $A \cdot F_1$. Considering sufficient supply from upstream, the period T_0 with $\eta \geq 1$ must be considered. The corresponding area is the $F_1 + F_2$ and the maximum transport capacity $A\ (F_1 + F_2)$.
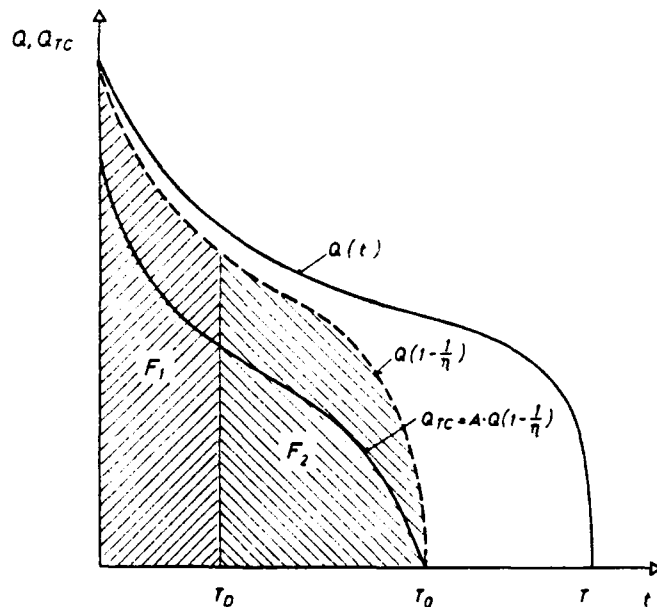
Fig. 4 Definition of maximum $(A \cdot F_1 + A \cdot F_2)$ and minimum transport capacity $(A \cdot F_1)$ over a period T. T_0 is the part of the period for which $\eta \geq 1$, T_D with $\eta \geq \eta_D$, and T_{2D} with $\eta \geq 2\ \eta_D$.

3.2 Discussion

A change in channel geometry (cross-section or alignement) results in a change of transport capacity (yields) as defined by eq. (7). Such a change may vary the slope, the efficiency factor β, or the area F by ΔF as a function of a variation of η. The change of one parameter may be dominant, but all of them may vary.

For two channel geometries, and in particular a material and an artificial one, transport capacities expressed in termes of annual yields can be defined and compared. According to the particular situation, maximum or minimum values on one of each my be considered. The ratio of transport capacities leads to an expression

$$\frac{GF_i}{GF_j} = \frac{\beta_i \ J_i^{1.6}}{\beta_j \ J_j^{1.6}} \ \ \frac{F_i}{F_j + \Delta F} \tag{8}$$

F_i and F_j may represent F_1 or $F_1 + F_2$ according to the particular situation. ΔF is dependent on the variation of the period with bedload transport (increase or decrease of T_0) and the change in the function $Q(\eta)$. Equation (8) allows to discuss, at least approximatively, the change in "efficiency" (parameter β) compared to a change in slope and the shift due to alteration in the condition for beginning of motion.

Another advantage of the discussion of transport capacities based on β, J and F or ΔF is that this expression (8) is not dependent on the fit of a transport formula to experimental results. This is described by the parameter α not appearing in this discussion.

A variation of the morphological conditions changes mainly β. Channel narrowing associated with a change from a braided river to a plane bed channel results in an increase for β from about 0.2 to 0.9. The change in η may be small compared to this if the slope does not change and η for the braided river was defind in the main bed load carrying channels. This shows the order of magnitude of a change produced by an engineering solution in the sediment regime of a river.

A disequilibrium probably exists along a river a channel. Expression (8) allows to find out within which limits transport capacity can be varied with variation of channel geometry to obtain a continuous sediment transport, what is the objective of engineering solutions. It may well be that along a channel a desired continuity of identical transport capacities cannot be realised, e.g. for a continuously decreasing slope. This limits the application of a dynamic equilibrum concept.

Many other kinds of comparitive calculations based on equation (8) can be imagined, which can visualize the effect of engineering solutions. In particular, changes in geometry resulting from other needs than sediment regime can be evaluated.

4. Training Works and Sediment Transport

4.1 Generalities

Straightening of a river course and narrowing of the cross-section is normally the consequence of major river training schemes. An increase in slope and a width reduction normally result in an increase of transport capacity. Fig. 5 show the relation between annual bedload yield (function of the discharge duration curve) and channel width for different slopes.
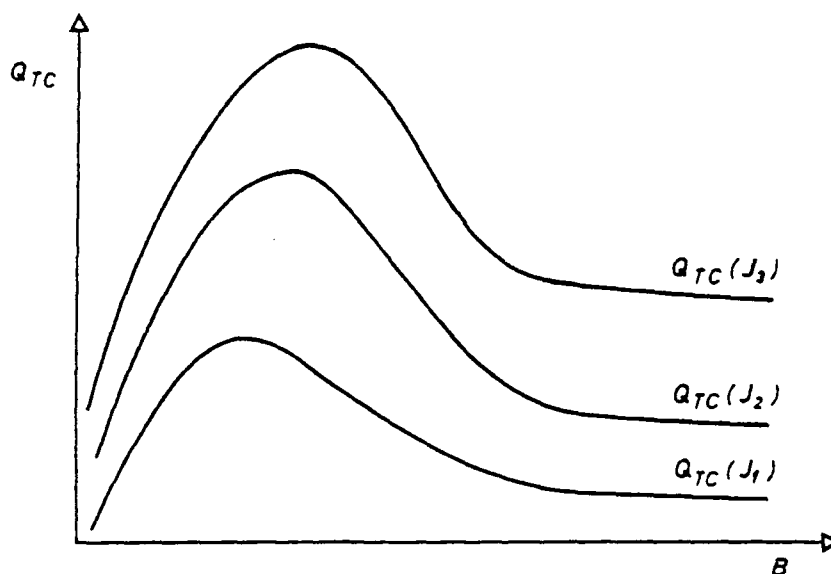
Fig. 5 Bedload yield as a function of channel width for different slopes ($J_3 \geq J_2 \geq J_1$) for given discharge duration curve

Obviously, an increase in slope means the shift from one curve to the upper one and thus an increase in transport capacity. In a certain range of channel widths, narrowing also produces an increase of sediment transport capacity. For very narrow channels, the opposite could happen, because of wall drag. For very wide channels, due to the braided pattern, transport capacity is almost independent of channel width. Finally, for given slope and duration curve, there is always a maximum of transport capacity (annual yield) which cannot be exceeded by any change of cross-sectional geometry.

Trained rivers have very often a channel width which corresponds to this maximum of transport capacity. Further narrowing of the section does therefore not induce a further increase in transport capacity. A reduction from natural width (braided channel) to an "optimum" width (corresponding to maximum transport capacity) induces a substantual increase in transport capacity. This is equivalent to an increase of the efficiency paramter β, discussed above, from about 0.2 to a value near 1.0. A width reduction is almost irrelevent on transport capacity if it for both widths the channel is braided. This latter case was characteristic for some of the early training works where the natural aggradation rate was not altered, but the deposition area reduced and thus the bed elevation rate increased (Jaeggi, 1989).

4.2 Overcapacities induced by River Regulation

An accelerated aggradation was the major feer of river engineers during the 19th century. If the river started to degrade after a regulation scheme was applied, this was considered to be a success. For this to happen, overcapacities had to be accepted. At that time, it was not possible, or not economic, to excavate the full desired channel cross-section. This work was left over to the river erosion. Since normally one objective of river training work was to lower flood levels, if possible even below ground level, this resulted in a potential erosion volume which corresponds to many annual bedload yields.

Normally, regulation works were executed upstream and extended over years and decades. As fig. 6 shows, the reach where a training scheme was finished delivered an extra load which saturated the overcapacity in the downstream reach. Only because of this overcapacity the downstream reach was able to evacuate natural supply plus the extra load. It is only a

certain time after this extra delivery had stopped, after the training works even in the tributerries were finished, that an erosion exceeding the originally accepted amount was the consequence. The river was still completing the natural supply by an extra load from the river bed. Many regulated Alpine rivers show such an erosive behaviour.
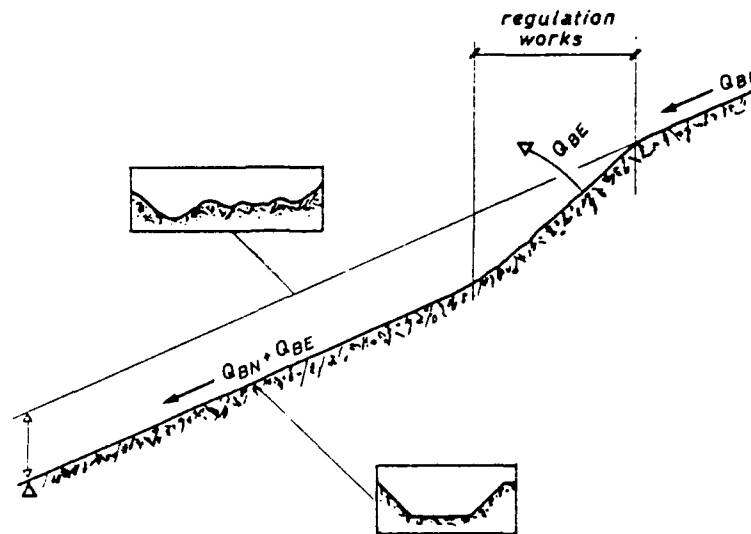

Fig. 6 Lowering of river bed, the aim of many river regulations, producing an extra load of sediment. The reach downstream of the construction site is maintained in equilibrium by the natural supply plus the extra supply

4.3 The River Danube Downstream of Vienna

From a multiple channel river the Danube near Vienna has been transformed in a more or less single thread channel for navigation purposes in the 19th century. Between this main channel and the remote flood protection dykes a natural habitat had been preserved or even redevelopped, which had become internationnaly famous. Upstream of Vienna, after World War II a series of hydro-power schemes again designed for improvement of the navigation have been developped. Downstream of Bratislava (CSFR) the scheme of Gabcikovo is now almost completed. Some years ago the building of the Hainburg power plant was supposed to fill the gap. National and international protests finally prevented that these scheme was realised and the river forest in the remnant flood plain was protected.

Since the last centuries regulation, but now with increased speed because of the hydro-power development in the upstream reach and gravel dredging for improvement of the navigation, the Danube erodes its bed. The project of building a power plant in Vienna would even increase the problem in the downstream reach. This was investigated recently with the aid of a numerical simulation (VAW, 1989).

It came out that the rate of erosion is reasonably small, despite the general trend of incision. The local erosion in the upper reaches (what means downstream of the last power plant) is sufficient to nearly saturate the deficit in transport capacity for a long time. Again, a comparatively steep downstream reach can thus be held in equilibrium over decades until the supply to this reach decreases. The incision rate is slow because the annual minimum transport capacities (yields), which are relevant for erosion, are small compared to the volumes involved. As fig. 7 shows, it would need over 80 years to erode the bed between 0.3 and 1.1 m.
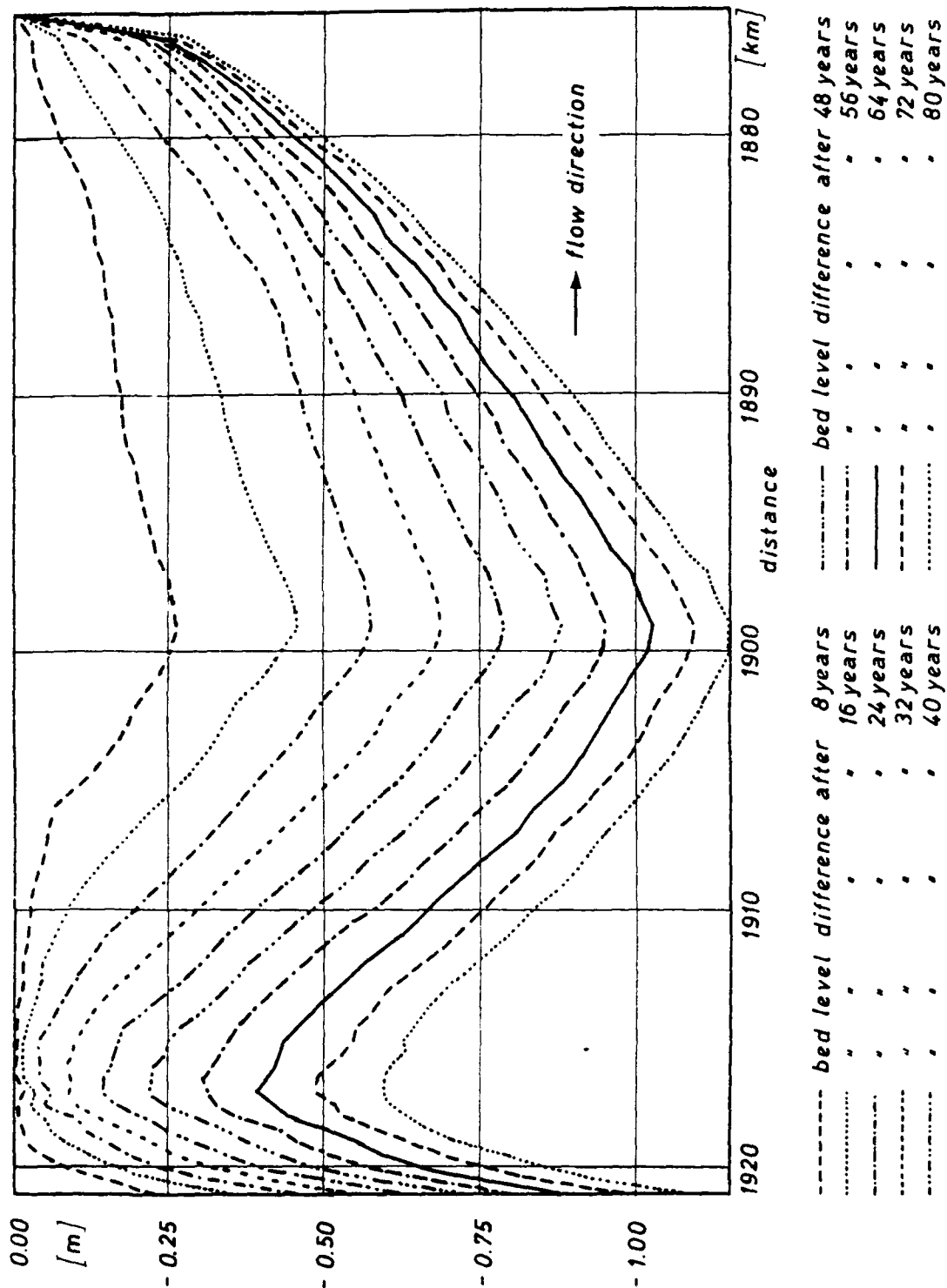
Fig. 7   Bed   level   differences   calculated in the Danube between
         Vienna (km 1921)   and   the   Austro-Czecoslovakian   border
         (km 1875) in a 80 year simulation

4.4 The Emme River

The Emme river is a tributary of the Aare river in Switzerland. Over about 60 km it has been trained in the second half of the last century and the first thirty years of this century. The valley slope is 0.8 % in the upper, 0.6 % in the middle, and 0.4 % in the lower reach.

An evolution as described in section 4.2 occured in this river. After completion of the lower parts of the scheme, the desired erosion was observed. Often, it was considered to be too slow and dredging plants were installed to speed up the process. However, after the extra supply from the regulation works progressing upstream was diminishing, bed erosion as shown in fig. 6, associated with a lowering of bed slope started. Drop structures were built to stop this process and to protect local structures like bridges and water intakes. After about 80 years about one third of the total head in the regulated river reach corresponds to the sum of the drops of these structures. Investigation of the bedload regime (Hunziker and Jaeggi, 1988; Jaeggi, 1989) has shown that the bed erosion had not stopped since the regulation scheme was applied and was still nowadays a consequence of the original overcapacity. Assuming that the actual training works would not be destroyed by further erosion, it would need another 50 years to reach some sort of equilibrium conditions. The actual situation and such a future evolution is shown in fig. 8. The actual sediment regime is therefore far away from an equilibrium one, with a two and a half fold increase of the yield along the channel.
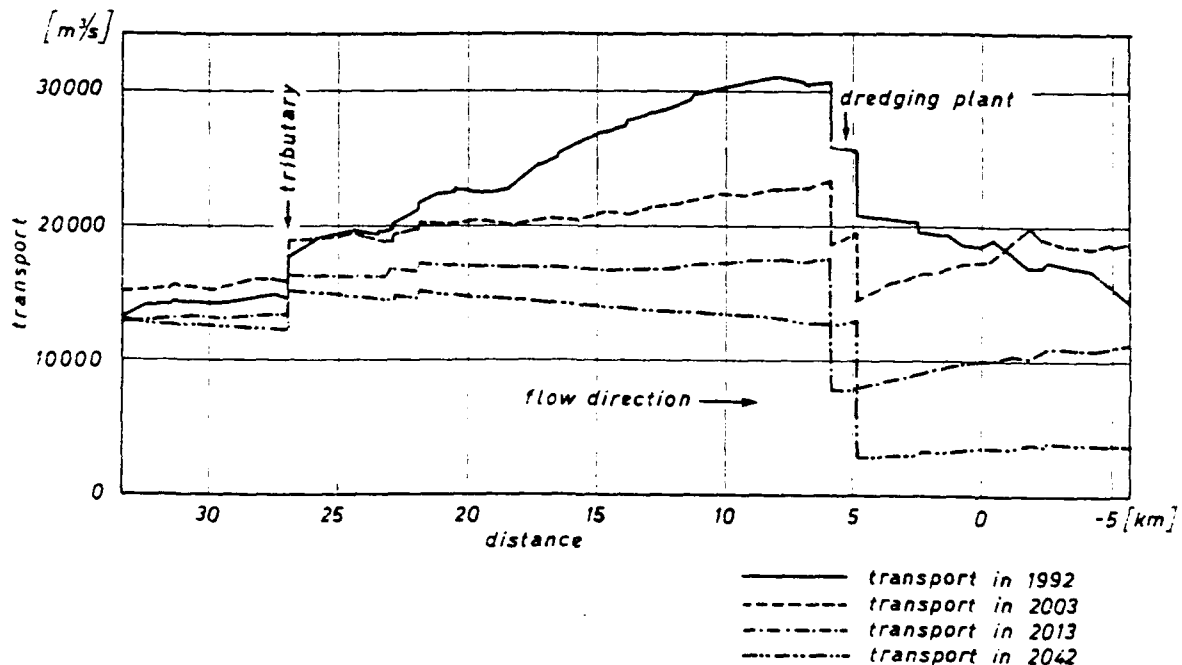
Fig. 8 Actual transport situation along the Emme river and future evolution (for idealised conditions)

5. Discussion of the Long Term Equilibrium

Analysis of transport capacity along a river reach, would very probably lead to a diagnosis of disequilibrium. The question is how to cope with this situation.

5.1 The Apparent Equilibrium

If transport capacities are small, for small slopes and/or coarse bed material, then a non-equilibrium situation does not mean that an immediate adjustment is induced. Channel geometry may be apparantely stable, although transport capacities and sediment supply do not correspond. The Danube downstream of Vienna reacts very slowly to the reduction to zero of tne sediment supply. A final equilibrium stage would be reached only after centuries. From one year to the other, theoretical bed level changes are in the order of millimeters or centimeters. This may be called the apparent equilibrium.

5.2 The Quasi-Equilibrium

On a differential scale, the equilibrium conditions are fulfilled. A difference between local transport capacity and sediment supply to a limited reach is compensated by local erosion or aggradation. On a larger scale, this local compensation may maintain a longer reach in equilibrium over a certain time period. One example is the regulated reach being fed by the extra load of erosion out of a freshly trained reach. As for erosion, in case of too low capacities a local buffer may take over the aggradation rate and supply a smaller rate to a downstream reach wich is then in equilibrium with this small rate.

Although it is obvious that a long term scale the situation will change, such a quasi-equilibrium may well be stable over years or decades.

5.3 Concept of Buffers

5.3.1 Aggrading Reaches
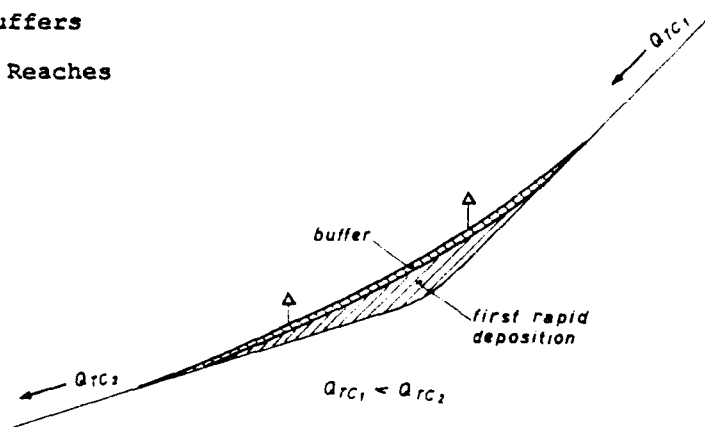
Fig. 9 Concept of a buffer in a aggrading reach

Naturally, the slope of alluvial rivers is decreasing downstream, and therefore normally also transport capacity. In case of an apparent equilibrium, aggradation is very slow because it spreads over a long distance. The potential aggradation volume (see fig. 9) can be viewed as a buffer. Assuming that the aggradation rate is constant, this means that there must be a constant decrease in transport capacity. In case of change in slope, as in fig. 10, if can be expected that a first quick aggradation fills up a first volume and allows that a linear buffer to take over the

loads of following years. In a situation like shown for the Danube, a buffer defined by a considerable length may take over many annual loads with a small change in bed level. A regulation scheme which leads to such an aggradation, or cannot prevent it, is still economic, if the effect of aggradation e.g. on flood capacity is feelt only after a period of fifty or hundred years. After that period, a new scheme may be applied.

Fig. 9 applies to natural rivers as well. It may be noted again that regulation schemes which had failed had, in such situations, just reduced the channel width and so the width of the buffer. Thus, a natural slow aggradation rate was changed to a dramatic one.

5.3.2 Degrading Reaches
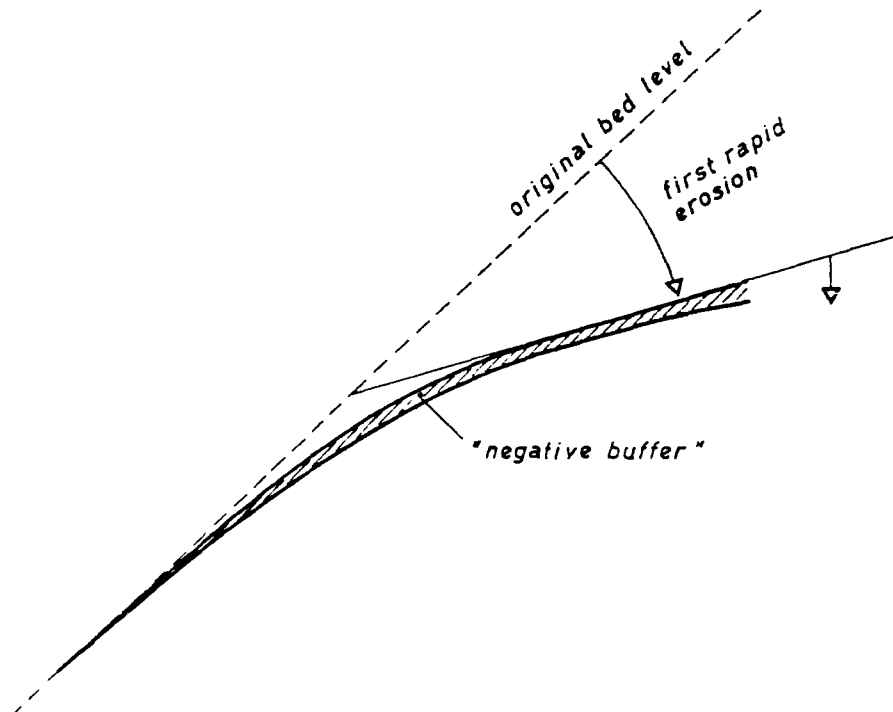
Fig. 10 Buffer in degrading reaches

A change in the transport capacity or the sediment supply may produce an erosion trend as shown in fig. 10. After a first rapid erosion again a transition reach developes whose contribution to the supply to the downstream reach may become substantial. Such an evolution was shown to happen in the Danube river after reduction of sediment supply. Again, this sort of negative buffer may have a sufficient influence so that a change in channel geometry becomes necessary only in a remote future.

5.3.3 Stepped Channels (artificial and natural)

Most of the steeper regulated rivers do form a sequence of artificial
drop structures, as described for the Emme river. Steep natural river from
step-pool systems (Whittaker, 1987) which have again a buffering effect on
sediment transport.
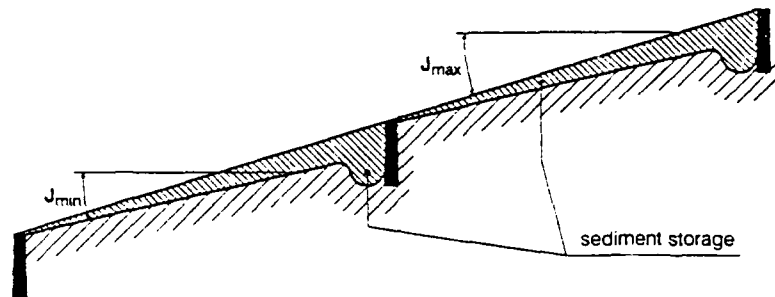
Fig. 11 Buffers in stepped channels

In both cases, the crest of the steps defines maximum transport
capacity, which is function of the valley slope. This is never or rarely
reached, because of natural or artifical overcapacity. A flatter slope
between the steps defines a minimum transport capacity, corresponding to
flow conditions for which the system is still stable with respect of
scouring the steps. If the natural supply to the system is within the
range defined by this two slopes, then the system remains stable. It is
therefore insensitive to variations in sediment transport, which would
just fill and empty the buffers. It is not surprising that this system of
including overcapacity and erosion control is quite popular.

It should be noted that Fig. 11 can strictly be applied only for
spacings of the steps of several hundreds of meters and more. For smaller
spacings, the concept applies to the energy line but not the bed level.
This takes then more a scour form. The scour is still a buffer, but its
shape depends not only on the energy line and the filling of the buffer,
but on local flow conditions also.

6. Dredging

Dredging is a solution often applied to cope with discontinuities of
transport capacity along a channel reach. Since a maximum of transport
capacity exists irrespective of any change in geometry it may often be the
only solution in case of large discrepancies.

Dredging is delicate if a certain transport rate downstream of the
intervention point is required. Very often, an erosion problem results. In
the presence of large buffering areas, the river's reaction to excessive
or insufficient dredging qunatities is slow. Problems with dredging in the
Alpine Rhine (Meyer-Peter/Lichtenhahn, 1963, Lichtenhahn, 1972 may be
viewed as an example of sucn a situation. There, after a last change in
the regulation scheme, accomplished in 1954, a dredging company had to be
paid for the excavation work. In the sixties, dredging became comercially
interesting and the dredged quantities excessive. The subsequent erosion
led to the collapse of the road bridge Buchs-Schaan in 1970.

7. Conclusions

River engineering solutions often change dramatically the sediment regime of a river. Often, such a change was the scope of the scheme. A dynamic equilibrium between sediment supply to a reach and transport capacity along the reach was considered to be the optimum solution. Effectively, buffers in long and wide river reaches or defined by artificial steps play an important role in sediment balance. Long term changes which are function of the behaviour in this buffers may lead to new conditions, in apparantely stable river reaches.

Computing Bed-load Discharge and Channel Adjustment in a Cobble Bed River for Flood Control Channel Design

by

Brad R. Hall[1], William A. Thomas[1], and Monte L. Pearson[2]

The Hydraulics Laboratory, US Army Engineer Waterways Experiment Station, has computed the existing condition sedimentation regime of the Truckee River at Reno, NV, in support of channel design modifications for flood control. The sedimentation analysis was complicated by the cobble-sized bed materials found in the channel, several existing bridges in the study area that significantly affect the channel hydraulics at flood discharge, and water diversion drop structures that provide local storage zones for bed-load transport.

Bed-load measurements on the Truckee River at the upstream end of the study reach were not available. The inflowing sediment load was determined by calculating the equilibrium sediment transport rate for a range of water discharges for a reach of the Truckee River within the study area called the equilibrium reach. The equilibrium reach is geomorphically stable and exhibits characteristics in channel planform, bed material, and channel banks that are similar to that portion of the Truckee River immediately upstream of the study area.

The one-dimensional sediment transport model "Sedimentation in Stream Networks," TABS-1, was used to quantify sedimentation processes. Sedimentation model adjustment was completed by comparing observed rating curve changes due to channel aggradation with computed shifts in the rating curve. Average annual and design flood sedimentation were quantified with the TABS-1 model.

INTRODUCTION

Study Area

The Truckee River study reach is located near Reno, NV, and extends from the Vista gage at approximately River Mile (RM) 43.9 to just upstream of the Booth Street bridge at RM 53.0. A map of the study area is shown in Figure 1. The Truckee River is a perennial stream characterized by pool and riffle channel morphology. Several bridge crossings and water diversion structures are found in the study reach. Man-made channel modifications, especially within the upper 3 miles of the study reach, have limited the amount of channel migration. Bed material size decreases through the reach, and the channel bed is armored at base flow discharge.

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Approach

The Hydraulics Laboratory, US Army Engineer Waterways Experiment Station, computer program "Sedimentation in Stream Networks," TABS-1, was used to quantify channel response to hydrologic variations. The Hydraulics Laboratory developed TABS-1 as a research tool to incorporate new and enhanced sediment transport technology in the widely used HEC-6 moveable-bed model (US Army Engineer Hydrologic Engineering Center, 1977). A detailed description of the assumptions, limitations, and data requirements for the TABS-1 model is provided in Appendix A of Copeland and Thomas, 1989. The study reach of the Truckee River lacks sampled bed-load transport rates, suspended sediment concentrations, and morphological surveys documenting lateral and vertical river channel migration to permit the normal numerical model adjustment and verification. This study combined a theoretical treatment of sediment transport with observed channel morphology to quantify channel bed dynamics associated with gravel and boulder streams. In addition to quantifying instantaneous sediment transport rates, the analysis allowed evaluation of channel bed aggradation and degradation response to long-term hydrologic sequences.

FIELD DATA

Hydrological and Geometric Data

There are three United States Geological Survey (USGS) stream gages in the Truckee River study reach (Figure 1). The gages and the period of record for streamflow data at each gage are listed in the following tabulation.

USGS Stream gages in the study reach.

Gage	Drainage Area, square miles	Period of Record
Truckee River at Reno, NV USGS Gage No. 10348000	1067	1907 - present
Truckee River near Sparks, NV USGS Gage No. 10348200	1070	1977 - present
Truckee River at Vista, NV USGS Gage No. 10350000	1431	1900 - present

Channel geometry was developed from channel and overbank surveys completed in 1975. Bridge geometry was surveyed in 1989 including low and high chord profiles and bridge pier location and orientation.

Sampled Sediment Transport

An extensive review of several technical libraries indicates that no measurements of bed-load transport have been documented for the project reach of the Truckee River. Sampling of suspended sediment concentration by the USGS in the Truckee River in the Reno, NV, area has occurred on a sporadic basis. Provisional records from the USGS provide the concentrations listed in Table 1. Suspended sediment concentration and river discharge have been sampled at the McCarran Boulevard bridge (RM 47.7) as part of separate water

quality and water supply studies (Table 2).

Bed Material Samples and Gradation

Twenty sediment samples were collected from the channel bed, bars, and bank at 15 locations spaced along the 9-mile study reach in June 1989. Samples were taken near the water's edge, from the channel bed, and from locations on point- and mid-channel bars. Bed material sample locations are plotted in Figure 2 along with a channel thalweg survey and computed water surface profile for a flood discharge of 25,000 cfs. Wolman count (Wolman, 1954) bed surface material gradations were collected (Water Engineering and Technology, Inc., 1989) providing an estimate of the surficial coverage of the channel bed by coarse grain sediments (i.e., cobbles) which are generally omitted in bulk material samples. Profiles of the percent finer for each sediment size class sampled in the grab samples and the Wolman counts used for the numerical model application are given in Figures 3 and 4, respectively. Grain size classifications follow the American Geophysical Union standard recommended by Lane, 1947.

Channel Characteristics

Field observations in 1989 of the project area indicate that the Truckee River can be generally considered as three separate and distinctive reaches in terms of channel and bed characteristics. These reaches are identified as (a) the reach from the upstream limit of the study area to the Highway 395 bridge, (b) from Highway 395 bridge to just downstream of the McCarran Avenue bridge, and (c) the reach from the McCarran Avenue bridge to the downstream limit of the study area at the Vista gage. Each reach is characterized by distinctive bed material sizes and channel characteristics.

The reach upstream of Highway 395 bridge is an armored pool and riffle channel. The river channel is incised into the coarse sediments of the Tahoe and Donner Lake outwash deposits. The large boulders of the Tahoe and Donner Lake formations are apparently armoring the bed causing an irregular series of riffles and pools whose distribution is controlled by the intersection of the Truckee River with boulder beds in the underlying deposits. The channel is relatively straight with local high energy gradients in the boulder riffles. Within this reach, bridge abutments and piers have caused local scour and deposition in the "downtown" subreach (RM 51.7 to RM 52.2). There is a noticeable absence of boulders in the river channel in the downtown subreach. Excavations adjacent to the channel indicate that boulders are present in the alluvial materials adjacent to the subreach, indicating that either the boulders were removed from the river channel since the existing channel improvements were constructed, or gravel and cobble sediments have deposited and covered the boulders in the lower energy gradient region upstream of the bridges. With the exception of the downtown subreach, the principal geomorphic activity of the Truckee River in this reach appears to be gradual bed degradation and winnowing of the finer (sand and gravel) materials encountered in the outwash deposits. As the Truckee continues to degrade slowly in this reach, the bed will maintain an armor of cobbles and boulders derived from the outwash deposits.

From the Highway 395 bridge to McCarran Avenue bridge, the Truckee River exhibits a slightly meandering planform with a marked decrease in the size of

the bed material. Several types of bars occur including point-, alternate-, and mid-channel bars. This reach is apparently a sediment storage reach where the coarse bed load that passes through the upstream reach is gradually deposited in several river bars. The bars are armored by gravels and cobbles. The distribution of the alternate and point bars suggest that the coarser point bars are formed and maintained by higher discharges and the alternate bars by more frequent lower stage flows (Water Engineering and Technology, Inc. 1989). This reach is generally experiencing slow lateral migration through point bar growth and bank migration.

Downstream of McCarran Avenue Bridge, the Truckee River changes to a gently meandering channel which is incising into fine-grained cohesive alluvium. This alluvium consists of slack-water deposits of silty clays that were most likely deposited during an earlier blockage of the Truckee River at Vista. The bed consists of irregularly distributed scour pools incised into the cohesive deposits. Gravel material appears infrequently on the bed, which generally consists of sand or a clean erosional flute in the underlying silty clays. The number of bars decreases considerably in this lower reach as the channel generally decreases in sinuosity. Steep channel banks may be unstable locally as indicated by several large rotational slides.

SEDIMENTATION ANALYSIS

Stability of the Truckee River
at Reno Stage-Discharge Curve

The stage-discharge rating curve developed by the USGS for the Truckee River at Reno, NV, gage (No. 10348000) was reviewed for any significant changes in the rating. The gage is located near the Highway 395 bridge at approximately RM 50.6. The measurements indicate a very stable stage-discharge relationship through 1977. After 1977, a slight increase in stage over time for low discharges can be detected. The rating curves used through 30 September 1977, 3 February 1982, and 17 February 1986 are compared in Figure 5. The rating curves indicate that for discharges less than 1,000 cfs, the stage has increased approximately 0.5 ft at the gage. The increase in stage for a given discharge rapidly diminishes for discharges greater than 1,000 cfs. Possible reasons for the change in the stage-discharge relationship are (a) a gradual channel aggradation at the gage location, and/or (b) changes in the relative roughness characteristics of the channel at discharges less than 1,000 cfs, and/or (c) changes in the channel morphology that has altered the stage-discharge relationship. The specific reason for the change in stage-discharge relationship cannot be determined from the rating curve records. However, if the change can be attributed to channel aggradation, the aggradational trend indicates that channel fill is occurring at a sluggish rate and that the aggradational trend does not significantly effect the stage discharge relationship for discharges greater than 10,000 cfs.

Sedimentation Model

Two modifications were made to the TABS-1 model for this study. The first accounts for the presence of bridge piers and girders in the geometric calculations for additional wetted perimeter and reduced flow area at bridge cross sections. The calculations are similar to the "normal bridge" method in the HEC-2 Water Surface Profile model (US Army Engineer Hydrologic Engineering

Center, 1982). The second modification allows for increasing the maximum transportable grain size in the computations. The maximum transportable grain size in previous versions of TABS-1 was 64 mm. A significant amount of the bed material in the study area is of a diameter greater than 64 mm. Thirteen grain size classes between 0.0625 mm and 512 mm were used in this study.

Hydraulic Roughness

Estimates of the sediment transport potential are dependent upon the proper designation of the hydraulic roughness. The roughness coefficients used for the channel portion of the hydraulic calculations were estimated with the analytical roughness predictors contained in the "Hydraulic Design Package for Flood Control Channels" (SAM). The SAM algorithm was developed at the Hydraulics Laboratory and computes a Manning's n value based on bed material grain size distribution and the relative roughness of the bed material to the channel hydraulic conditions. The Manning's n is calculated using the roughness predictors developed by either Brownlie (1983) or Limerinos (1970). The Limerinos relationship is better suited to roughness prediction for coarse grained channels with no significant bed forms. The computed values of Manning's n are 0.029 for the river reach from Vista gage (RM 43.9) to the Steamboat Creek confluence (RM 45.2), 0.035 from Steamboat Creek to the Lake Street Bridge (RM 51.9), 0.030 from Lake Street Bridge to Arlington Street Bridge (RM 52.3), and 0.035 upstream of Arlington Street Bridge.

Determination of Inflow Sediment Load

Since bed-load measurements on the Truckee River at the upstream end of the study reach were not available, the sediment inflow to the study area was determined by calculating the equilibrium sediment transport rate for a range of water discharges for selected reaches of the Truckee River where sediment transport is in equilibrium with the bed materials. This reach of river is termed the "equilibrium reach." For gravel and cobble transport the portion of the Truckee River from approximately RM 50.5 to 51.5 exhibits similar characteristics in channel planform, bed material, and channel bank characteristics as that portion of the Truckee River immediately upstream of the study area. For bed-load transport of sand, the downstream limit of the study area (RM 43.9 to RM 45.3) is used for the equilibrium transport calculations. The locations of these reaches are shown in Figure 6.

Determining the equilibrium bed-load quantity requires an iterative application of TABS-1 on the equilibrium reach until the estimated inflowing load for all size classes to the equilibrium reach is approximately equal to the average transport capacity (by size class) in the reach with only minor amounts of aggradation and degradation. Excessive scour should not be computed in equilibrium reach calculations since excessive scour indicates that the coarse bed materials are being removed from the bed and the channel is not armoring. The bed gradation at the end of the equilibrium reach calculations should be similar to the bed material gradation at the beginning of the simulation. If this condition is met, then the bed material gradation is in equilibrium with the computed transport rate.

Equilibrium Bed-load Transport of Gravels and Cobbles

Several sediment transport functions were tested for the equilibrium bed-load

calculations. The Meyer-Peter and Muller (MPM) bed-load equation (Meyer-Peter and Muller, 1948) provided the most consistent bed elevation variation and uniform transport rate by size class for the equilibrium reach, and was chosen for this study. The equilibrium sediment load for discharges of 1,000, 10,000, and 30,000 cfs for a simulated duration of 30 days at 0.1-day time-steps was computed. The calculated equilibrium bed load of gravel and cobble in the equilibrium reach for discharges of 1,000, 10,000, and 30,000 cfs is shown in Figure 7.

Published research on critical shear stress values for gravel transport indicates that there is a wide variability in the recommended method and appropriate value of critical shear stress. The equation used in this study to calculate dimensionless shear stress is:

$$\tau^* = \frac{\gamma RS}{(\gamma_s - \gamma)D}$$

where τ^* is the dimensionless critical shear stress, γ is the specific weight of water, R is the hydraulic radius, S is the slope of the energy gradient, γ_s is the specific weight of the sediment, and D is the nominal grain diameter of the sediment particle. A practical application of the MPM bed-load transport equation in a steep gravel-bed river is given in Carson and Griffiths (1989). They determined that raising the value of τ^* in the MPM equation from 0.047 to 0.059 gave the best agreement for observed gravel transport on a braided gravel-bed river on a channel slope of 0.0048. Wiberg and Smith (1987) developed a method of computing τ^* based on grain size and near-bed hydrodynamic characteristics for coarse, mixed grain size bed sediments. Their results indicate that the value of τ^* for sediments with the characteristics of Truckee River riffles is in the range of 0.03 to 0.06. Boulder transport due to a dambreak flood surge on the Rubicon River in California was analyzed by Scott and Gravlee (1968). They determined the maximum tractive force at several locations by measuring the diameter of the maximum size boulder moved by the flood surge, the water surface slope, and the depth of flow. The value of τ^* for cobble and boulder transport measured by Scott and Gravlee varied between 0.025 and 0.08. Due to the variability of analytically computed and field measurements of τ^* on gravel and boulder bed rivers, and the cited references that indicate that critical shear stress should be increased for determining bed-load transport of gravel, the recommended value of τ^* of 0.047 in the MPM bed-load transport equation was retained for computing bed-load sediment transport.

Equilibrium Bed-load Transport of Sands

Negligible amounts of sediments smaller than coarse sand (0.5 mm) are present in the bed material samples upstream of the gravel and cobble equilibrium reach (RM 50.5 to 51.5). Thus bed material transport of sediments this size class and smaller could not be determined from the equilibrium reach calculations described in the previous paragraphs. To estimate the quantity of sediment smaller than coarse sand entering the upstream end of the study reach and moving as bed load in the downstream end of the study reach, additional calculations were performed for the Truckee River between Steamboat Creek (RM 45.3) and the Vista gage (RM 43.9). To be consistent with the mode (i.e., bed load) of sediment transport being simulated through the upstream

end of the study reach, the Meyer-Peter and Muller transport function was used. Additional fine sediment may be transported through the study area as suspended load, but sampled suspended load data indicate that the flow has sufficient energy to pass the suspended load through the study reach as wash load. The suspended sediment data sampled at the Sparks gage (RM 50.4), McCarran Boulevard (RM 47.7), and the Vista gage (RM 43.9) are compared in Figure 3. Regression analysis indicates insignificant differences in the water discharge versus concentration data for each gage. The hydraulic characteristics vary from a channel slope of approximately 0.005 at the Truckee River at Sparks gage to approximately 0.0005 at the Truckee River at Vista gage, indicating that the suspended sediment is actually wash load and does not need to be considered in the analysis of channel bed profile changes.

Bed material samples from channel bars at RM 45.6 indicate that 10 percent of the material is finer than 0.5 mm. Preliminary calculations using total load sediment transport equations (e.g., Tofalletti, Laursen) indicate that transport of material smaller than 0.5 mm would pass through the upper end of the study area as suspended load. Thus, the sediment load for medium sand, fine sand, and very fine sand determined in this reach can be added to the load calculated for gravel and cobble sediment inflow to the study area. The total (sand, gravel, and cobble) bed-load inflow rating curve developed using the MPM relationship is plotted in Figure 9. The equilibrium reach bed load was adjusted by comparing computed long term channel response with observed prototype channel conditions. The procedure by which the bed-load inflow transport rate was adjusted is described in the following paragraphs.

Long-Term Channel Response

A water discharge-exceedence frequency hydrograph that approximates the average annual hydrograph for the Truckee River at Reno for 1977-1986 was developed. The average annual hydrograph and water surface elevation at the Vista gage are tabulated below.

Average Annual Hydrograph Characteristics

Water Discharge, cfs	Exceedence Frequency	Vista Gage Elevation, feet
5480.	0.10	4378.5
2450.	0.20	4375.0
1410.	0.33	4373.5
707.	0.46	4373.0
375.	1.00	4372.5

Estimate of the Average Annual Bed Load

The hydrologic conditions for 1977-1986 for the study reach was simulated to compare computed with measured shifts in the stage-discharge rating curve for the Truckee River at Reno gage (Figure 5). Computed bed elevation changes at the end of the 9-year simulation period are plotted in Figure 10. The results indicate that extensive deposition occurs at the downstream portion of the study area. The computed rating curve shift for the 9-year simulation period for the Truckee River at Reno gage also indicates aggradation at the gage location in excess of the measured gage shift.

The simulation results using the equilibrium reach bed load (Figure 9) indicate that the computed bed-load transport into the study area is excessive. The total bed-load transport rating curve was adjusted downward until reasonable agreement between computed and observed sediment deposition quantities and zones, as measured by bed elevation changes through the study area after the 9-year simulation period, and comparable changes in both magnitude and trend in the observed shift of the stage-discharge rating curve for the Truckee River at Reno gage were obtained. The bed-load sediment inflow rating curve was reduced to 33 percent of the bed-load transport rate computed from the equilibrium reach. Further reduction of the bed-load inflow rate resulted in excessive scour at the upstream end of the model. The adjusted bed-load inflow rate used for existing condition simulations is plotted along with the equilibrium reach bed-load rating curve in Figure 9.

The computed rating curve shift for the Truckee River at Reno gage using the adjusted bed-load inflow rating curve is given as follows.

Water Discharge, cfs	Computed Shift, ft	Observed Shift, ft
707.	1.4	0.5
1410.	0.8	0.3
2450.	0.1	0.2
5480.	0.0	0.1

Computed bed elevation changes for the study reach after the nine year simulation period are plotted in Figure 11. Another way to display model results is to show the accumulated weight of sediment passing each cross-section. Computed accumulated bed load through the study reach after 1 year and after 9 years simulation time are plotted in Figure 12. Storage of bed material upstream of the three irrigation diversion drop structures between RM 49.1 and 50.1 during the first year of simulation is evident by the reduction in accumulated bed load in this region. This results in scour and associated increase in bed load downstream of the drop structures from approximately RM 49 to RM 48. After nine years simulation time, the storage of sediment upstream of the drop structures is filled and the channel bed is armored downstream of the drop structures, resulting in relatively uniform transport of bed load from the upstream limit of the study area at RM 53 downstream to approximately RM 48. Downstream of RM 48, a relatively uniform reduction in the total bed-load transport rate from RM 48 to RM 44 is computed, indicating sediment deposition and channel aggradation in this reach.

The 9-year simulation indicates that significant amounts of deposition occur downstream of McCarran Avenue Bridge at RM 47.7. Extensive bar development is not apparent in the prototype at this location. Sediment storage in the prototype is observed primarily between Highway 395 and the McCarran Avenue bridge. Possible reasons for this discrepancy between the computed and observed conditions are errors in the computed hydraulic conditions downstream of RM 47.7 resulting in computed sediment transport being too low, or errors in the computed hydraulic conditions between Highway 395 and McCarran Avenue Bridge resulting in computed sediment transport capacity being too high in this reach.

The 9-year sedimentation analysis of the study reach provides a means for estimating the average annual bed-load transport at several locations. The

TABS-1 results indicate sediment storage between RM 47.3 and RM 44.4. The average annual bed load at the upstream study limit, RM 47.3, RM 44.4, and the downstream study limit are given in Table 3. The computed average annual bed-load inflow to the study reach is approximately 51,000 tons per year. Sampled suspended sediment concentration combined with average annual flow-exceedence statistics yields an average annual suspended sediment discharge of 250,000 tons per year. The computed bed-load discharge is approximately 17 percent of the total sediment discharge. The computed average annual trap efficiency for bed load of the study reach is 71 percent.

Single Event Analysis

The 100-year recurrence interval flood simulation was completed to assess the quantity and location of sediment scour or deposition within the project reach for the design flood. To minimize the effects of initial bed material and bed profile gradations, the flood hydrograph was appended to the end of the 9-year simulation of average annual hydrographs. The differences in bed elevations and sediment scour or deposition between the end of the 9-year spin-up hydrograph and the 100-year flood hydrograph were used to quantify 100-year flood sedimentation parameters. Computed bed elevation changes at the end of the 100-year flood hydrograph are plotted in Figure 13. The accumulated bed load passing each cross section in the study reach at the end of the 100-year flood is plotted in Figure 14. The 100-year flood simulation results indicate the following:

    a. Passage of the inflowing sediment load through the downtown reach to approximately RM 51.8.

    b. Channel scour and associated increase in sediment load from 50.8 through 51.8. The scour is primarily due to channel adjustment and armoring to the high water discharge.

    c. Transport of the increased load to RM 49.

    d. Gradual trapping and reduction of the sediment load from RM 49 to RM 47.

    e. Transport of the reduced load out of the study reach.

SUMMARY AND CONCLUSIONS

Assessment of channel stability on instantaneous hydraulic conditions (e.g., shear stress, depth-averaged velocity, and stream power) does not consider long-term sediment transport conditions, and does not address the effects of sediment inflow to the study area on channel stability. Channel changes result from the balance of sediment transport into a reach, the sediment transport capacity of the reach, and the sediment availability in the reach. Simulation of the sediment transport through the project reach is required to address the sediment balance and subsequent channel stability.

Sediment transport theory was combined, using the TABS-1 numerical model, with evidence developed from field reconnaissance to quantify the bed-load inflow rating curve for the Truckee River at Reno, NV. The technique used is called the equilibrium reach method, and develops the sediment load based on sediment

transport theory, channel thalweg stability, and bed material gradation stability. Multi-year simulations using the equilibrium reach bed-load inflow resulted in excessive deposition, as indicated by the field reconnaissance and geomorphological analysis, at the downstream end of the numerical model. To compensate for this discrepancy, the bed-load inflow rating curve was adjusted downward to 33 percent of the equilibrium reach sediment inflow. TABS-1 simulation using the adjusted sediment load inflow rating curve resulted in reasonable agreement between observed and computed stage-discharge rating curve shifts; however, the model computed excessive sediment deposition downstream of the prototype sediment deposition zone. Additional adjustments are needed to reduce degradation between RM 53 and 51 and aggradation between RM 46 and 44. However, the study approach allows a reasonable estimate of annual sediment yield for flood control channel design. The bed-load inflow value of 51,000 tons per year may be high, but the design value is bracketed between that and the calculated outflow of 15,000 tons per year.

ACKNOWLEDGMENTS

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Table 1. Sampled suspended sediment concentration, USGS data base

Truckee River near Sparks, NV   USGS Gage No. 10348200

Date	Water discharge, cfs	Concentration, mg/l
5 Feb 1980	346	26
28 Feb 1980	790	23
12 Mar 1980	533	14
20 Mar 1980	480	14
4 Apr 1980	479	5
10 Apr 1980	563	9

Truckee River at Vista, NV   USGS Gage No. 10350000

Date	Water discharge, cfs	Concentration, mg/l
9 May 1979	1600	18
20 Dec 1981	7500	950
13 Mar 1983	5500	884

Table 2. Sampled suspended sediment concentration,
Truckee River at McCarran Boulevard.

Date	Water discharge, cfs	Concentration, mg/l
17 Dec 1985	347	5
21 Jan 1986	682	9
20 Feb 1986	14000 *	1345
25 Mar 1986	3430 *	1407
15 Apr 1986	2699	93
5 May 1986	2814	57
12 May 1986	2636	22
21 May 1986	2407	32
10 Jun 1986	1266	31
7 Jul 1986	628	32

* estimated water discharge from Truckee River at Reno USGS stream
gage records

Table 3. Average annual bed load of the Truckee River

	Bed load, Tons per Year			
Location	Sand	Gravel	Cobble	Total
RM 53.1	28,960	22,040	17	51,000
RM 47.3	24,420	20,130	0	44,550
RM 44.4	14,150	2,690	0	16,840
RM 43.9	12,820	1,940	0	14,760
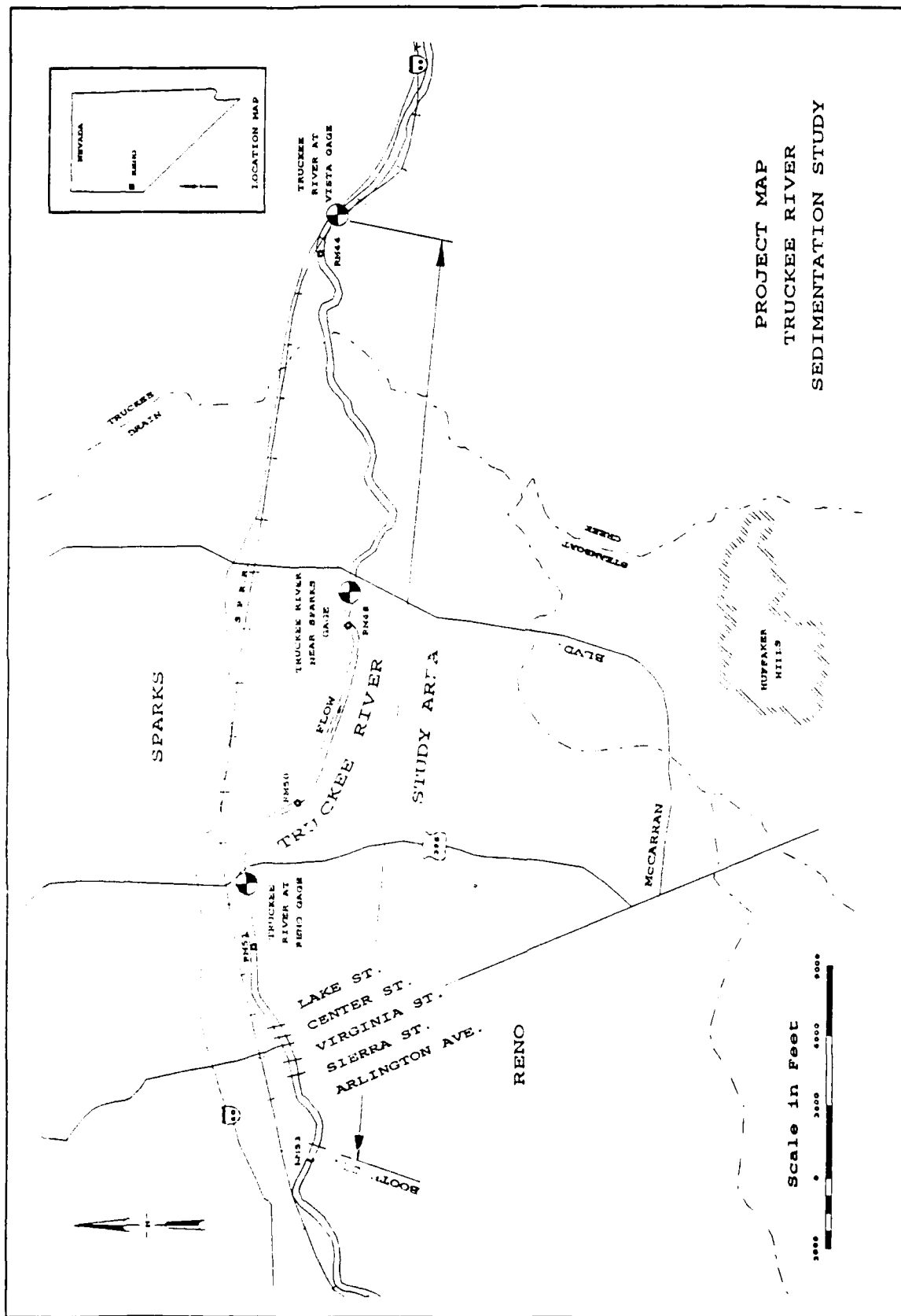
Figure 1. Vicinity map.
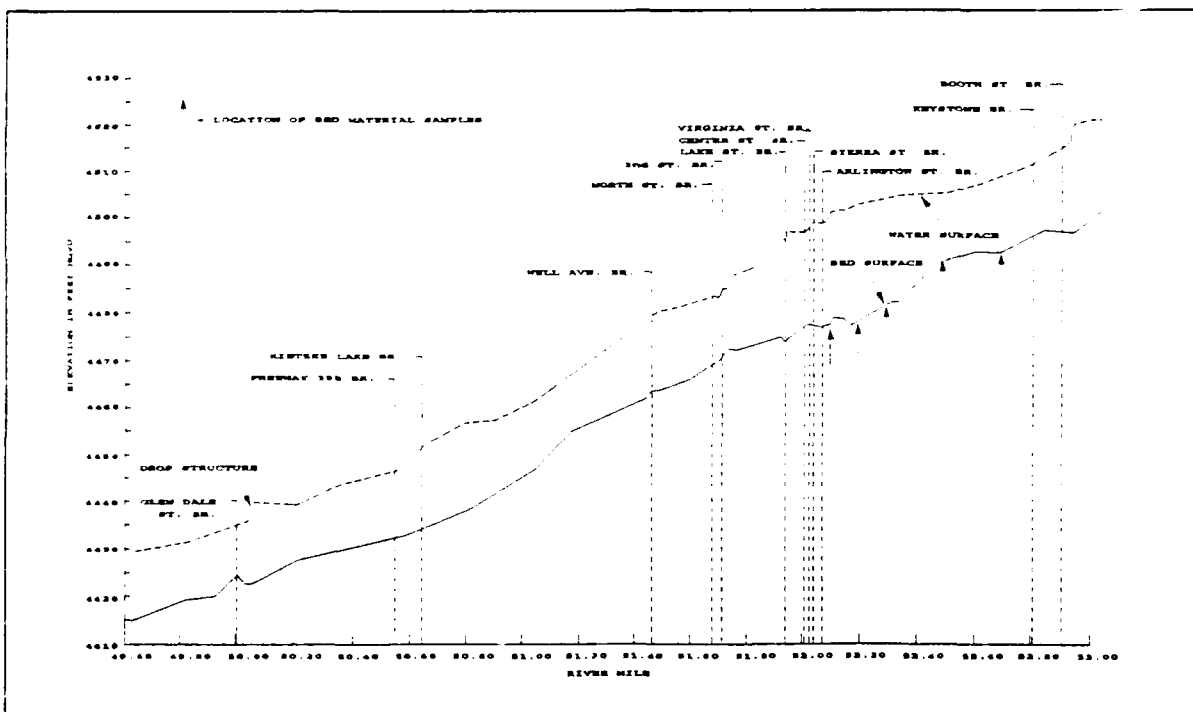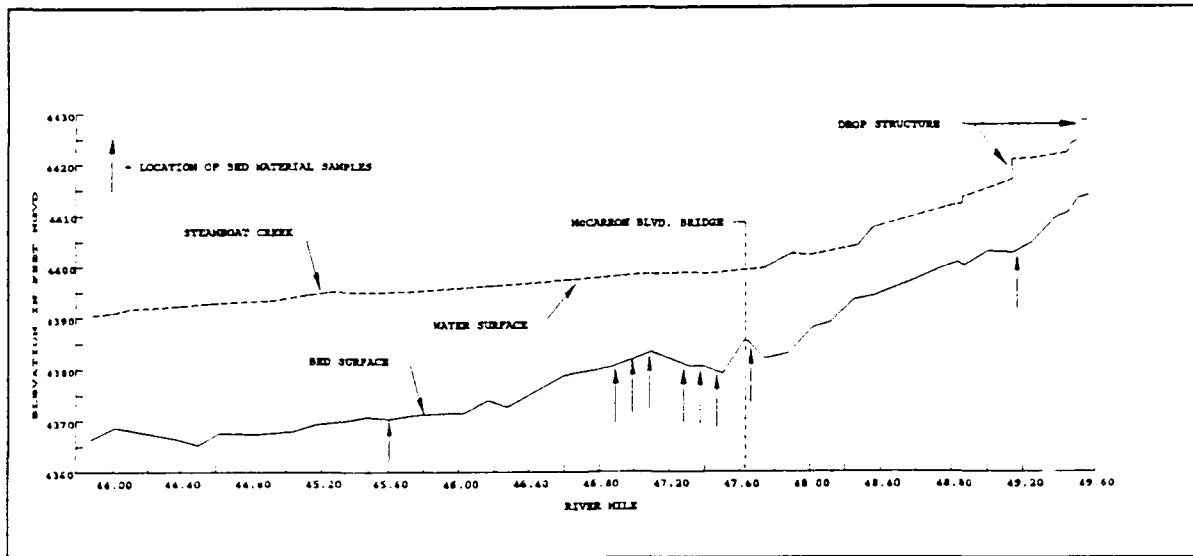
Figure 2. Bed material sample locations.
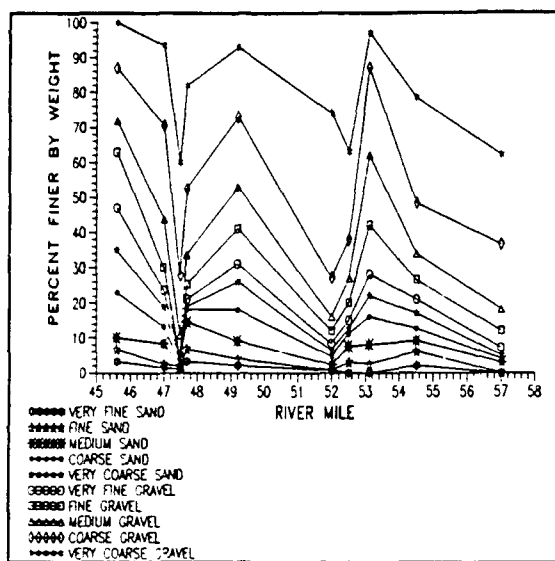
Figure 3. Bed material sample gradations.
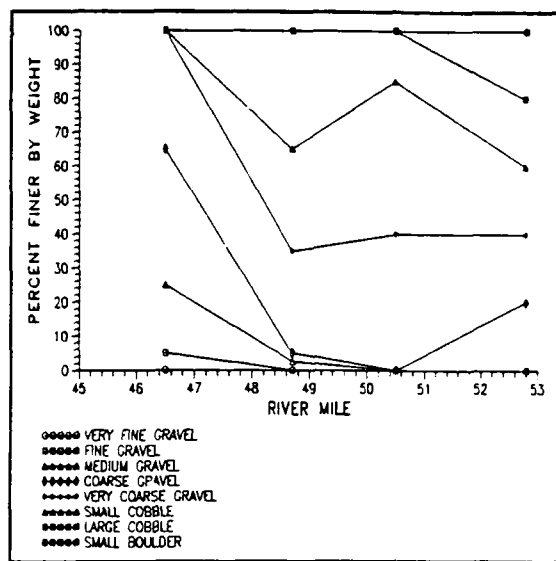
Figure 4. Wolman count material gradations.

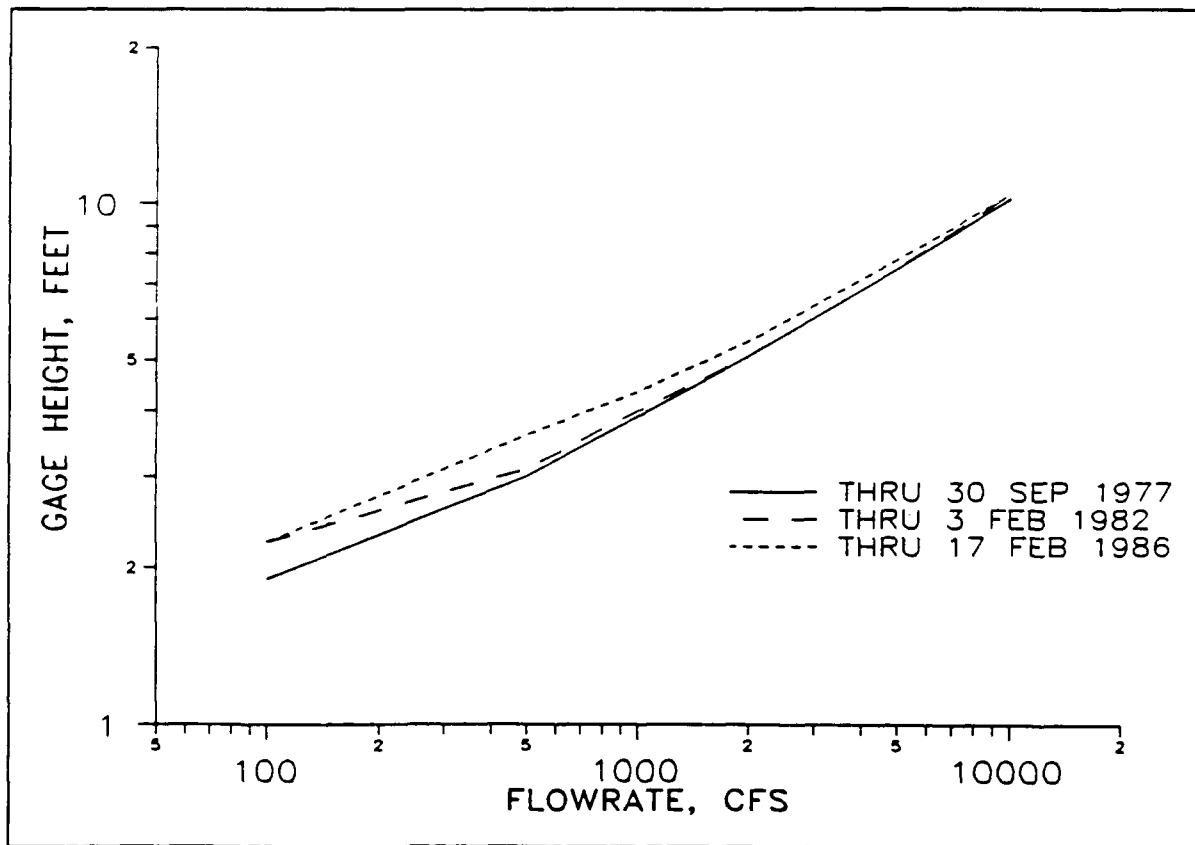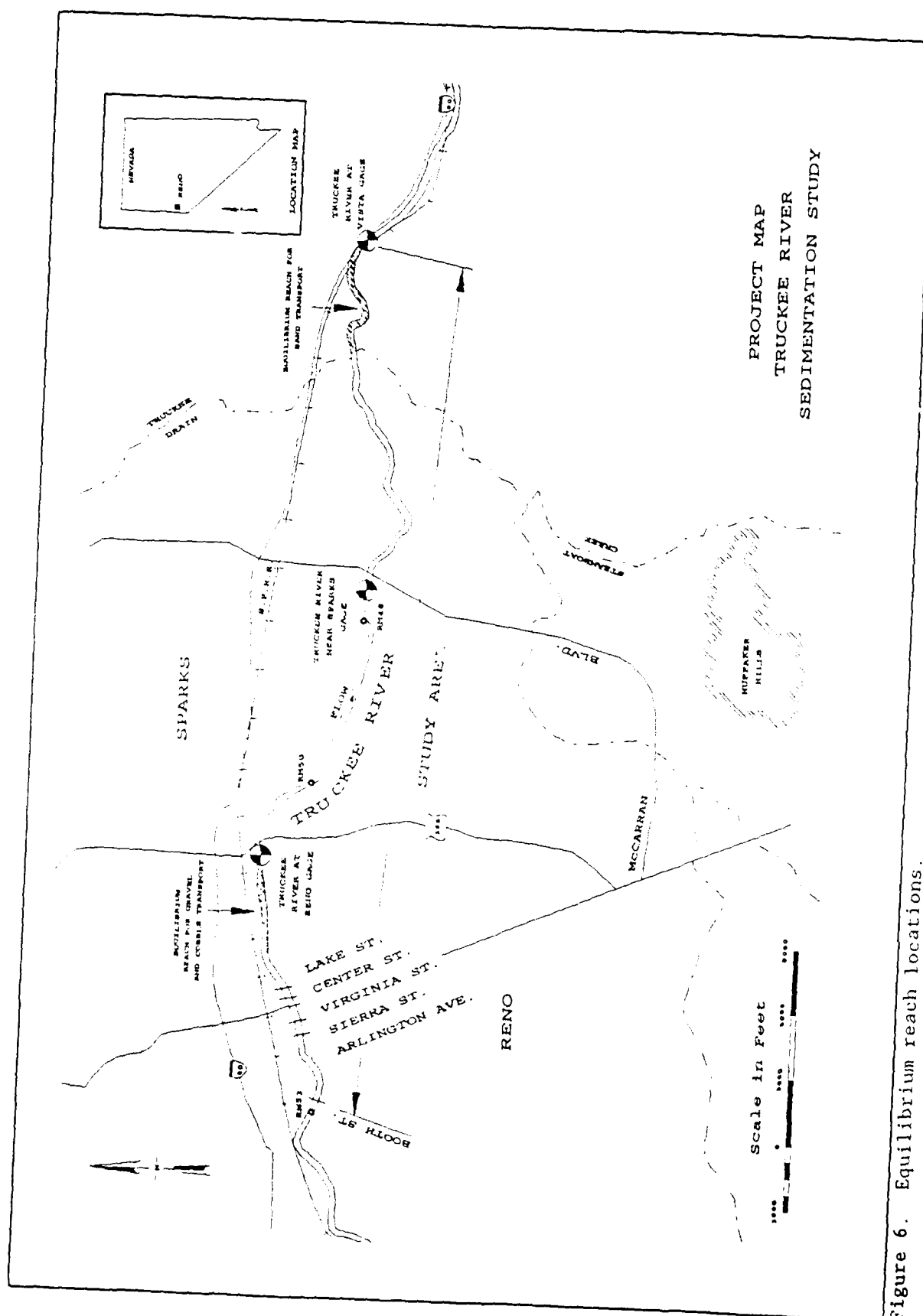**Figure 5.** Stage-discharge rating curve, Truckee River at Reno, Nevada

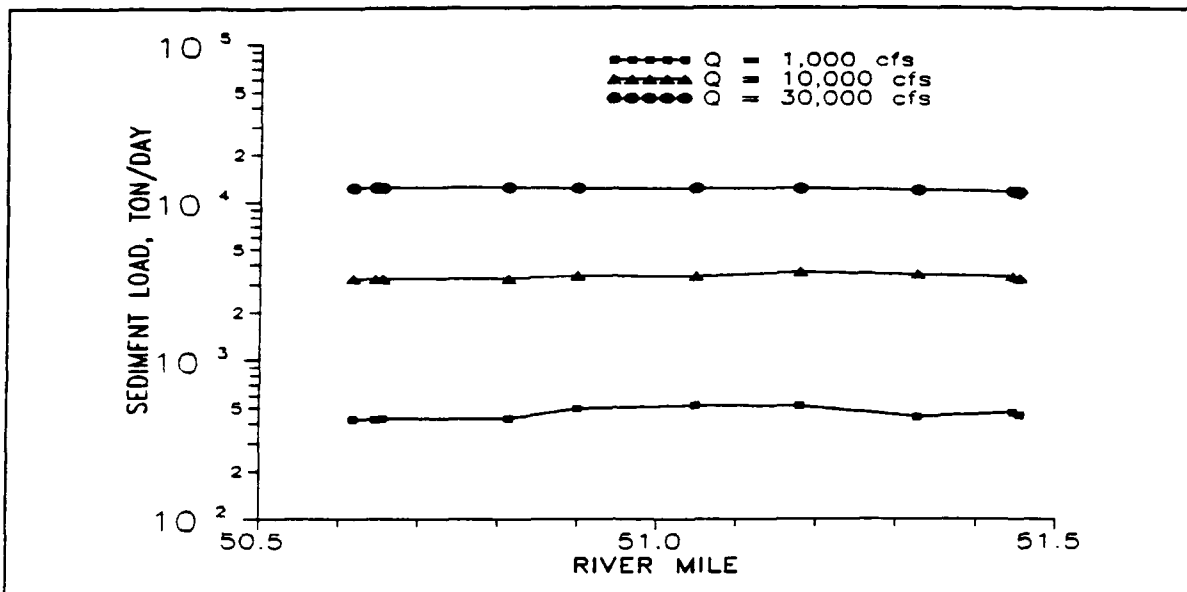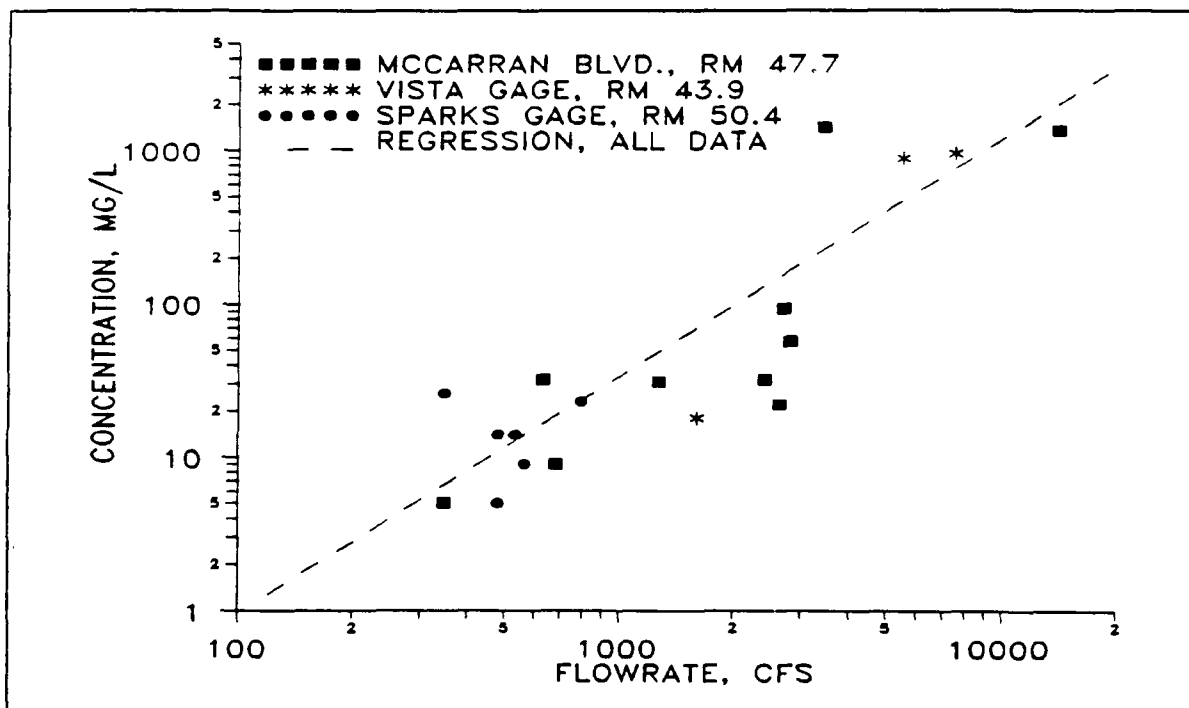Figure 6. Equilibrium reach locations.

Figure 7. Equilibrium reach bed-load transport.

Figure 8. Sampled suspended sediment concentration.
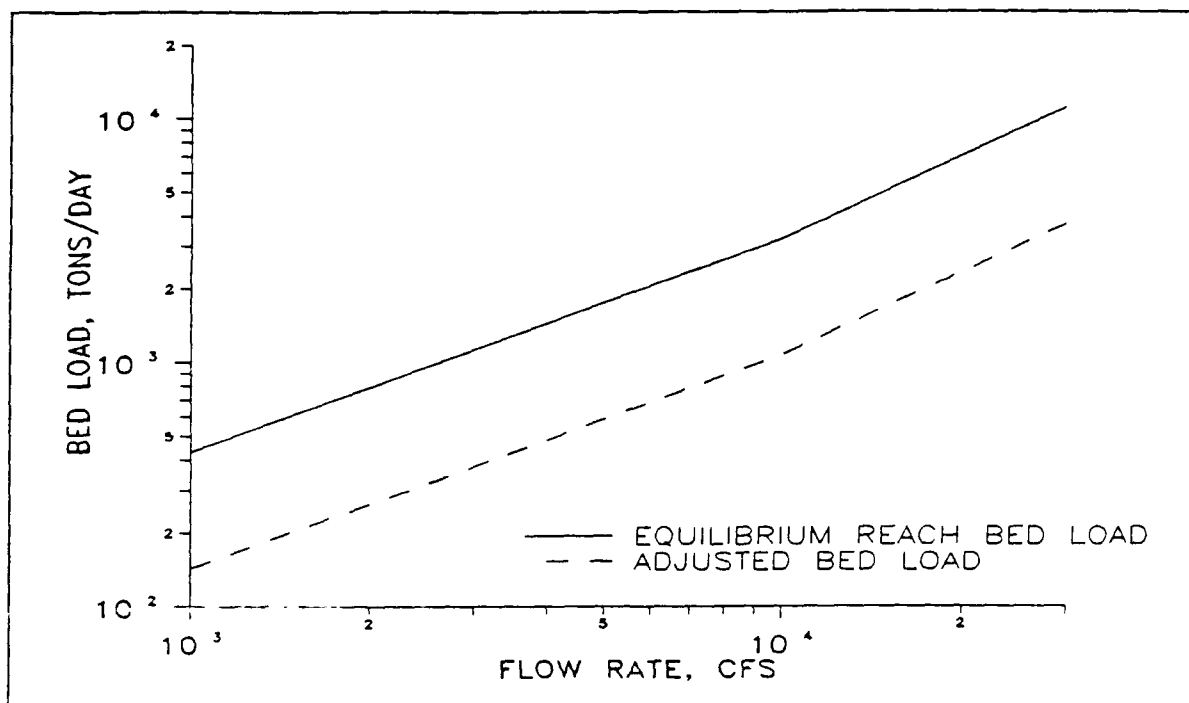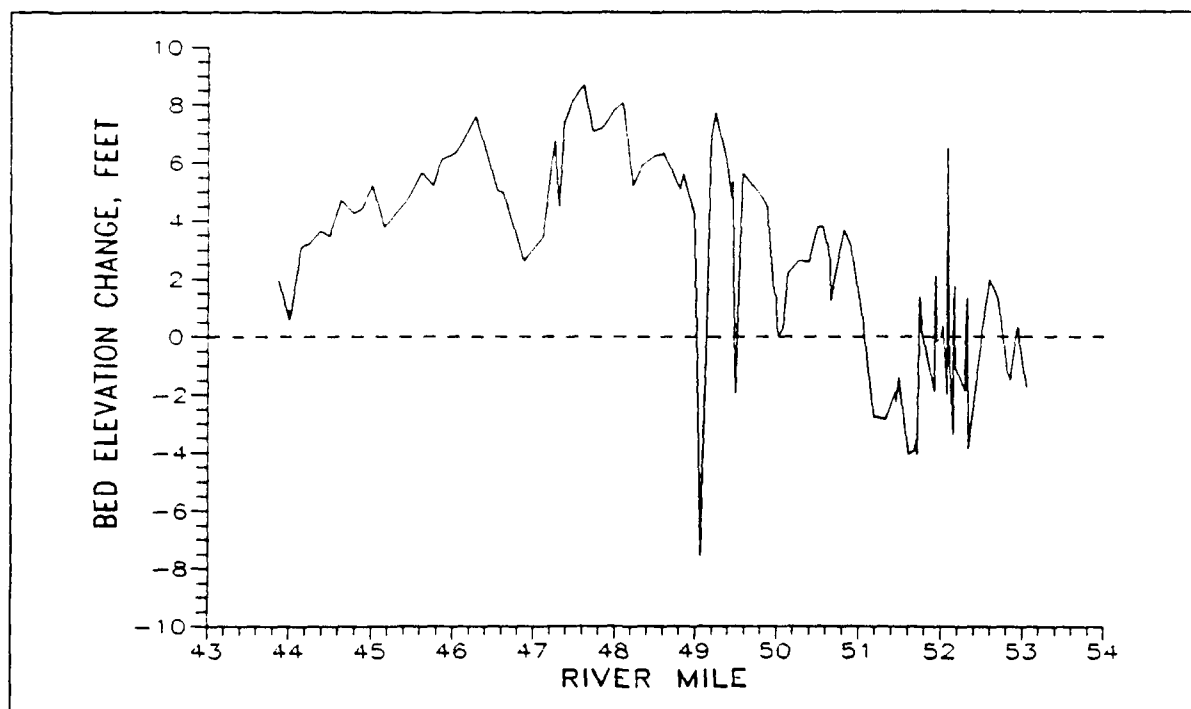
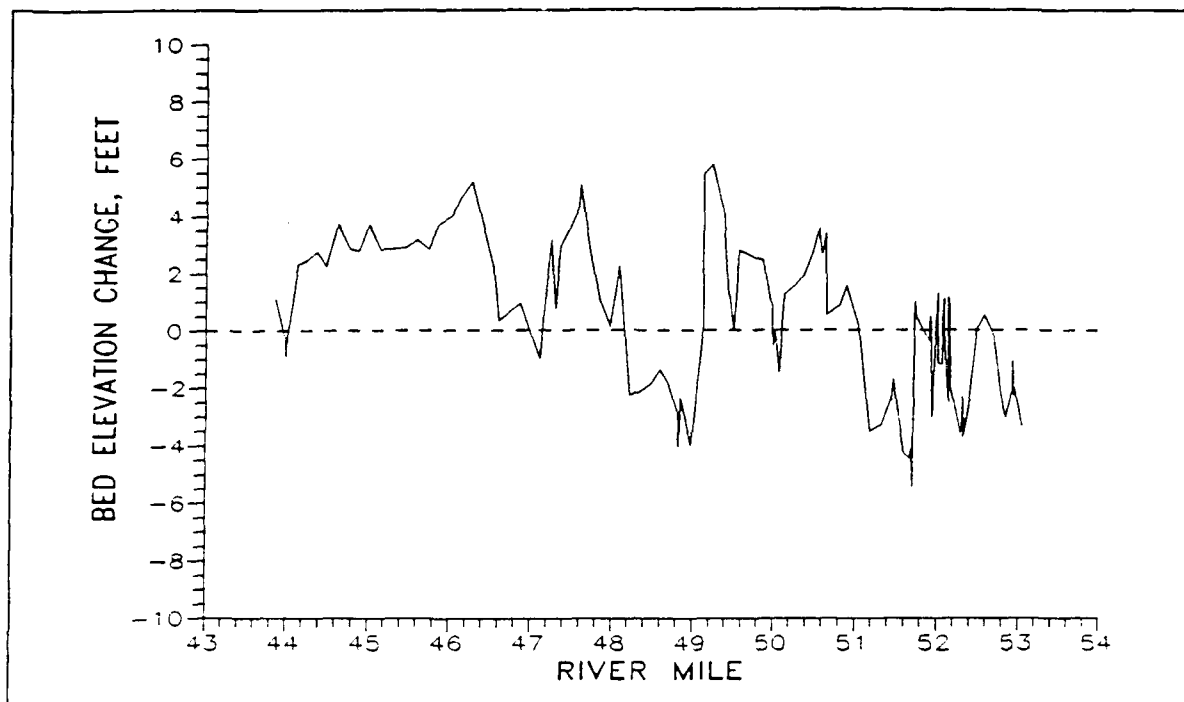Figure 9. Bed-load inflow rating curves.

Figure 10. Bed elevation change, equilibrium reach bed-load inflow.

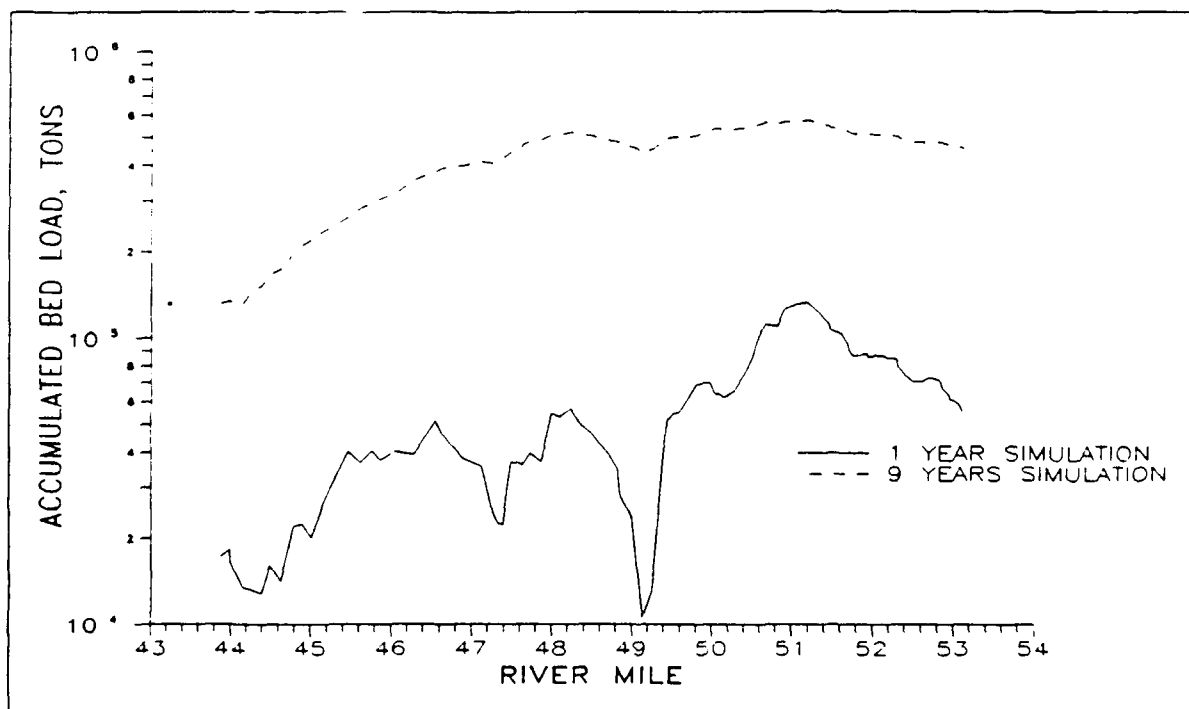Figure 11. Bed elevation change, adjusted bed-load inflow.

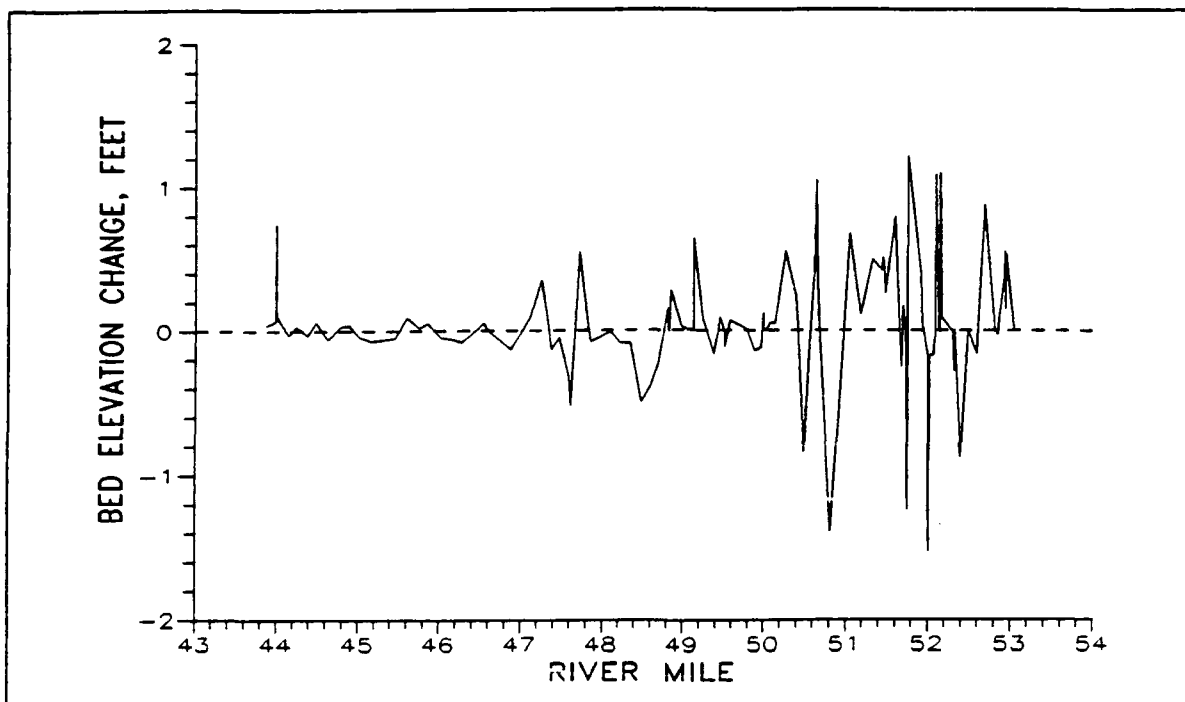Figure 12. Accumulated bed load, adjusted bed-load inflow.
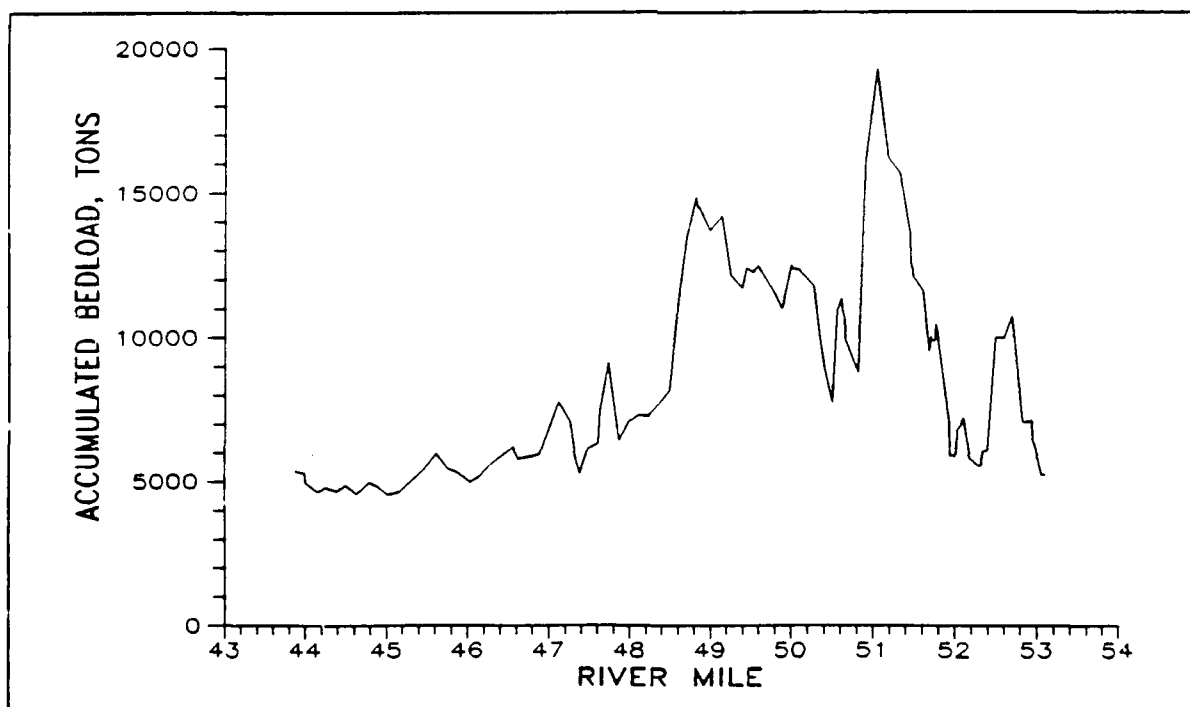
Figure 13. Bed elevation change, 100-year flood.

Figure 14. Accumulated bed load, 100-year flood.