UNIVERSITY OF SOUTHAMPTON

ALPINE PROGLACIAL FLUVIAL SEDIMENT TRANSFER

Jeffrey Warburton

Thesis submitted for the Degree of Doctor of Philosophy

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ABSTRACT

FACULTY OF SCIENCE

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ALPINE PROGLACIAL FLUVIAL SEDIMENT TRANSFER

by Jeffrey Warburton

This thesis considers sediment yield and transfer processes in three glacier basins in the Val d'Hérens, southern Switzerland. Sediment yields were estimated for the 1986 ablation season for the: Bas Glacier d'Arolla, Haut Glacier d'Arolla and Glacier de Tsidjiore Nouve. Yield estimates were derived from continuous discharge and suspended sediment concentration series combined with estimates of bedload transport. 1986 sediment yields showed Tsidjiore Nouve to have the highest yield (4174 $t.km^2$), Bas Arolla was second (4093 $t.km^2$) and Haut Arolla the lowest (2422 $t.km^2$). Bedload accounts for 58-64% of the total load at Bas Arolla and 36-51% at Tsidjiore Nouve. The three basins accounted for 86% of the measured sediment output from the Val d'Hérens region.

Detailed field study of sediment transfer processes, sediment yield and channel change during the 1986 and 1987 ablation seasons in the proglacial zone of the Bas Glacier d'Arolla permitted evaluation of proglacial sediment sources and yield. Valley train sediment transport dominated proglacial sediment transfers. Tributary streams and hillslopes contributed minor amounts of sediment. Channel change was episodic and switched from single-thread to multi-thread to single-thread over the ablation season. Meltwater floods were responsible for catastrophic changes in channel form. Thalweg behaviour during floods controlled channel form and sediment yield.

Intensive field measurements (25th May to 30th July) were sufficient to define a proglacial sediment budget at Bas Glacier d'Arolla. The budget indicated that proglacial sediment sources contribute 23% to the total basin sediment yield. The overwhelming proportion (95%) was derived from the valley train during the brief period of meltwater flooding July 15th-18th 1987. Expressed in terms of work and power, the budget suggests that 4 basic process-morphologic subsets can be distinguished: (1) a valley train-channel subset; (2), valley train margin subset of quasi-continuous processes; (3) a subset of hillslope processes, and (4) slopewash.

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Even with such able assistance any errors or inequalities in this thesis are entirely of my own creation.

Finally, to Mum, Dad, John and Janet thanks for everything.

Chapter 1.

INTRODUCTION

1.1 Introduction

1.1.1 Background

Glaciated areas play an important and unique role in the hydrology of alpine areas, as is highlighted by the number of international hydrological symposia and publications which have focussed on this theme (e.g. Young, 1985). In Alpine areas meteorological, glaciological, and periglacial phenomena combine with hydrological processes in complex associations (Roots and Glen,1982). Equal attention must be paid to the sediment system, which is inseparable from hydrology in the glaciated catchments (Figure 1.1).

Alpine proglacial sediment transfer poses a major intellectual and technical challenge to the geomorphologist, and at the same time represents a very practical challenge to those whose task it is to utilise and manage the water resources of the mountain catchments. It is these intellectual, practical and technical challenges which form the focus and motivation of this thesis.

The apparent bias towards the study of the hydrological aspects of alpine catchments can be explained by programmes such as the UNESCO-sponsored International Hydrological Decade (IHD - 1965-1974); the International Hydrological Program (IHP); and the International Association of Hydrological Sciences (IAHS) which stimulated and encouraged national hydrological monitoring programs. This investment has been rewarded: "In the last 20 years very considerable advances have

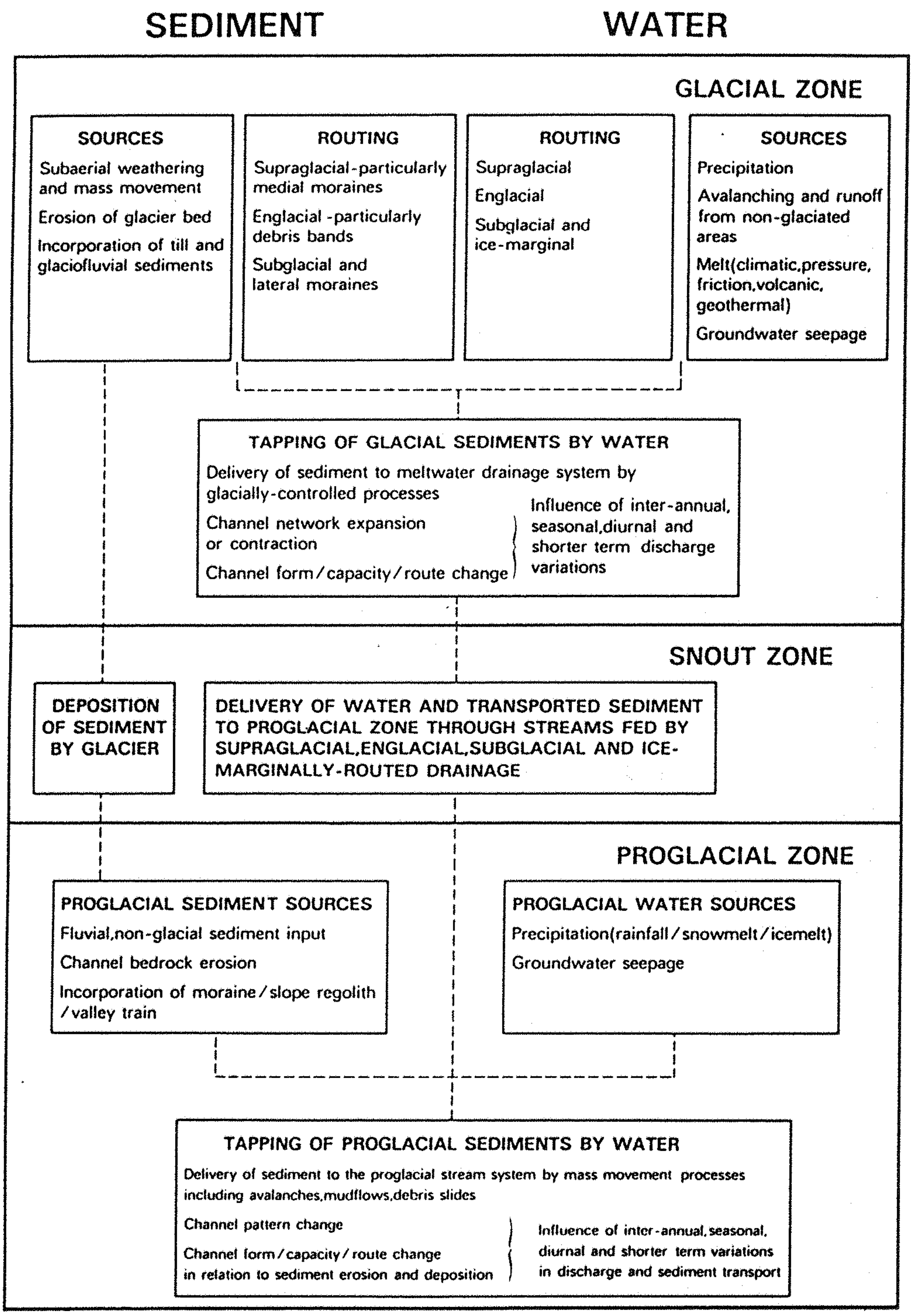


Figure 1.1 Description of the basic structure of the glacio-fluvial sediment system in an alpine glacierised catchment (from Gurnell and Clark 1987).

been made in our understanding of hydrological processes in high mountain areas" Young (1985). This however, is not true of sediment based studies: as Gurnell (1987a) points out there is a 'dearth' of alpine glaciated catchments where integrated studies of the sediment system or even just the monitoring of total sediment yield has been undertaken. The principal task of this thesis is to fill this gap by providing estimates of total meltwater-stream sediment yield, and by assessing the importance of fluvial processes in modifying these yields. Interest in the glaciated catchment water-sediment system is also justified because of the use of glacierised areas for electric power generation and because of the problems of the management of environmental hazards (e.g. floods on meltwater rivers; bursting of glacially and sub-glacially impounded lakes; and damage by sediment load to turbines) (Embleton, 1982).

Despite several published volumes which have examined the sedimentary deposits, landforms and processes of glacierised catchments (e.g. Price, 1973; Jopling and MacDonald, 1975; Davidson-Arnott et al., 1982; Eyles, 1983) an integrated approach, examining the water and sediment system as a whole, has rarely been attempted. The most noteworthy exceptions are Church (1972), Østrem (1975), Hammer and Smith (1982), Fenn (1983) and Gurnell and Clark (1987). This latter volume, Gurnell and Clark (1987) "Glacio-fluvial sediment transfer - an alpine perspective" is particularly relevant to this thesis since it emphasised alpine catchments; summarises previous research on the Tsidjiore Nouve and Bas Glacier d'Arolla catchments (both studied in this thesis); and provides the most recent and complete synthesis of the subject.

1.1.2 Research Aims

The principal aim of this thesis is to assess the importance of fluvial processes in transporting sediment within alpine proglacial zones. Fenn (1983), working in the same area, attempted to specify the general nature of proglacial streamflow, suspended sediment transport and channel form dynamics. This approach differs from Fenn (1983), in that it attempts to investigate sediment transport directly and to quantify the various sources of proglacial sediment which contribute to the sediment load of the meltwater stream. In this respect it is a natural progression from Fenn's work in that it attempts to provide a physical interpretation of the proglacial streamflow and sediment output series. Four fundamental questions are considered. These are inevitably handled in a site specific manner, but their implications are applicable to the Alpine proglacial zone as a whole:

1) What are the proportions of bedload, suspended load and solute load in the main meltwater channel?

2) How is the sediment load in a proglacial stream modified by contributions from sediment sources along the valley train and what are the proglacial fluvial sediment sources?

3) What are the consequences of sediment transport in modifying the morphology of the proglacial zone?

4) What are the major transport processes, storage elements and linkages in the proglacial fluvial sediment budget?

Alternatively, the above four questions could be summarised under four respective headings: total sediment yield, sediment source, channel change, and the sediment budget. This is clearly an over simplification, because all four themes are intimately related. These basic themes are in accordance with the recommendations for further research in alpine catchments specified by Fenn (1983, p.406) who stated " There is a particular need for more determinations of the absolute volumes of bed, suspended and dissolved load at selective sites (at the snout, and at various distances from the snout) in glacial streams, so that estimates of sediment yield from particular contributing zones may be more accurately based". Fenn (1983) also suggested that the inter-relationships between proglacial channel form, flow and sediment discharge (i.e. the whole proglacial process-form response system) should be investigated. Answers to the above questions are of importance to alpine hydro-electric schemes in relation to both the management sediment entering galleries and pumping systems and the problems created by highly mobile proglacial channels diverting around water intakes. These problems, the management of sedimentation and the control of floods, often through channel improvements and bank protection are two of the general problems faced by river engineers in many environments and are also areas of uncertainty (White, 1987). What is needed is greater understanding of the physical processes relating to:

a) The quantities and nature of sediments entering river systems.

b) The movement of heterogeneous sediment mixtures.

c) The erosion of river banks as influenced by the nature of materials forming the bank e.g. vegetation.

d) The importance of unsteady flow on sediment transport and channel form.

e) Prediction of channel planform and its change through time.

.... "solutions to these problems will be greatly valued by the engineering profession" (White, 1987).

The immediate objectives of the project are thus very specific, but the underlying aims are of great generality.

1.1.3 Basis of the Sediment Budget Approach

In fluvial geomorphology there is a strong emphasis on the analysis of the yield or 'output' of sediment from drainage basins. This dates back to early uses of stream load data to provide initial estimates of subaerial erosion rates. For example, Everest as early as 1831 made one of the first quantitative assessments of river load by calculating the amount of material transported by the Ganges (Chorley et al., 1964, p. 280-281). In glacierised catchments streamload has long been used as a measure of glacial erosion; Scheuchzer (1723) (In Charlesworth, 1957) remarked that "glacier streams were thickly loaded with fine mud that renders them turbid and milky" but quantitative statements regarding the load came latter. Helland (1877), working on the Jostedalsbreen Icefield in Norway, made one of the earliest estimates of glacial erosion using measures of the sediment yield of streams. Based on measurements of the quantity of water in rivers and the amount of mud carried away from the glacier, Helland estimated that 6900 m3 were removed from the 870 km^2 basin annually. Although initially useful, these measurements were of limited value since they ignored transport of bedload, subglacial storage, the importance of non-glacial sediment sources and the role of the proglacial zone in regulating sediment transport (Wright, 1937). Charlesworth (1957, p. 220) stated these limitations concisely - "glacier-milk is but an insecure foundation upon which to base any quantitative conclusions, especially if the data have not been collected by continuous readings taken at all seasons and from the whole basin". Even so, the method still remains intact, because measurements and calculations are relatively simple. Better estimates of erosion, as this thesis will show, involve far greater effort.

Sediment output data are only of limited value in denudation studies because the role of storage in modifying the relationship between erosion and yield, is rarely evaluated (Richards, 1987). The monitoring of drainage basin sediment yield provides only an integrated 'black-box' view of how sediment output responds to average conditions. Monitoring of processes and the construction of a sediment budget can provide additional information about the sediment system. This is especially important at the upper ends of river systems where there is a sensitive balance between sediment supply and transport (Bathurst et al., 1986a).

A sediment budget is a quantitative statement of the rates of production, transport and discharge of detritus (Swanson et al., 1982). Closely related to the budget is sediment routing which in its strictest sense is the computation of movement of sediment through a series of landscape units and the modification of sediment during the passage (Swanson et al., 1982). More generally it is the concept of sediment movement through a landscape unit. Application of the sediment routing approach has been rare in drainage basin studies. One example is Swanson and Friedrikson (1982) who examined the impact of forest practices on sediment routing in a small drainage basin in the Western Cascades, Oregon.

The sediment budget approach provides a useful framework for understanding sediment movement through a catchment or landscape unit. Dietrich et al. (1982) define three prerequisites in constructing a sediment budget. These are firstly, recognition and quantification of transport processes; secondly, recognition and quantification of storage elements; and thirdly, identification of linkages amongst transport processes. Although conceptually elegant, this approach is weakened by sampling problems; lack of defined processes; changing boundary conditions and process assemblages; problems in measuring the rates of transfer processes; the definition of recurrence intervals (episodic processes); and the lack of appropriate process models. To some extent, conceptual generalisation of the sediment budget can overcome limitations in study time (changing boundary conditions) and sampling problems. It is essential to establish at an early stage an approximate budget (based on field observations and/or a pilot project) which will greatly simplify field operations.

The goal of any sediment system study is a sediment yield model based on physical processes. However, before this goal can be achieved, progress through a hierarchy of development stages or levels of sophistication must take place. Three stages can be identified. The first stage, the identification and definition of the basic processes, has largely been achieved through catchment studies over the past 25 years (e.g. Rapp, 1960). Secondly, construction of sediment budgets provides an understanding of the linkages between transport and storage (e.g. Dietrich and Dunne, 1978). Thirdly, the development of physically-based sediment routing and sediment yield models (Simons et al., 1982). Similarities between Kuhn's view of science (Kuhn, 1961) and these stages are evident. The three levels can be thought of as the entitation and quantification phase; the budgeting phase (both levels falling within the inductive stage specified by Kuhn); and the modelling phase. Currently the emphasis is on sediment budgets which is not surprising because "Before we can develop sediment yield models, and a better understanding of the linkage between erosion and sediment transport, we must improve our observational data base and process understanding" (Hadley and Walling, 1984). It appears that the prevailing Kuhnian paradigm with regard to alpine sediment systems is that of the sediment budget. It remains to be seen whether the shift to a modelling dominated view will be the product of internally generated evolution or an externally-forced 'revolutionary' paradigm replacement.

Sediment budgets are also useful in uniting contemporary process studies with traditional interests in landscape evolution (e.g. Davis, 1899) A sediment budget, although constructed from a relatively short run of data provides a framework for studying the sediment system over longer periods of time, especially when residence times can be evaluated through the application of dating methods. However, storage remains the most essential but poorly understood part of any geomorphic system.

The sediment budget framework also has the advantage of defined linkages which provide the basis for a study of energy flows between structural components. This approach could provide a basic integrating theme in physical systems (Gregory, 1987), and in the alpine context may be particularly useful in process studies where sediment transport is an easily-observable surrogate for energy - at least when not supply limited (Caine, 1976; Barsch and Caine, 1984).

One major drawback with the sediment budget approach is that it is based on an a priori model which is essentially deductive and therefore necessitates a rigorous definition of the sediment budget. How well a budget quantifies and characterises a geomorphic system depends on how well transfers, storages and linkages are specified; the precision and duration of field measurements; and the scale of spatial and temporal segmentation of the basin under analysis (Swanson et al., 1982). Chapter 2. provides a more detailed discussion of the limitations specific to this project. In most studies a complete budget is not always specified. There is usually a semi-quantitative fulfilment, with only the general form of the budget being established and the magnitude of the components measured. This is still a worthy achievement, since it will form a valuable basis for further studies and revision of the general budget/model structure. Therefore, in formulation a budget may be essentially inductive, but once structured it may form the basis of a deductive framework - a not uncommon pattern in field science.

1.1.4 Defining an alpine proglacial sediment budget structure.

The purpose of this section is to outline the sediment budget model used in the present study. Ideally the construction of a sediment budget should involve a pilot study which recognises and quantifies approximately the major processes that generate and transport sediment. In this study the sediment budget model (shown in Figure 1.2) was generated from field observations and measurements, carried out in Switzerland in 1985 and 1986, and from conceptual models proposed by Caine (1974), Dietrich and Dunne (1978) and Fenn (1983). Figure 1.2 divides processes into glacial, slope and fluvial sub-systems. The interaction between glacial and slope sub-systems is omitted from Figure 1.2 for simplicity. This thesis is principally concerned with the valley-train fluvial elements (highlighted on the budget diagram by the heavy box) and as such is a component of the holistic model. This model is descriptive rather than normative since it is based on field observations in the Arolla valley. If the model had been generated in the Ferpècle Valley, 7 km away, a proglacial lake would have been included in the system. Nevertheless it is proposed that the model has considerable generality, and could easily be adapted to other proglacial catchments.

The decision to examine just the proglacial fluvial component of the sediment budget is justified from a number of standpoints. Firstly, non-fluvial slope processes tend to be episodic on a long timescale, and cannot always be adequately characterised by direct

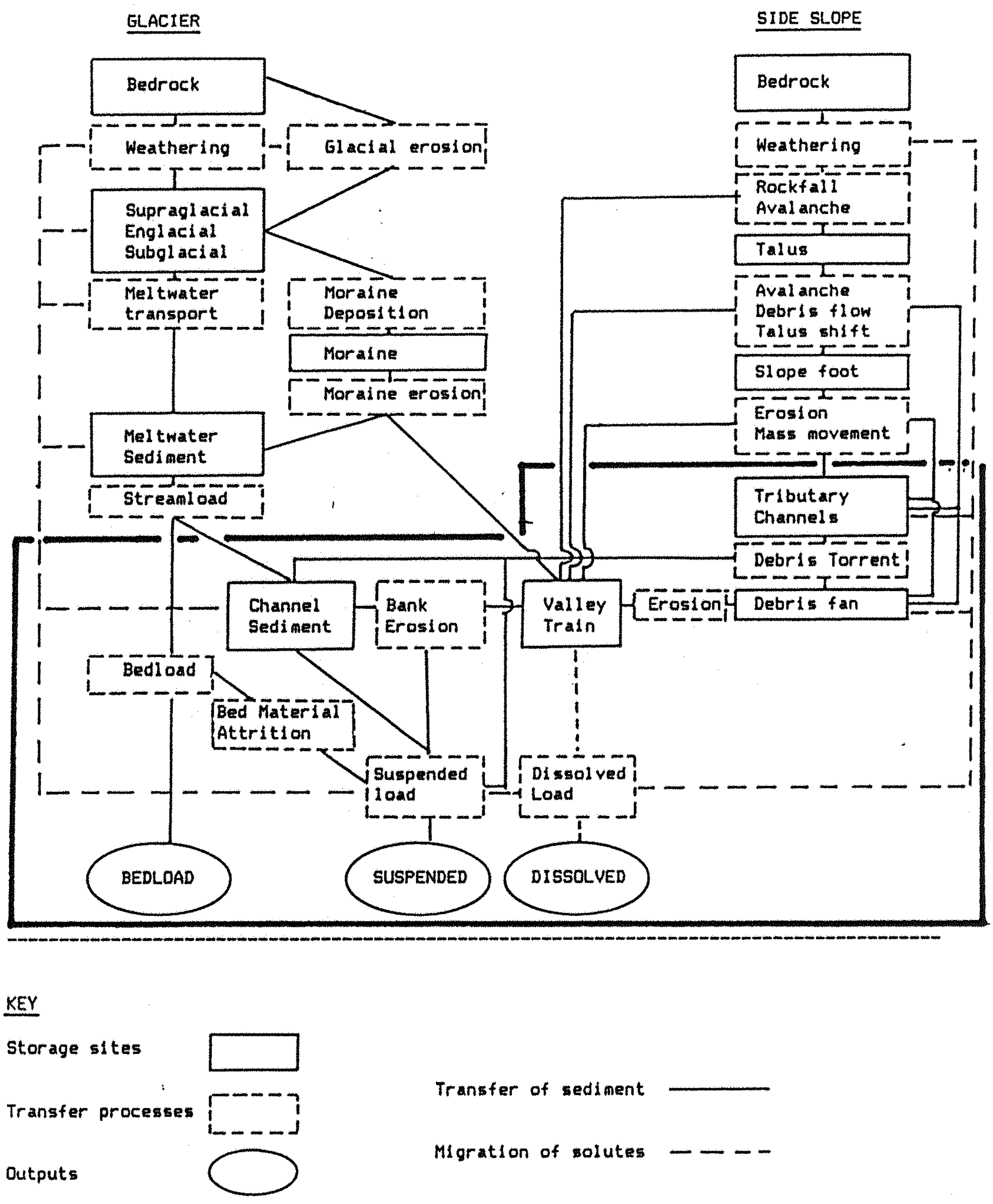


Figure 1.2 Preliminary sediment budget model for alpine proglacial sediment transfer processes. (Based on the conceptual models of Caine (1974), Dietrich and Dunne (1978) and Fenn (1983) and field observations in the Bas Arolla catchment, Switzerland).

techniques within a two year period of study. Secondly, contemporary glacial processes are not easily investigated since the major sedimentary processes are active within or under the ice and would thus demand detailed examination and inference in their own right. Such a study would justify a project in its own right. Thirdly, there exists a reasonable understanding of proglacial fluvial processes, and data are available for the catchments studied in this project. Finally, it can be argued that the proglacial fluvial network forms a distinct physical sub-system (Figure 1.2). Just as a drainage basin is defined as a fundamental geomorphic unit (Chorley, 1969) the drainage zone of the proglacial area may be thought of in a similar way. It is still a landscape unit which collects, concentrates and promotes the movement of water and sediment (Roehl,1962; Derbyshire et al., 1979) but, in addition, it is also the landscape unit that transmits the products of glacial erosion. The proglacial zone can therefore be visualised as a truncated or beheaded drainage basin whose upstream limit is the ice margin (Figure 1.5). Alternatively it could be argued that the fluvial network extends beneath the ice towards the topographic limits of the glacierised catchment. Definition of formal boundaries is necessary in order that system elements and processes can be recognised and quantified (Clark, 1987a). If this is done, a sediment budget can be at any scale: catchment wide, stream reach or even within channel (aggregate bar movement). For the purpose of this study, the drainage area of the proglacial zone is treated as a bounded unit.

In specifying the budget, emphasis should be placed on transport rates rather than processes, because the latter are often difficult to define, e.g. it is difficult to explain how fine sediment deposited in a bedload trap got there. Even so computation of transport rates are in themselves highly variable and as such should be treated cautiously.

The proglacial sediment budget model proposed in this study is temporally constrained both on long and short-timescales. Clearly, from two seasons data long-term extrapolation is not justified and in the short-term, because data have only been collected during a part of the ablation season (May-September), the full annual sediment flux cannot be constructed. Having said this the majority of sediment load activity takes place during this period, and alternative methods (e.g. assessing the rate of purging of Grande Dixence sediment traps) can be used to produce longer-term 'proxy' sediment records.

1.2 Study Areas

1.2.1 Background

The data on which the present study is based were collected in two May - September field seasons in 1986 and 1987. During 1986 the sediment discharge from three catchments (Glacier de Tsidjiore Nouve, Bas Glacier d'Arolla and Haut Glacier d'Arolla in the Pennine Alps of Valais, Southern Switzerland) were measured. In 1987, a more detailed investigation was undertaken on the Bas Glacier d'Arolla proglacial zone in order to provide data for the development of a proglacial fluvial sediment budget. This hierarchical approach was thought to offer the best design, in the absence of resources for a more comprehensive monitoring program. This section provides a description of the study area in four parts: an outline of the general characteristics of the research area with detail of the research catchments; a summary of the main features of the alpine proglacial zones with detail of the research sites; a description of the Grande Dixence Hydro-electric scheme whose management has relevance to the estimation of longer-term sediment yields (Chapter 3); and a synopsis of previous work in the area.

1.2.2 Research Area

The Arolla valley, where this research is based, is the western branch of the Val D'Hérens, Valais, Switzerland (Figure 1.3). The main river, the Borgne d'Arolla, is fed by the Arolla glacier complex which includes two of the glacierised catchments investigated in this research; the Haut Glacier d'Arolla and the Bas Glacier d'Arolla. Two other major glaciers, Glacier de Pièce and Glacier de Tsidjiore Nouve (the third glacier investigated in this research), contribute meltwater from the western side of the Arolla valley (Figure 1.3).

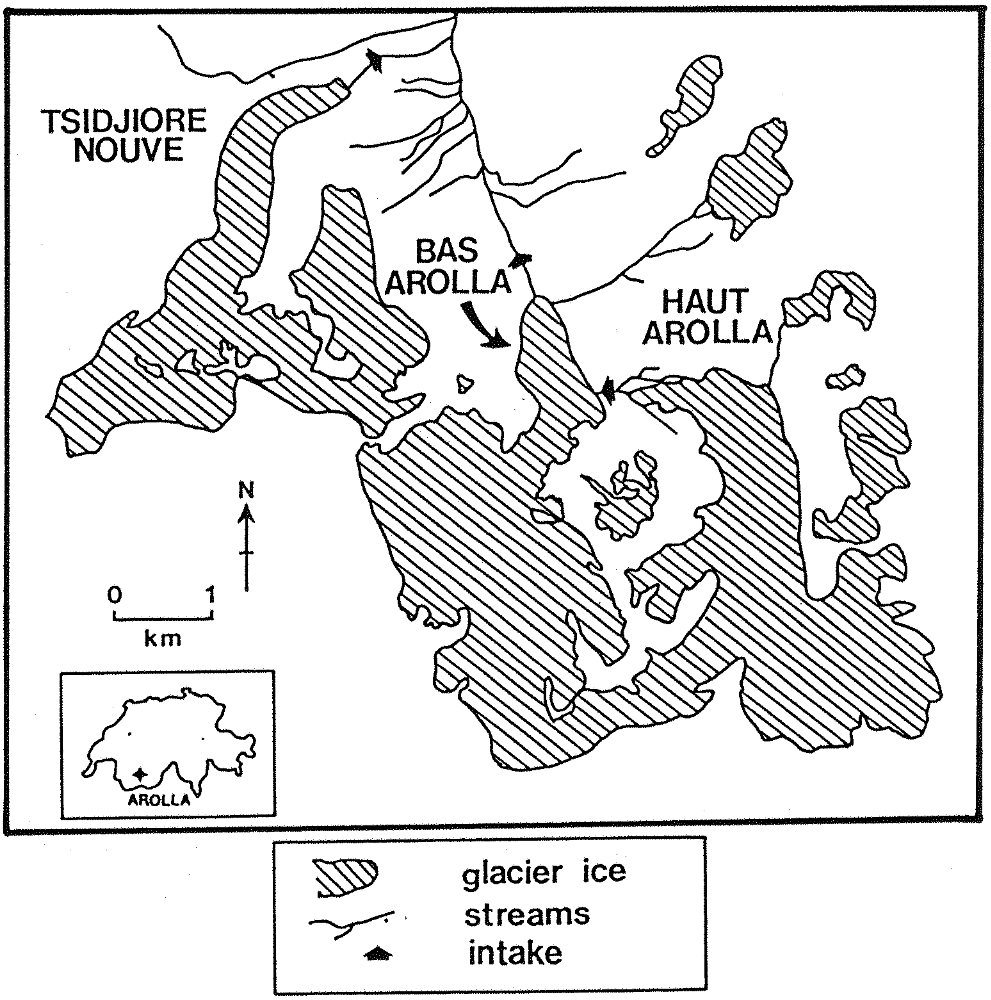


Figure 1.3 Location map of the three study glaciers.

During the Little Ice Age significant glacier advances occurred in the Val d'Hérens between 1550-1850. This was followed, up until 1925, by a period of cumulative spasmodic advances interspersed with retreat phases after which there was general retreat. Recently, Glacier de Tsidjiore Nouve and Bas Glacier d' Arolla have undergone distinct advances. Glacier de Tsidjiore Nouve is particularly susceptible to rapid fluctuations (Whalley, 1973). In contrast the Haut Glacier d'Arolla, in recent times, has shown marked contraction, but since about 1980 it has been advancing slightly. There are no long-term measurements of glacier snout position for the Haut Arolla glacier, so its movements are only estimates based on observations by local people and Grande Dixence employees. The implications of these contrasting glacier dynamics for the water/sediment systems of these glacierised catchments are difficult to infer.

Table 1.1 summarises the characteristics of the three catchments studied in this thesis. Information on glacier catchment details and proglacial characteristics are given along with information about the location of the main monitoring stations during the 1986 and 1987 measurement seasons. Figures 1.4 and 1.5 show the general characteristics of the three proglacial zones. At this point it is useful to note that the three catchments have differences in catchment area, glacierised area (%) and the altitudes of the snouts. However, this is dependent on how the catchments are defined in relation to re-routing of sediment charged waters by the hydro-electric scheme (Figure 1.3). The present Bas Arolla catchment is managed, to the extent that meltwater from Haut Glacier d'Arolla and glaciers Mont Collon, Vui be and Bertol are abstracted and diverted for hydro-power generation and sp do not flow into the lower Bas Arolla glacier system. Contrasts between the present 'managed' catchment and the 'natural' Bas Arolla catchment are shown in Table 1.1. Therefore, only the Tsidjiore Nouve and Haut Glacier

Table 1.1 General characteristics of the Haut Arolla, Bas Arolla and Tsidjiore Nouve study basins.

	UNITS	TSIDJIORE NOUVE	BAS AROLLA	HAUT AROLLA	UNMANAGED **
Catchment area	km -2	4.79	7.56	11.74	25.13
Glacierised area	km -2	3.38	5.31	6.33	13.81
Percentage glacierised	%	70.6	70.2	54.0	55.0
Bedrock geology *	-	Gneiss/ Schist	Gneiss/ Schist	Gneiss/ Schist	
Orientation of accumulation zone	-	N	N	N	-
Orientation of ablation zone		NE	N	NW	-
Glacier length	km	5.0	5.0	4.2	-
Maximum altitude of catchment	m.a.s.l.	3796	3716	3585	-
Maximum altitude of glacier	m.a.s.l	3795	3538	3498	-
Altitude of snout	m.a.s.l	2270	2140	2560	-
Altitude of gauging station	m.a.s.l.	2120	2105	2495	-
Distance from snout to gauging station	m	370	330	950	-
Channel length (From snout to gauging site)	m	330	350	1005	-
Average valley train slope	degrees	13.3	5.0	3.9	-
Average channel slope	degrees	15.6	4.5	3.7	-
Altitude of main monitoring sites 1986	m.a.s.l	2130	2105	2495	-
Distance from snout to main site 1986	m	330	290	950	-
Altitude of main monitoring sites 1987 - upper	m.a.s.l	-	2130	-	-
- lower	m.a.s.l.	-	2107	-	-
Distance from snout to main site 1987 - upper	m	-	100	-	-
- lower	m	-	285	-	-

NOTES:

* Gneiss - Arolla series of Dent Blanche nappe, Schist - Schistes Lustres series

** This is the 'natural' unmanaged catchment which includes: Bas Glacier d'Arolla; Haut Glacier d'Arolla; and glaciers Mont Collon, Vuibe and Bertol.

Figure 1.4 Bas Glacier d'Arolla proglacial zone, July 1987.

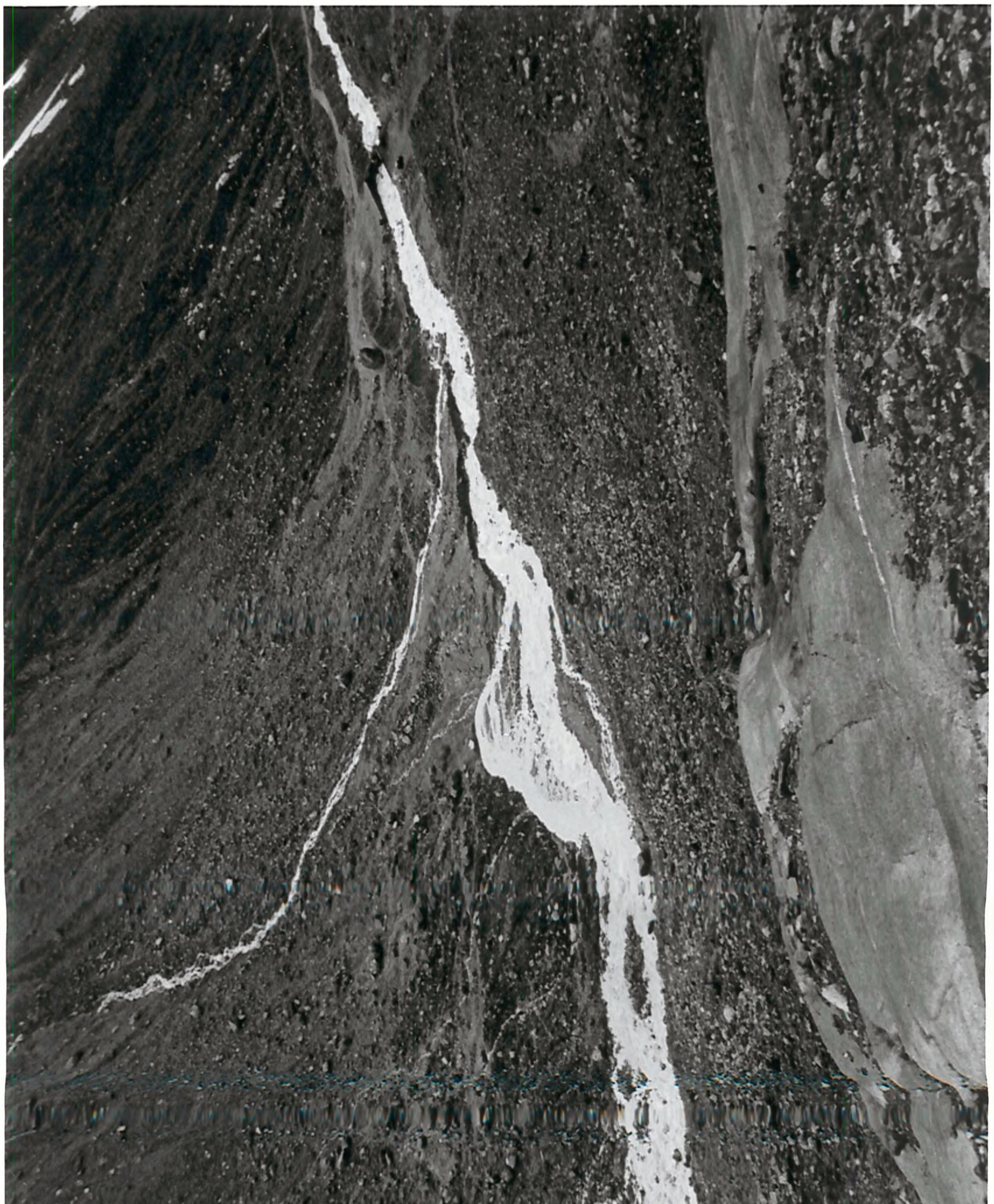
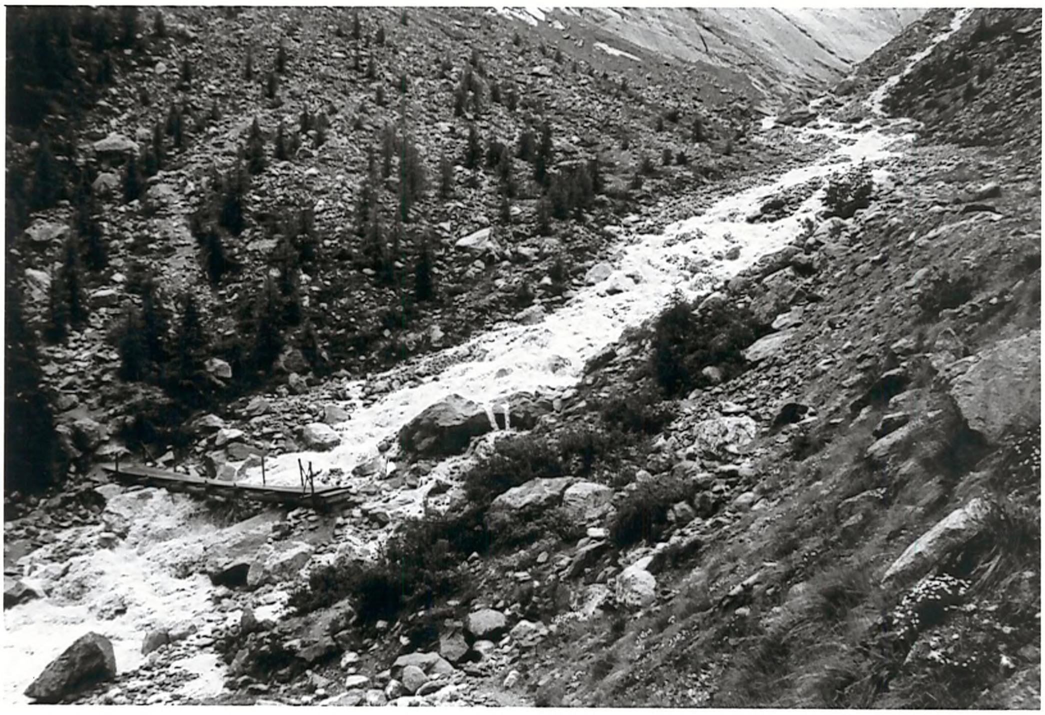


Figure 1.5 A) Upper photograph - Tsidjiore Nouve proglacial zone, July 1986.
B) Lower photograph - Haut Arolla proglacial zone (lower section) and meltwater intake structure, July 1986.



d'Arolla basins can be thought of as tru ly natural, the Bas Glacier d'Arolla basin has a substantially reduced catchment and glacier area generating runoff from the basin. Under these circumstances interpretation of long-term hydrological and sediment data from this glacier should be attempted with caution since interference, by man, over the last 30 years in the glacio-fluvial system cannot easily be evaluated.

Table 1.1 also shows the general characteristics of the three proglacial zones. Of particular note are contrasts in the gradients of both the valley trains and stream channels; and distances between gauging site and snout. In 1986 all three catchments were monitored but in 1987 all effort was focussed on the Bas Arolla proglacial zone (Figure 1.4).

Volumes and patterns of runoff vary between the three basins. Figure 1.6 (based on data supplied by Grande Dixence) shows ablation season hydrographs for Tsidjiore Nouve Bas Arolla and Haut Arolla catchments for 1986 and for the Bas Arolla catchment in 1987. Runoff volumes differ greatly between the basins but relate well with catchment area (Table 1.1). There are also notable yearly variations in the pattern and timing of events throughout the season for the same basin (e.g. Bas Arolla 1986 - 1987). In particular the role of large flood events is important. Three floods occurred in the Bas Arolla catchment during 1986 and 1987 (Figure 1.4), on July 6th 1986, July 15th-18th 1987 and August 24th 1987. Similar events also occurred at the other two sites but it was only at Bas Arolla that the events were documented in detail. The implications of these flow variations for the sediment system will be discussed in later sections (Chapters 6 and 8).

1.2.3 Proglacial fluvial processes

The hydrology and fluvial geomorphology of alpine

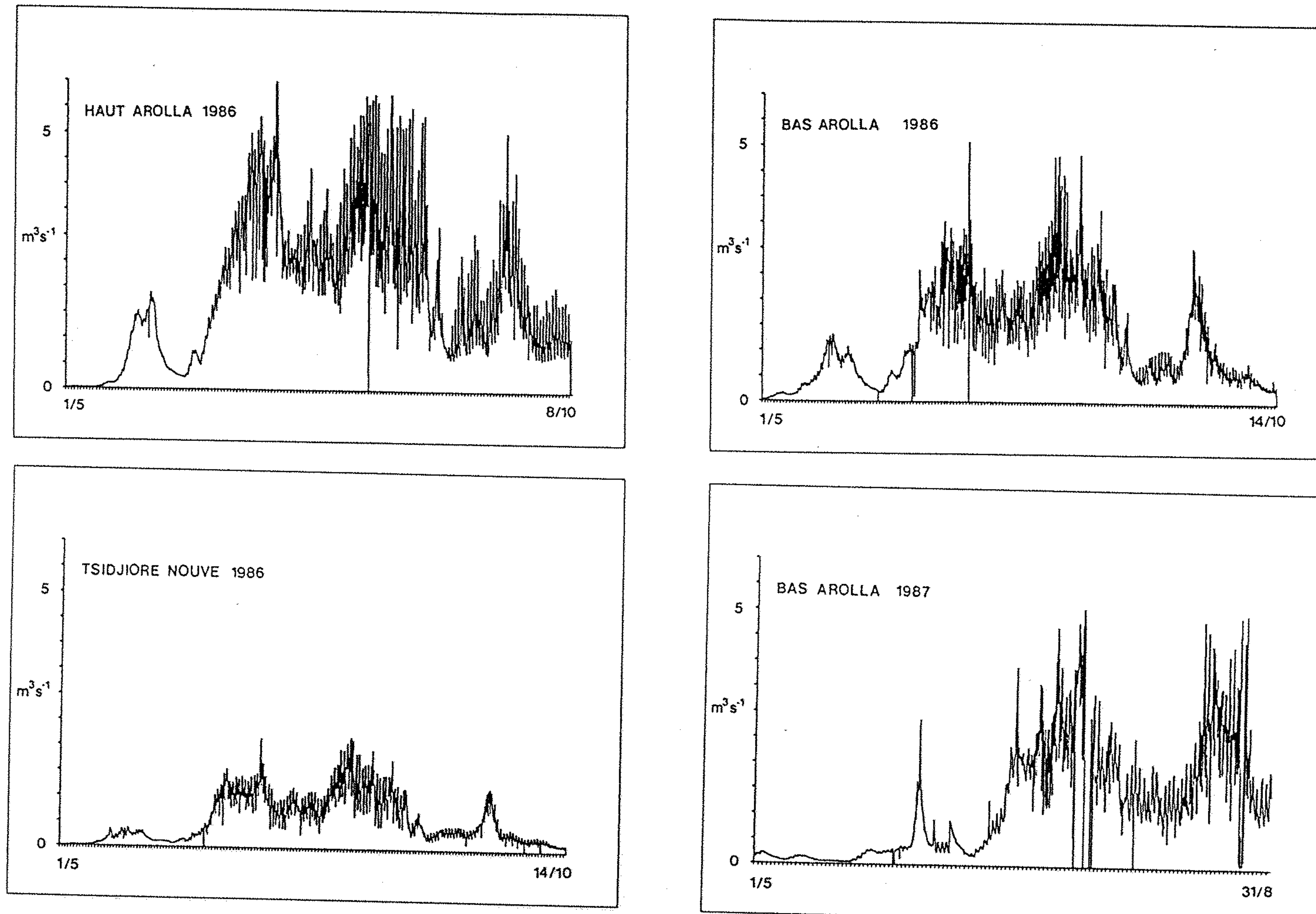


Figure 1.6 Ablation season hydrographs for Bas Arolla, Tsidjiore Nouve and Haut Arolla catchments in 1986 and for Bas Arolla in 1987.

proglacial streams is highly distinctive (Table 1.2). Strictly speaking the proglacial zone is the area proximal to the glacier and therefore includes a suite of channel and slope processes (Figures 1.4 and 1.5). This project concentrates on the channel processes and therefore only considers the area of this zone affected by these processes. The area affected by fluvial processes is usually termed the valley train and in alpine glaciated valleys consists of outwash constricted by the valley wall or side slopes (Fahnestock, 1963). Valley train deposits are of the 'Scott type' (Miall, 1978) and consist of massive, crudely-bedded gravels, crudely-stratified gravels and sand lenses. Channels are steep, single thread and low sinuousity with occasional braiding. They migrate back and forth across the valley train eroding primary landforms and sediments. The multi-storey gravel units are built up by aggradation of longitudinal bars with the lenses of sand representing deposition in abandoned channels or bar-edge sand wedges. These fluvial deposits combine with poorly sorted diamicts and colluvium. Some downstream sorting and particle rounding may also be evident in the general case. However, in the study area considered here the short distance between the glacial snout and hydro-electric water intake means that such sorting and rounding is less likely to be evident.

Flows are highly variable: "meltwater discharges are characterised by violent summer floods superimposed on well-defined seasonal and diurnal discharge cycles ... (Eyles, 1983)". Floods are important in terms of sediment transport and channel change, and Haeberli (1983) has produced a general review of these events for the Swiss Alps, while Beecroft (1983) discusses a specific example of a 'sudden break' outburst flood hypothesized as emanating from a water pocket in Glacier de Tsidjiore Nouve in 1981. Channelised flow is turbulent and water surfaces and velocity profiles are highly irregular. In some cases, step-pool or

Table 1.2 Alpine Proglacial Stream Characteristics

FLOW	Highly variable
PATTERN	Low sinuosity, single thread, occasional braiding
GRADIENT	Steep
CHANNEL BANKS	Non-cohesive, easily eroded, very rarely bedrock controlled
STREAM LOAD	High sediment and solute loads
CHANNEL CHANGE	Rapid channel shifts, sedimentary structures are frequently formed and destroyed, large lateral and vertical changes in the thalweg
CHANNEL VEGETATION	Poorly established
SEDIMENT SOURCES	Abundant, poorly sorted diamicts and colluvium, large range in bed materials form sand to boulders (1-2 m)
SEDIMENTOLOGY	Proximal gravels, massive crudely bedded gravels and cudely stratified gravel sheets (Scott type - Miall, 1978), longitudinal bars and sediment clusters are common, some downstream sorting and particle rounding
WATER TEMPERATURE	Low (0 - 2 °C) - high viscosity and increased bedload entrainment potential

'staircase' type channels with tumbling flow develop (Whittaker, 1987). Channels respond continuously to rapidly varying water and sediment flows (Fenn and Gurnell, 1987) and as such appear to be self-formed (Richards, 1982). However, this is not strictly true since large boulders are not moved directly by the flow and some remain immobile. For this reason and because of the combination of random influences (e.g. bank heterogeneity) and non-stationarity in external factors, equilibrium forms tend to be approached but not actually achieved. Nevertheless, channel change is rapid due to the easily eroded non-cohesive, sparsely vegetated channel banks. The abundance of sediment and the highly erosive nature of the glacio-fluvial system contribute to high sediment loads, which sometimes produce aggradation in the main stream channel. These processes show marked similarities with those of arid environments "In general, fluvial processes in the glacial environment are similar to those in arid environments characterised by 'flashy' discharges, sparse vegetation cover and substantial fluxes of debris" (Eyles, 1983).

1.2.4 The Hydro-electric power scheme

All three glacial basins studied in this investigation fall within the catchment of the Grande Dixence Hydro-electric scheme (Figure 1.7). This is Switzerland's largest hydro-electric development with a catchment area of 357 square kilometres, 50 % (180 square kilometres) of which is covered by snow and ice. This immense scheme, which is "certainly the greatest hydro-electric exploitation of any region of Europe" (Elgi, 1978), was started in 1950 and completed in 1965. The construction phase of the Grande Dixence scheme included: 77 separate water intakes; 4 major pumping stations; 100 kilometres of subterranean gallery; the highest (284 m) concrete gravity dam in Europe, and a storage lake with a capacity of 400 million m^3; and two power generating stations at Fionnay and Nendaz

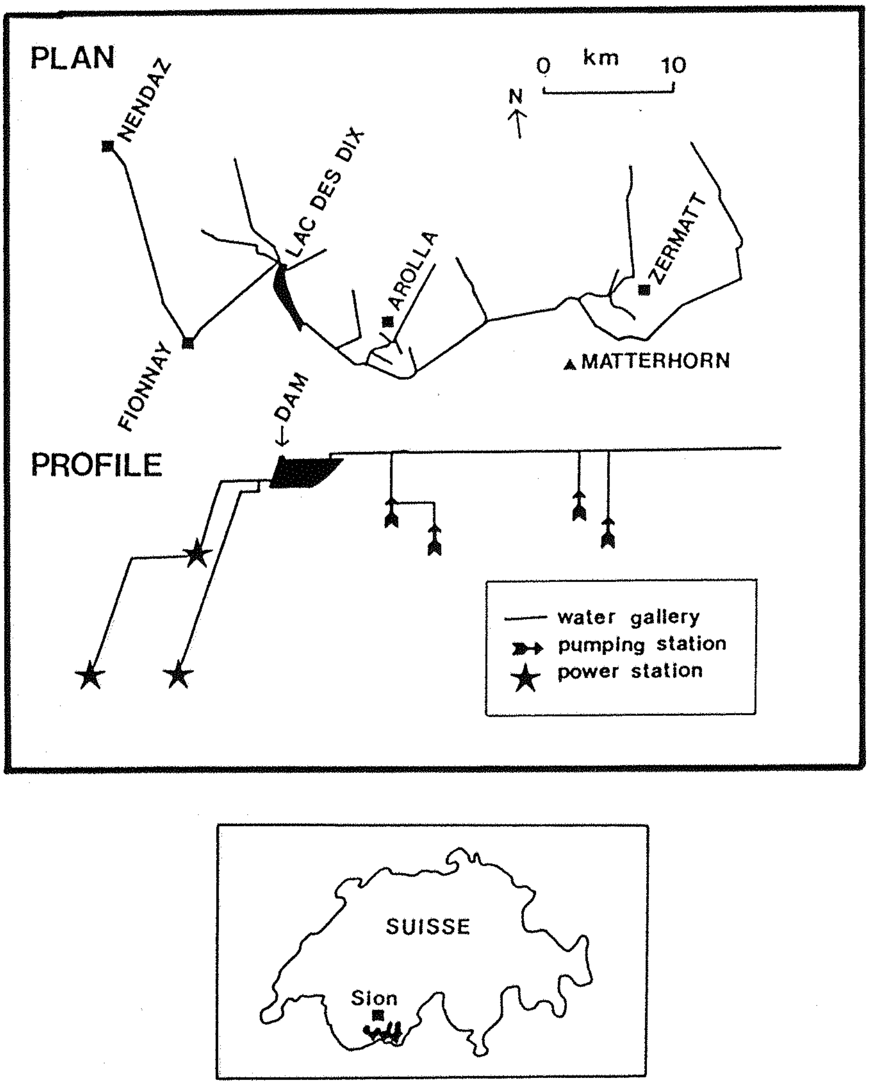


Figure 1.7 The Grande Dixence hydroelectric scheme.

(Figure 1.7).

Within the context of the study area, water from the Bas Glacier d'Arolla joins water from the Ferpècle valleys in a 'gallery flow' compensation chamber at an altitude of approximately 2074 m in the valley side opposite the Arolla pumping station. This water is then fed to the pumping station where it meets with Tsidjiore Nouve glacial meltwaters routed by a gallery to this same point. Water is pumped from the valley floor at 2009 m up 312 m to the principal high-level feed gallery which drains into the Lac des Dix, the main storage reservoir (2160 m). Haut Glacier d'Arolla water directly enters this high-level gallery at the head of the Arolla Valley. The maximum flow rate in this gallery, at Cheillon, is 82 $m^3 s^{-1}$.

The function of the storage reservoir is to collect summer meltwater for use in the winter months. The storage scheme supplements other power sources in meeting daily and seasonal peak load demands (Bezinge and Kasser, 1981). Most of the stored water flows through large pressure tunnels, first to the generating station at Fionnay (1486m) and then in a second stage to the power station at Nendaz (478m). Both power stations have six horizontal sets each consisting of a generator and 2 Pelton turbines. The Nendaz turbines generate 80MVA each whilst the smaller Fionnay turbines produce 60MVA each. The combined gross annual output is 1680 GWhr, 86% of which is generated in winter.

Measurement of discharge from glaciers within this scheme is important for modelling and forecasting in order to design optimal pumping programmes. The Grande Dixence scheme has been studied by Lang and Dayer (1985) with a view to producing such a management program. An important implication of this is that discharge records, based on hourly mean values, are available for the three glacier basins studied (Figure 1.6).

1.2.5 Brief synopsis of previous work

Previous research, relevant to the present study, which has been based in the Arolla valley is summarised in Table 1.3. Three points are clearly evident:

1) Interest in the area greatly increased after the early 1970's which parallels the general increase in glacial hydrological research at this time (Fenn, 1983). British Universities, particularly the University of Southampton, and Universite Libre de Bruxelles have dominated the work on current processes in this particular valley.

2) Tsidjiore Nouve has proved to be the most popular research site which probably reflects the factors of easy accessibility and minimal disturbance by man. Much of the major work on the Tsidjiore Nouve catchment is summarised in Gurnell and Clark (1987).

3) Research has studied the components of the proglacial fluvial regime but no attempt has been made to construct a proglacial fluvial sediment budget. A particular omission is the lack of detailed study of channel processes and channel change. Notwithstanding this apparent drawback, existing research has provided an invaluable background to the present study and is referred to in detail in the subsequent chapters.

Year	Author	Glaciology	Glacial Hydrology	Glacial History Moraines	Proglacial
TSIDJIORE NOUVE					
1973	Whalley			X	
1974	Small and Clark			X	
1976	Bezinge			X	
1976	Rothlisberger			X	
1978	Lemmens		X s/ec		X s/ec
1978	Lemmens and Roger		X s/ec		X s/ec
1978	Souchez, Lemmens, Lorrain and Tison	X			
1978	Souchez and Lorrain	X			
1978a	Tison	X			
1978b	Tison	X			
1979	Rothlisberger and Schneebeli			X	
1979	Small, Clark and Cawse			X	
1981	Beecroft				X ss
1981	Small and Gomez			X	
1981	Souchez and Tison	X			
1982	Gurnell				X ss
1982a	Small			X	
1982b	Small			X	
1983	Beecroft				X y
1983	Small			X	
1983	Gomez and Small			X	
1983	Gurnell				X q
1984	Gurnell and Fenn				X s/ec
1984	Gurnell and Fenn				X ss
1984	Gurnell and Fenn				X ss
1984	Small, Beecroft and Stirling			X	
1985	Fenn, Gurnell and Beecroft				X ss
1987	Fenn		X s/ec		X s/ec
1987	Fenn and Gurnell				X c
1987	Gurnell				X ss
1987	Gurnell				X y
1987	Small				X y
				X	
BAS AROLLA					
1911	Mercanton			X	
1949	Knecht and Susstrunk	X			
1951	Clark and Lewis	X			
1951	Haefeli	X			
1967	Haynes	X			
1973	Ferguson		X c		
1976	Rothlisberger			X	
1976	Whalley			X	
1979	Small, Clark and Cawse			X	
1981	Park		X c		
1982b	Small			X	
1983	Gurnell				X q
1987	Clark, Gurnell and Hancock	X rs		(X rs)	X rs
1987	Fenn		X s/ec		X s/ec
1987	Fenn and Gurnell				X c
1987	Small			X	X y
HAUT AROLLA					
1979	Small, Clark and Cawse			X	
1982b	Small			X	
1985	Gomez and Small			X	
1987	Fenn				X s/ec
1987	Small			X	X y

(NOTE: X = Topic of interest. Subscripts indicate type of study - s/ec = solutes/electrical conductivity, ss = suspended sediment, y = sediment yield, q = discharge, c = channel processes, rs = remote sensing)

Table 3.1 Brief Synopsis of previous work in the Arolla area

1.3 Thesis Structure

This thesis is divided into ten chapters, based on the sediment budget/ approach. Each deals with a particular aspect of the general problem outlined in the introduction. This division into chapters is somewhat arbitrary since the contents of each section can only fully be understood in the context of the other sections. In order to avoid a lengthy literature review, references are introduced in the form of 'mini-review' sections at appropriate points in the discussion. This approach has the advantage of placing the most relevant references close to the appropriate subject matter.

Following the introduction and description of the study area (Chapter 1), the general approach to measuring and monitoring fluvial sediment transfer in the proglacial zone is outlined (Chapter 2). This leads to Chapter 3 which considers the components of sediment yield from the study basins (bedload, suspended load and solute outputs). Next inputs and outputs of sediment from the proglacial zone are considered in Chapter 4. Chapter 5 discusses tributary streams as a source of sediment and examines their storage and transport characteristics. Valley train sediment sources are then investigated with the discussion divided between channel change (Chapter 6) and valley train bluff erosion (Chapter 7). Of particular interest in chapter 6 is the role of flood events in transporting sediment. Following on from this, Chapter 8 examines the importance of large flood events and the origins of small sediment pulses in order to illustrate additional features of the proglacial sediment system and their relationship with glaciological and hydrological controls outside the proglacial zone. Based on the results presented in chapters 3, 4, 5, 6 and 7 a tentative sediment budget for the fluvial component of the proglacial sediment system is constructed (chapter 9). Finally, chapter 10 summarises the major conclusions and recommendations for improvements in measuring techniques and highlights areas for further research.

Chapter 2.

MEASURING AND MONITORING FLUVIAL SEDIMENT TRANSFER IN PROGLACIAL ZONES - FIELD AND LABORATORY TECHNIQUES

2.1 Introduction

A primary component in constructing a sediment budget is the quantification of contributions from sediment transport processes (Lehre, 1982; Duijsings, 1985). This chapter attempts to provide a structured framework for measuring proglacial fluvial sediment transport processes and is therefore concerned with this requirement. The main system of interest is the fluvial sediment system: however, where glacial and slope systems have a direct effect, they are considered.

This chapter provides a summary of the main techniques and methods used in data collection, which is the key to sediment budget specification. Data collection is defined as the measurement and collection of field attributes or samples, and the processing of this information into a form suitable for higher analysis or graphical presentation. Therefore all elements of the field monitoring programme, calibration procedures, abstraction of data from charts, laboratory analyses and data processing are included in this chapter. Most of the techniques described could be executed under field conditions or in places were only primitive laboratory conditions are available. Goltermann et al (1978) associate techniques of this kind with 'level 1 methods' in terms of their sophistication; that is, they are implemented at the most basic of levels. This is an important pre-requisite given the difficulties of storing and transporting samples from the field to the laboratory. Emphasis is placed on providing a critical assessment of the techniques used, especially where they deviate from standard methods. Data are presented which aid in the interpretation of the field records, and provide an insight into the limitations of individual techniques. These details are necessary because they provide estimates of the accuracy of techniques, an understanding of the limitations of the data (important in subsequent analyses) and a review of the techniques most appropriate for work in the proglacial fluvial environment.

2.2 Overview of Measurement Programme

Although any field measurement programme is a compromise between the ideal and the practical, this is especially true in monitoring proglacial fluvial sediment transfer. It is essential that the field programme is designed to minimise the constraints imposed by restricted resources and a short-term measurement programme, whilst still providing the necessary resolution to characterise the relevant processes.

The measurement programme was designed to provide a short-interval, long-duration data base with a hierarchical, spatially and temporally 'nested' structure. The structure is based partially on the recommendations of Fenn (1983) who concluded that measurements of proglacial time series should be taken every hour or on a continuous basis to characterise their marked temporal variability, and that some account should also be taken of the spatial variability in process which can also be considerable. It is this heterogeneity within the proglacial sediment system which makes representative sampling difficult. Care needs to be taken in designing a measurement scheme which adequately accounts for these variations. In the present context an inadequate monitoring programme could produce excessive generalisations in process estimates which would adversely affect the estimated sediment budget, particularly in the case of its smaller components. A successful field monitoring programme is dependent not only upon good design but also on reliable performance of the equipment used. Problems with instrumentation caused some difficulties in the present programme, and these will be discussed where relevant.

The field programme was divided between two ablation seasons, 1986 and 1987. In the 1986 field season three proglacial zones were studied and in 1987 research was concentrated within th proglacial zone of the Bas Glacier d'Arolla.

The three-site study of 1986 was designed to establish the degree of inter-basin variability in sediment yield, to establish reliable monitoring techniques, and to provide a sediment yield context for concentrating on the sediment budget of the Bas Arolla proglacial zone in 1987. The 1987 programme was designed to determine sediment sources and their inputs into the sediment system of the Bas Glacier d'Arolla proglacial zone so that the sediment budget of the proglacial zone could be evaluated. An important aspect of the 1987 programme was the gauging of inputs and outputs of suspended sediment through the proglacial zone, which required two main monitoring stations to be established at proximal and distal ends of the channel reach (Figure 2.1).

The single monitoring stations in 1986 were established inside the Grande Dixence underground water intakes in the Glacier de Tsidjiore Nouve and Bas Glacier d'Arolla basins and on the proglacial stream water intake at the Haut Glacier d'Arolla basin. The instrumentation was essentially the same at each of the three monitoring sites (see below and Figure 2.2) except that at the Haut Glacier d'Arolla site, instruments were powered by a Rutland wind generator which 'trickle charged' a 12 volt battery, whereas mains electricity was available inside the water intakes at the other two sites. Unfortunately

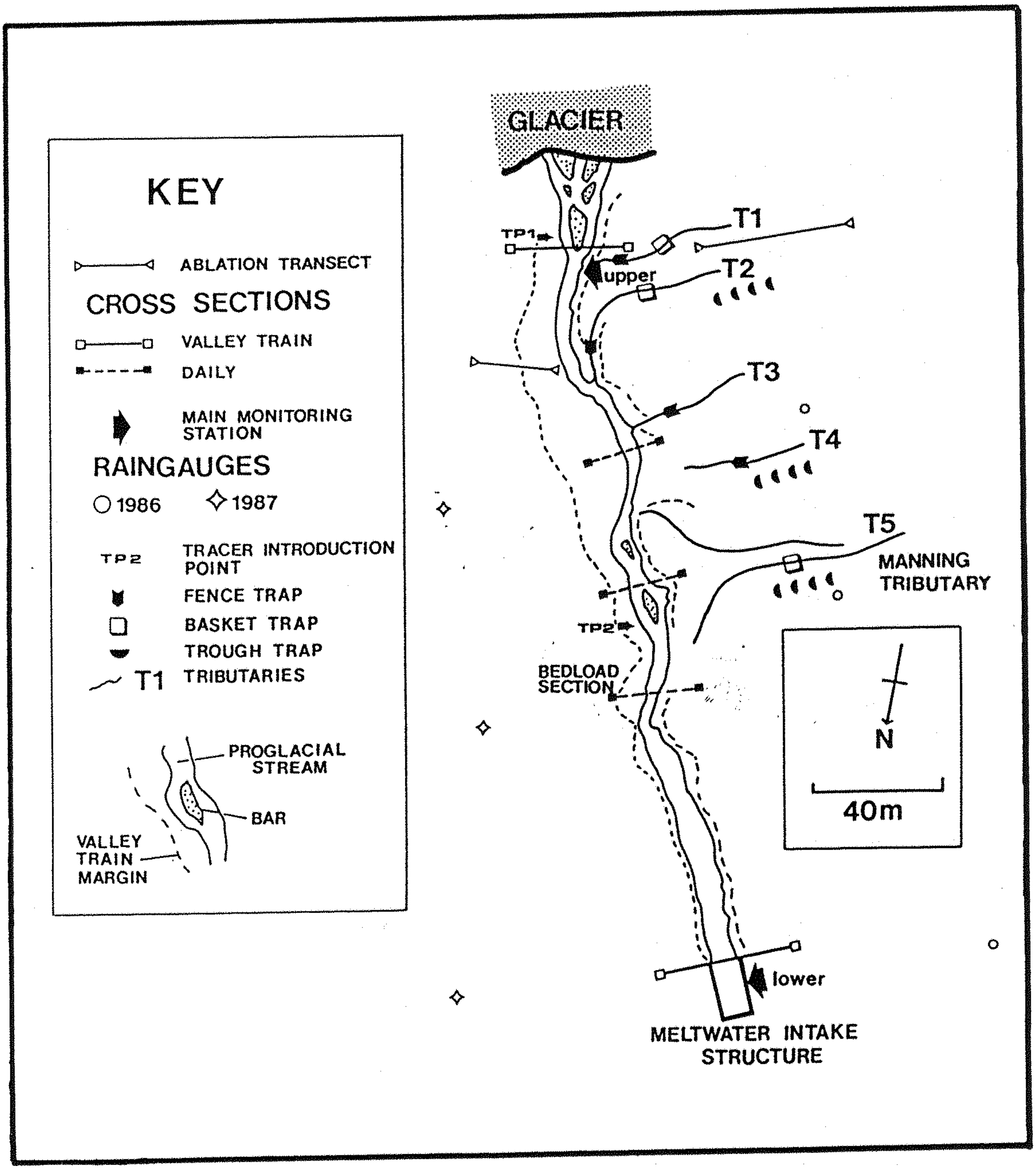


Figure 2.1 Main monitoring and measurement sites Bas Arolla proglacial zone 1986 and 1987. Only the upper and lower valley train cross-sections are shown.

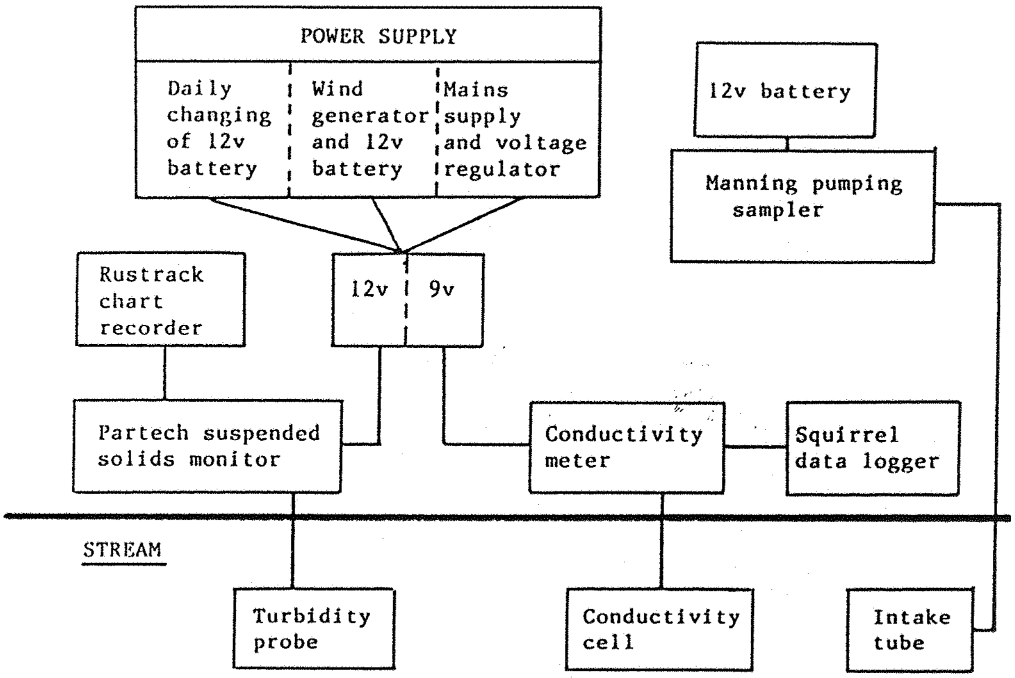


Figure 2.2 Main site instrumentation and power supply.

the wind power was not sufficient at the Haut Arolla site and a battery changing/charging routine had to be established. In 1987 two monitoring stations were established on the Bas Glacier d'Arolla proglacial stream (Figure 2.1). The two stations were not linked to mains electricity and were powered from batteries which were charged daily. This arrangement was maintained until the 30th July after which the downstream station was moved inside the water intake and connected to mains electricity in the same location as in 1986 and the upstream station was removed.

It can be seen in Figure 2.3 that the field monitoring period extended from mid-May to early-September so that field observations were confined within a 'window' of time, albeit the most important period of sediment transport activity. Within this period measurements were taken at five temporal scales: 'continuous', hourly, daily, weekly and 'event based' (flexible timescale). Continuous data were not truly continuous since the Rustrack recorder used in monitoring makes a chart mark every two seconds. The sediment budget is based on measurements made from late May - late July, 1987 and is therefore a partial season budget.

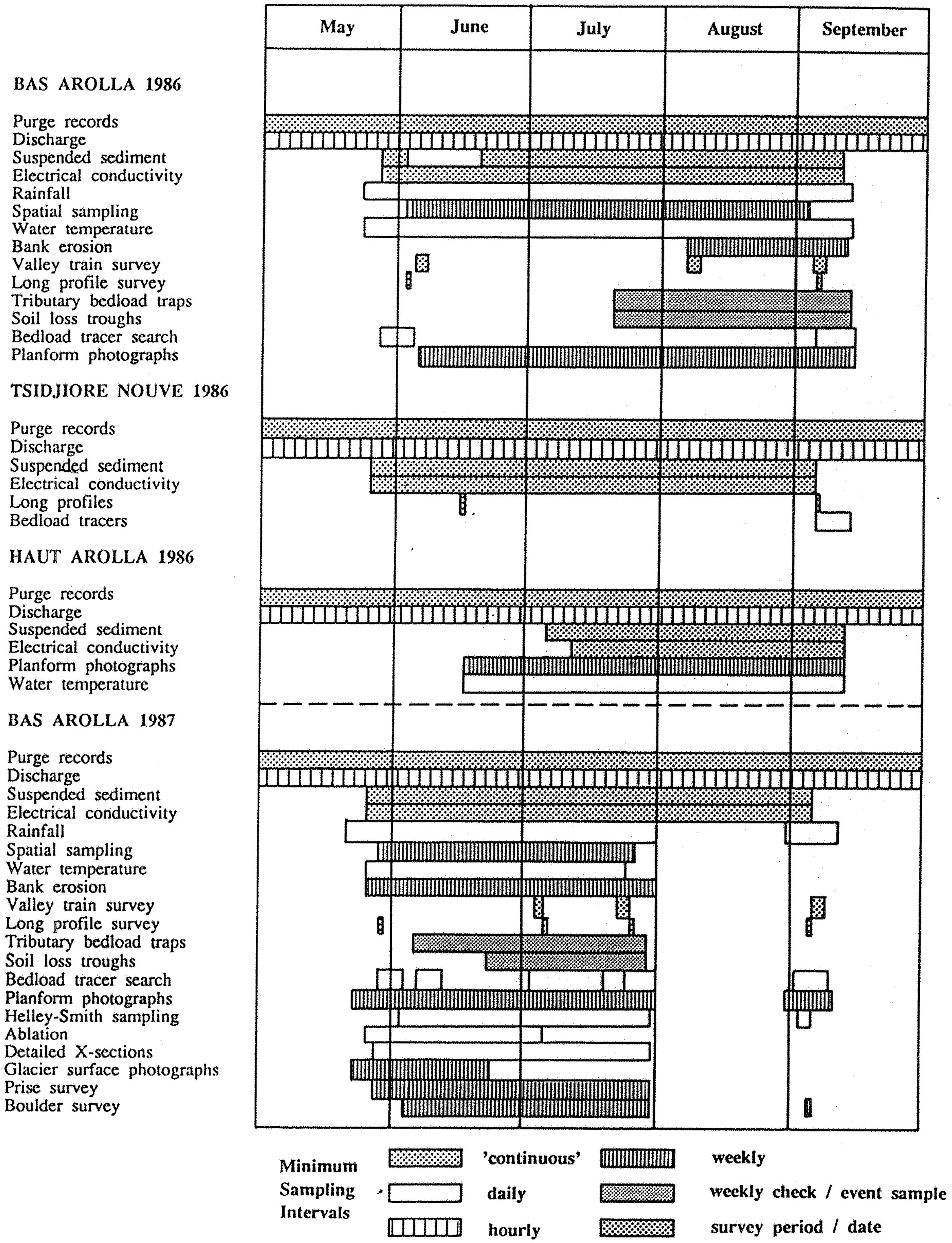


Figure 2.3 Summary diagram of fieldwork sampling strategies, survey dates and periods of observations.

2.3 Field methods

2.3.1 Main Monitoring Stations and sediment traps

Main monitoring stations were instrumented to record turbidity and electrical conductivity. As mentioned previously individual station characteristics differed little apart from the power source (mains, battery-charging from mains or wind generated). Turbidity was recorded as a continuous trace on a Rustrack chart recorder and electrical conductivity was stored at 20 minute intervals on a Squirrel data logger. Electrical conductivity was not recorded continuously since tests of logging conductivity at intervals of 1 second to 1 hour added little information and a 1 hour logging interval was economic in using storage capacity. In addition to these 'fixed' site instruments, a Manning pumping sampler (Figure 2.2) was used automatically to collect meltwater samples for calibration purposes (Section 2.4.1). Siting of the turbidity and electrical conductivity probes and the sampler intake hose (Figure 2.2) required care to prevent silting, burial or artificial turbulence created by the automatic sampling mechanism. Particular care needs to be taken in siting turbidity probes, and guidelines for doing this are given in Appendix 1., Section 3.

Three basic criteria were used in locating the main stream monitoring sites on the Bas Arolla proglacial stream in 1987: (1) The channel reach had to be stable and single thread: (2) The location had to be 'natural'; that is, unaffected directly by activities associated with the hydro-electric scheme, and (3) it had to be close to a Grande Dixence installation where discharge measured at the water intake could be closely related to stream channel discharge and where access to battery charging facilities was readily available.

The combination of the data generated by the portable monitoring equipment (Figure 2.3) and the operation of sediment traps Grande Dixence water intake/sediment trap of the structures provided an ideal opportunity accurately to gauge the output of sediment in the proglacial stream. Figure 2.4 shows the design of the meltwater intake structure used on the proglacial streams of Tsidjiore Nouve and Bas Arolla. Streamflow is collected in the water intake (gravel trap) structure which traps the bedload gravels and a proportion of the suspended load. Water is then diverted through a sand trap, where more of the suspended load (coarse - medium grain sizes) are deposited, and over a weir into the hydroelectric tunnel system which feeds a large storage reservoir. A stage recorder continuously monitors discharge behind the weir at the outlet from the sand trap (site D). A range of variables was monitored at different locations on the water intake in order to monitor sediment transport both directly and indirectly. Indirect methods were aimed at calibrating the amount and calibre of sediments collected in the sediment traps with a view to estimating sediment yield and possibly extrapolating yield estimates over a longer time period than the two field monitoring seasons. A summary of the measurements made in and around the intake structures (sites A, B, C and D) is given in Table 2.1. The two sediment traps are purged of sediment periodically, usually in response to automatic sensing of sediment accumulation although some manual purging occasionally occurs. The response of the stage recorder (site D) to purging is a fall in the water level to one of two levels depending on which trap is purged. This allows the type and timing of purges to be identified, which can be combined with estimates of trap volume and sediment packing density to estimate sediment yield from the traps (Chapter 3). The main disadvantage of this system is that during flood peaks, purge gates remain open, diverting water away from the instruments located at site D. However, during these periods, and probably

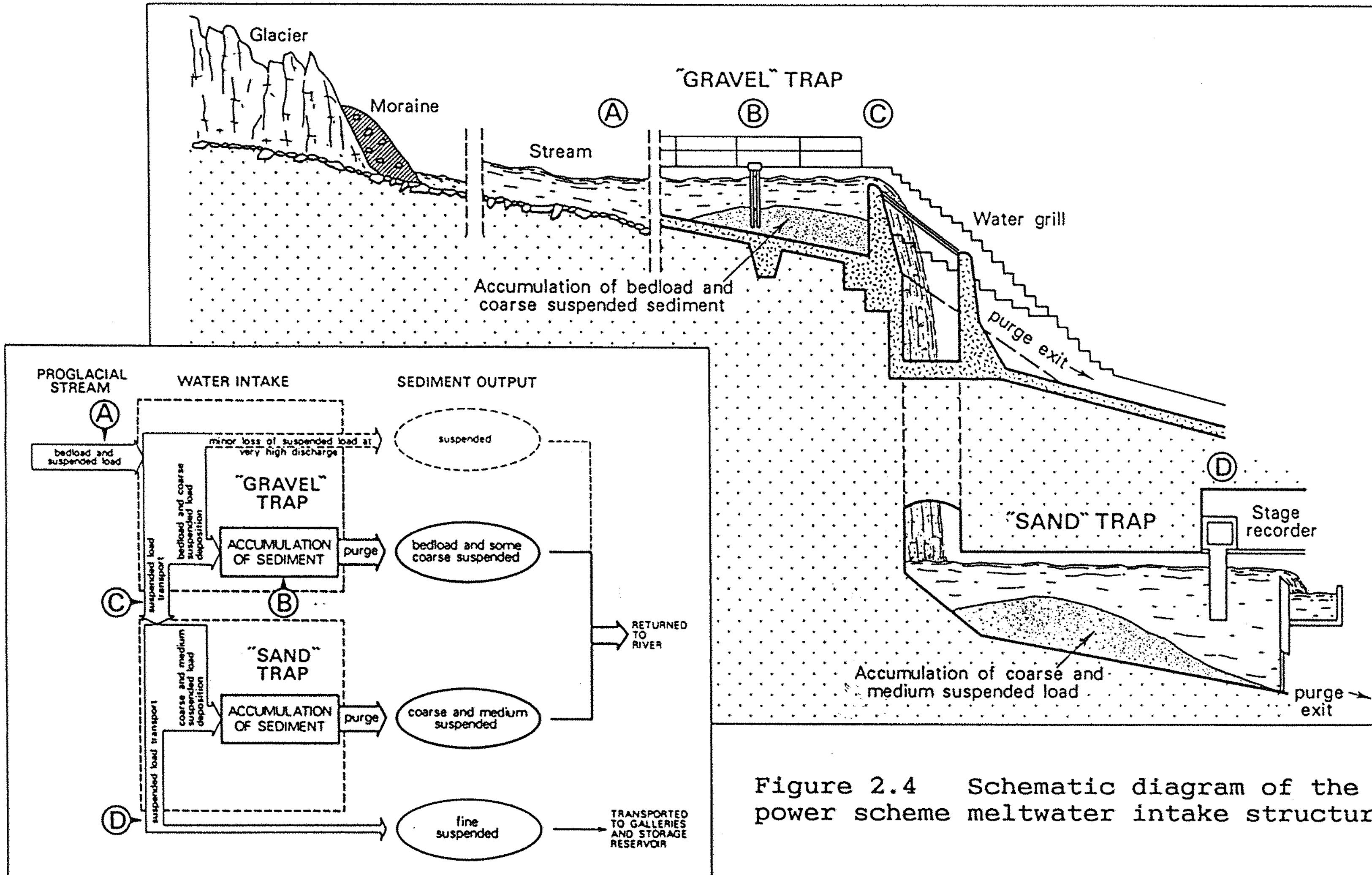


Figure 2.4 Schematic diagram of the hydro-electric power scheme meltwater intake structure.

Table 2.1 Summary of measurements made at the Bas Arolla meltwater intake structure 1986 and 1987.

MEASURED PROPERTY		SITES A	B	C	D
SEDIMENT	GRAIN-SIZE	*	*	*	*
	PACKING DENSITY		*		
BEDLOAD	HELLEY-SMITH	*			
	ACCUMULATION SURVEY		*		
SUSPENDED	TURBIDITY	*		*	*
	BOTTLE SAMPLING	*	*	*	*
SOLUTES	ELECTRICAL CONDUCTIVITY	*		*	*
	BOTTLE SAMPLING	*		*	*
DISCHARGE	VELOCITY/AREA	*			
	SALT GAUGING	*			
	STAGE RECORDER				*
TEMPERATURE	THERMOMETER	*		*	*
	THERMISTOR (continuous)				*
POWER	BATTERY	*		*	
	MAINS				*

SITES: A - Stream channel
B - Intake/gravel trap
C - End of intake/prise top
D - End of sand trap

well before the peak flood flow, conventional stream sites would be rendered non-operational by the high flows. This artificially imposed ceiling to the discharge records also acts as a safegaurd, since field equipment housed in the intake structure will not be lost in flood events.

Suspended sediment concentration was monitored using a continuously recording turbidity meter at site D, where there is mains electricity and where the turbidity record is unaffected by varying light and turbulence conditions. Sampling of the same water parcel (flow time estimated from salt tracing) as it passes sites A, B and D allowed calibration curves to be constructed which permitted estimation of suspended sediment concentration variations at these three sites and the amount of suspended sediment deposited in the sediment traps.

Surveys of sediment accumulation in the gravel trap together with measurements of sediment packing density, based on the accumulation of sediment in containers suspended in the trap, can be used to estimate the amount of sediment removed when the trap is automatically purged. A technique for estimating the volume of sediment deposited in the gravel trap was devised (Figure 2.5) whereby a 4 kg weight, suspended from two ropes one on either side of the trap was lowered to the bed of the trap. Given the width of the trap and the length of the two ropes from either side of the trap, when the weight is resting on the bed, the height of the sediment surface and distance from the side of the trap can be calculated. This procedure can be repeated at a number of points across the trap until enough points have been determined to construct a cross-profile (Figure 2.5). Since the gravel trap also trapped the coarse fraction of the suspended load, the suspended sediment calibration curves for sites A and C were used to estimate deposition of this load in the

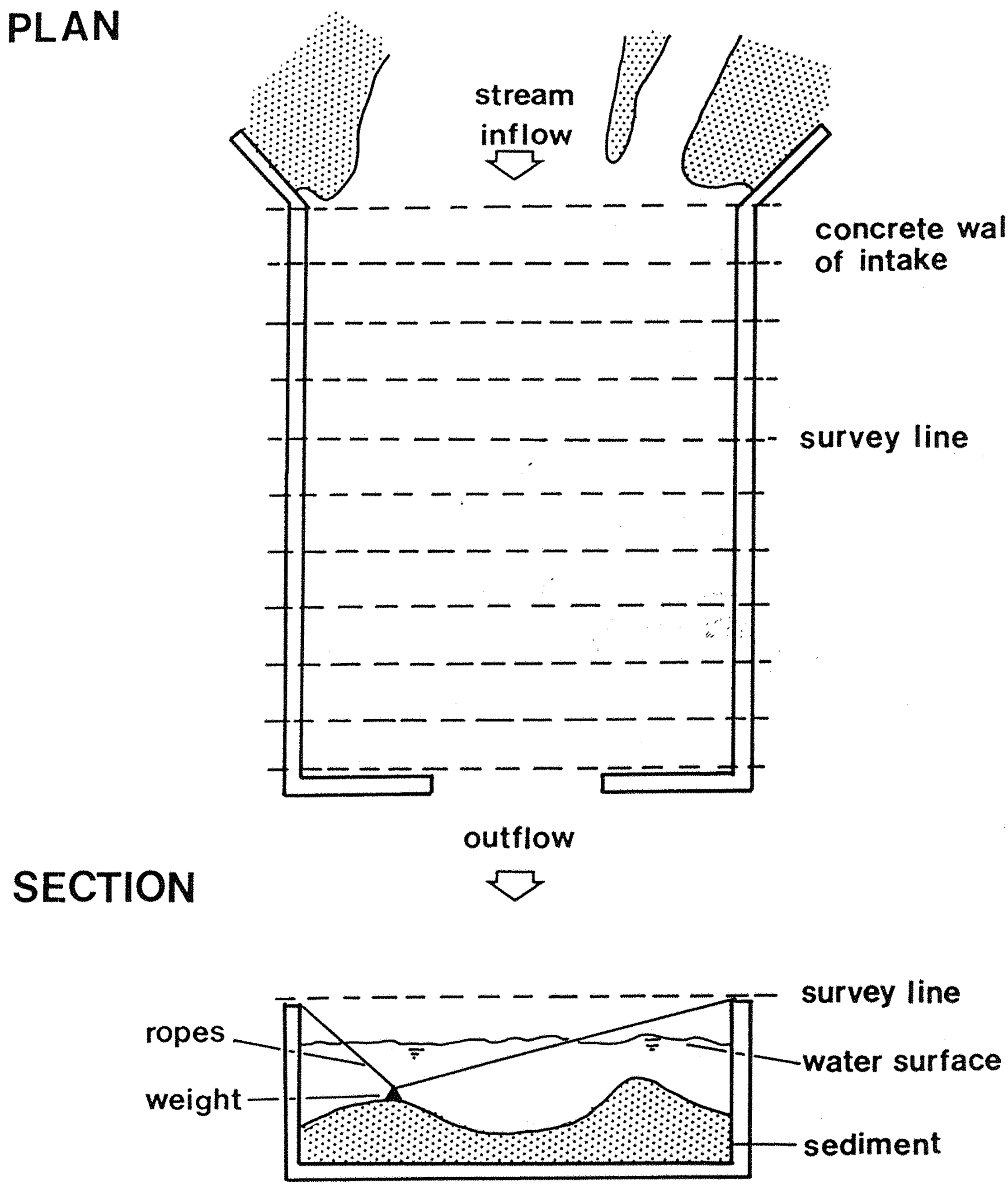


Figure 2.5 Gravel trap survey procedure using a weight suspended from two ropes held at the trap sides (see text for full details).

trap. Subtraction of this load from the total accumulated in the gravel trap gives a corrected estimate of the bedload.

At the Haut Glacier d'Arolla site such measurements are not possible because only a single sediment trap, for both gravel and sand, is installed. Access to this trap is extremely difficult and so estimates of sediment discharge, based only upon approximate values for the trap dimensions and an assumed packing density were made.

Further details of the method of monitoring variables at both the main monitoring stations and at other sites will now be discussed.

2.3.2 Discharge

The majority of discharge measurements were derived from records supplied by Grande Dixence S.A. who measure discharge as water is diverted through their meltwater intake structures. Figure 2.4 shows the water intake structure installed on the proglacial streams at the Bas Glacier d'Arolla and Tsidjiore Nouve and indicates the point at which discharge is gauged. Flow is gauged by means of an established relationship between stage (monitored using a float and counter-weight stage recorder mounted over a stilling well) and discharge over a rigid concrete weir. Discharge was also gauged at various sites on the proglacial stream and its tributaries, using both the velocity-area method (Gardiner and Dackombe, 1983) and 'slug injection' dilution gauging (Østrem et al., 1964; Newson and Harrison, 1978). Velocity was determined using three methods: current metering (Braystoke); salt tracers; and floats. Detailed velocity measurements were also taken at the times when bedload transport was sampled. In these cases three determinations, of 30 second duration, were made at 0.6 depth in each vertical. Discharge determined from the Grande Dixence records was considered to be superior to the data determined using 'in channel methods' where the fast, turbulent flow of the stream and changing cross-section decreased the measurement accuracy.

2.3.3 Precipitation and ablation of snow

The only climatic variables monitored in this study were precipitation and snow ablation. More comprehensive climate data were available from a meteorological station at Bricola at 2415 m in an adjacent valley 8 km away. Three raingauges were used in 1986 and four in 1987. All were sited in the proglacial zone of Bas Glacier d'Arolla and were of the same design (15 cm diameter, 0.5 l collecting bottle). Rainfall was calculated in mm as a 3 gauge mean except when the gauges had been tampered with (a frequent occurrence). Gauges were visited daily. Snow ablation was measured in front of the Bas Glacier d'Arolla along two transects using bamboo poles as markers. One transect was located on the western slope of the proglacial zone 40 m down from the snout and the second on the eastern slope 50 m from the snout (Figure 2.1). Seven poles spaced 4-5 m apart were used in each transect. The sites were visited at 11.00 am daily and the distance from the top of the pole to the snow surface was measured to determine the amount of melt over the previous day.

2.3.4 Electrical conductivity

Electrical conductivity was monitored using a variety of conductivity meters (WPA CM25, WPA CM35 and WPA CM135) and was recorded either continuously onto a Rustrack chart recorder, or logged every hour or fraction of the hour (dependent on the capacity of the logger) on a Grant Squirrel data logger. Conductivity cells were mounted on Dexion and submerged in the main flow of the channel. Occasionally the cells became blocked with sediment but generally they performed well. Cross calibrations between conductivity meters at the upper and lower 1987 main monitoring sites were established using a portable conductivity meter. Results showed a near-perfect linear relationship between the two sites. Electrical conductivity was also sampled manually during 1987 at 14.00 hours over a spatial network of sites, on a weekly or more frequent timebase dependent on local conditions, using the same downstream station sequence during every survey. At the time of every conductivity reading or station visit, water temperature was measured using a mercury glass thermometer submerged in the water for 30 seconds. It is usual to convert electrical conductivity measurements in non-glacial streams to a reference temperature of 25 ^{o}C (Goltermann et al., 1978). However, Østrem et al. (1964) and Collins (1977) have shown that simple temperature correction factors are unreliable for glacial meltwaters. As a result of the restricted range of observed water temperatures (0 to 2 ^{o}C) uncorrected conductivity data are used in this study.

500 ml water samples were taken to determine the concentration of material in solution in relation to the electrical conductivity of the water. These samples were initially filtered through Whatman 40 filter papers then through Whatman GFA filters and the filtrate was stored in glass bottles for determination of total dissolved solids (Section 2.4.1).

2.3.5 Suspended sediment concentration

Determination of suspended sediment concentration has been a major preoccupation of this work. It is acknowledged that the appropriate techniques will vary depending on local conditions (particularly suspended sediment concentration), so that the approach adopted in this study may not be transferable to sites with markedly lower suspended sediment transport. The procedure for determining a continuous record of suspended sediment concentration relied on the calibration of the turbidity meter records. The background to this method and details of calibration are given in Appendix 1. In essence, the suspended sediment concentration was monitored using a continuously recording turbidity meter linked to a mains electricity supply or subject to daily battery changes. The turbidity record was translated into estimates of suspended sediment concentration using a calibration curve based upon manual sampling and filtration of water samples from the proglacial stream. Water samples were collected using a USDH-48 sampler and were filtered through pre-weighed Whatman 40 filter papers. The suspended sediment concentration was estimated from the weight of sediment trapped on the filter papers and was related to the turbidity of the water sample at the time of sampling. Suspended sediment concentrations estimated from samples of varying turbidity can be used to construct a calibration curve which relates the two variables (Figure 2.6). As a result of major shifts in sediment concentration over the ablation season several calibration curves were constructed to relate to different turbidity ranges on the meter and to differences following re-calibration of the meters to ranges additional to those originally set on the meter (Appendix 1.).

In addition to these calibration samples, water samples were also collected from a spatial network of sites (at the same time and location as spatial sampling of electrical conductivity) 500 ml polypropelene bottles (30 mm neck diameter) were used to collect the samples because use of the USDH-48 sampler proved to be too time consuming and cumbersome. Spatial samples need to be collected rapidly, and in the same sequence, in order to sample approximately the same water parcel as it moved downstream (Gurnell, 1982). On one occasion the

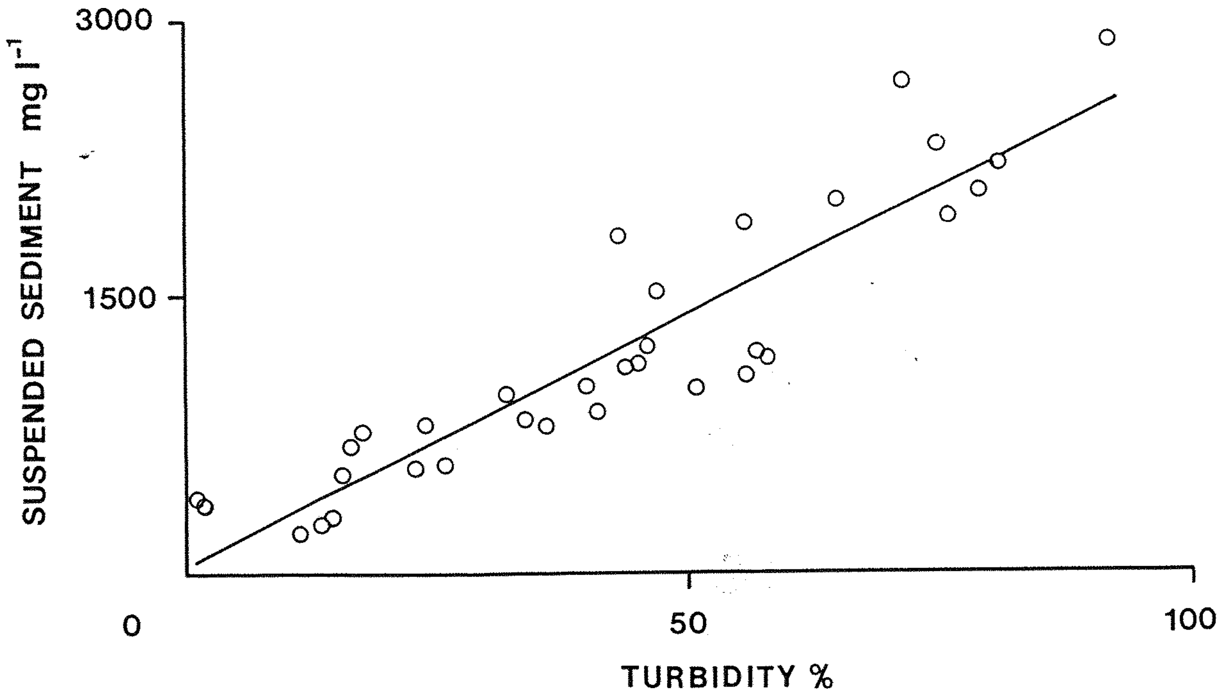


Figure 2.6 Turbidity (%) - Suspended sediment concentration (mg l^{-1}) calibration relationship for Haut Arolla proglacial stream 1986 (Partech set on Range 2).

efficiency of the method was tested because a sediment pulse was released from the glacier at the same time as the spatial sampling commenced. It proved possible to sample in advance of the sediment wave at all of the sampling sites.

A Manning pumping sampler (Figure 2.2) was employed to provide additional turbidity calibration samples and to sample tributary flows. Problems were encountered when using the sampler in shallow tributaries because of siltation of the intake. It should also be noted that a calibration experiment cast some doubt on the comparability between hand samples and pump samples (Appendix 1.).

2.3.6 Bedload transport

Information regarding the movement of bedload was derived from several sources:

a) The 'purging' of the 'gravel' trap on the Grande Dixence meltwater intake structures (Figure 2.4, Section 2.3.1).

b) The accumulation of bedload in small-scale basket traps located in the bed of tributary streams.

c) Accumulation of sediment behind fence traps set up across tributaries.

d) The gauging of instantaneous bedload movement using a Helley-Smith type bedload sampler.

e) Bed material tracer studies.

These studies were carried out in both the mainstream and some of tributary streams.

The functioning of the Grande Dixence gravel trap has been outlined in section 2.2 and is discussed in Chapter 3. The data required for the calibration of the trap include the timing of purging of the trap, the volume of accumulated sediment in the trap when it is purged and the packing density of the trapped sediment.

As mentioned in Section 2.2, trap purge times are determined from water stage records, and trap volume is estimated from field survey using a weight suspended from two ropes (Figure 2.4). A source of error in identifying purges from stage records is operator error (Figure 2.7). This is generally small but tends to increase with the frequency of purging. Over the full ablation season this amounts to a maximum error of +/- 1.8% in total gravel yield when the purge counts from two different operators were compared (Figure 2.7). Field survey using the weight and line method is also subject to an average error in depth estimation of +/- 8.1%. Packing density was estimated from traps placed on the gravel bed of the intake. Once filled with sediment they were removed, the contents were weighed, and a volume - weight conversion factor calculated. Further discussion of these results is to be found in Chapter 3.

Three basket traps of dimensions, 45 x 35 x 25 cm depth, were set in the bed of three of the west bank tributary streams (Figure 2.1). The traps were installed and visited periodically to determine whether sediment had accumulated. Due to the ephemeral nature of stream flow in the tributaries, sediment accumulated aperiodically on an 'event' basis. Following deposition, traps were excavated, the volume and weight of sediment were determined, and sub-samples were taken for grain-size analysis.

Small fence traps were also installed on tributary streams, across the full width of the channel (Figure 2.1). The mesh-size of the fence was 20 mm and so was only suitable for trapping coarse material. These fences were of only limited success, since most were destroyed during the first major flow event in July 1987. The problem appeared to be that organic debris blocked the mesh, producing water ponding behind the fences which in turn lead to the collapse or bursting of the fence.

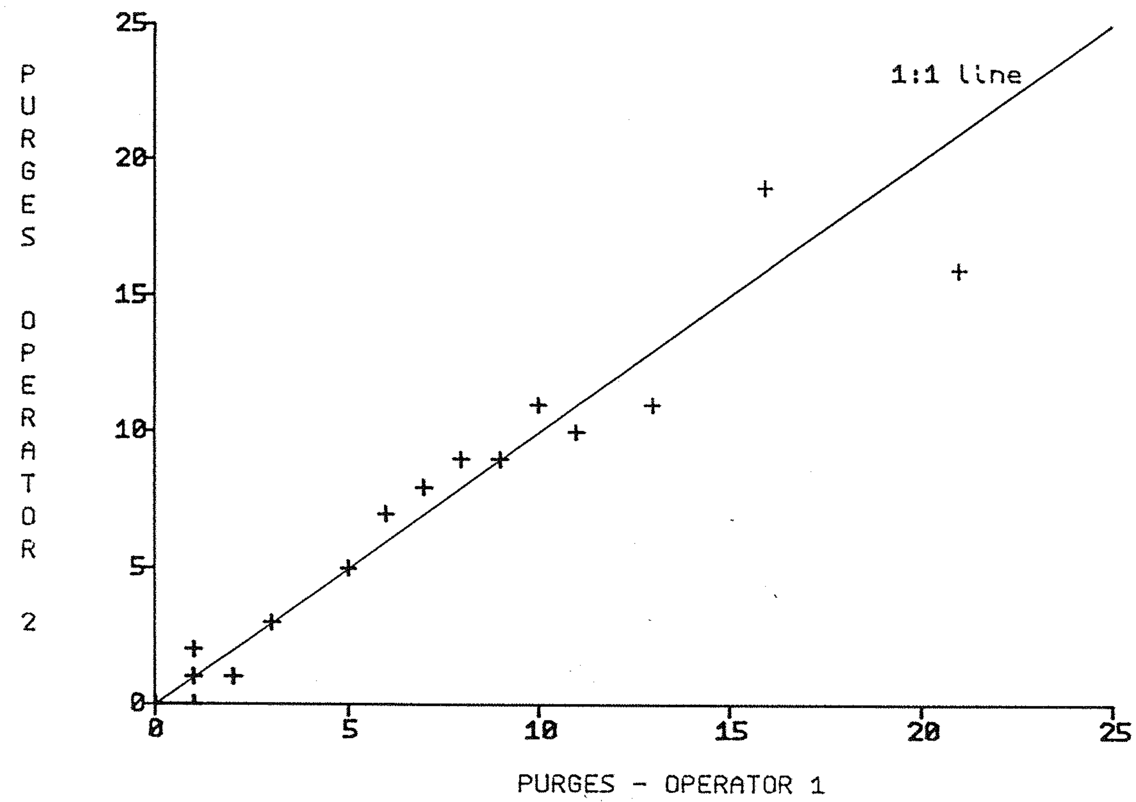


Figure 2.7 Operator error in identifying purges from the stage records. The graph shows the number of purges identified by two different operators for the same stage records.

In order to provide improved temporal resolution over the results from the various traps described above, instantaneous measurements of sediment movement were gauged using a Helley-Smith type bedload sampler. Although direct measurements of bedload in motion have been most successful using fixed structures (e.g. Leopold and Emmett, 1976), these techniques are generally inapplicable to mountain streams where access is usually problematic, where there is a wide range of sediment sizes and where sediment is transported in threads along the channel bed. Under these circumstances a portable sampler offers greater flexibility. The most advanced sampler is the design of Helley and Smith (1971). This pressure difference sampler has been extensively calibrated (Hubbell et al., 1981) and can provide reliable direct measurements of bedload transport. Under the correct conditions, the sampler has a sampling efficiency (the ratio of the amount of bedload collected to that which would have passed through the space occupied by the sampler in the same time, had it not been present) approaching 100% (Emmett, 1980). Originally developed by Helley and Smith (1971) with a 3-inch (76 mm) orifice, the sampler has proved to be very effective (Emmett, 1980). However, for use in very coarse bedload streams, such as those encountered in the present study, a larger, 6-inch (152 mm) orifice sampler has been developed (Bathurst, Leeks and Newson, 1986b). This is an exact 'scaled-up' version of the original Helley-Smith design, except that, a 'boxtail' and wading rods were incorporated to make it more stable when in use. The sampler is equipped with 0.25 mm mesh sampling bags. Samples collected by this means were weighed, the largest clast measured and the majority retained for detailed grain size analysis. Measurements of the flow depth and the velocity were taken concurrently with the bedload measurements and the stream cross profile was also surveyed.

Movement of bed material was also studied using tracers. A bulk sample of clasts, ranging in size from approximately 3 cm to 40 cm intermediate axis, was collected from the bed of the proglacial stream. These stones were dried, painted (using masonry paint) and returned to the stream bed (Figure 2.1). The distance travelled by the tracers was determined from a down stream search. The tracer site was visited daily and a downstream search was undertaken if there appeared to be movement of tracers from the introduction point. Once found, the clasts were measured and labelled and then distance of movement determined.

2.3.7 Soil loss measurements

In an attempt to estimate soil movements adjacent to tributary channels a series of Gerlach-type trough traps were set up in association with a variety of ground-cover types. These traps are similar in design to those described by Statham (1981) consisting of a 50 cm length of guttering embedded in the slope. No lid covered the traps so that they sampled material from surface wash processes and rainsplash. Traps were visited weekly in order to determine the level of sediment accumulation. Traps were cleaned at the end of each sampling season and the residue retained for particle size and organic matter analysis.

2.3.8 River channel and Valley Train form measurements

Changing valley train morphology largely reflects the action of fluvial processes. In order to characterise these changes fluvial forms must be measured at both daily and seasonal timescales; at large and small scales; in main channel and tributary sites; and in order to represent the complete morphology, in planform, cross-section and long profile (Richards, 1982).

Several form measurements were surveyed:

a) Detailed stream cross-section surveys
b) Cross-sections of the valley train
c) Long profiles of the channel
d) Planform photographs and plane table mapping
e) Bank erosion profiles
f) Tributary long profiles
g) Tributary cross-sections
h) Boulder movement survey

Figure 2.1 shows the principal positions of the morphological measurement sites. Daily cross-sections were surveyed in three positions representing upper, middle and lower reaches. The lower of these sites also served as the bedload monitoring section. Valley train cross-sections were surveyed at 28 positions, spaced at 9 m intervals down valley. The positions of the upper and lower sections of this series are shown in Figure 2.1. Valley train surveys were carried out twice in 1986 and 1987. Main channel long profiles were surveyed between the upper valley train cross-section and the meltwater intake (Figure 2.1). Survey of tributaries was difficult due to the shifting ephemeral nature of the channels and so the most reliable survey results were obtained in the quasi-stable lower reaches of these channels (Figure 2.1). Danger from falling debris prevented morphological measurements being extended all the way to the glacier snout.

Form measurements a, b, c and f were all determined using either a Quickset level with distance measured optically or a surveying tape. In all cases cross-sections and profiles were monumented, and tied into fixed points in order that they could be accurately and repeatedly resurveyed. The lack of good ground control is the main problem with surveying in proglacial areas. The planform of the stream network was mapped using plane table methods or oblique high angle photographs of the proglacial zone. The latter method was employed because plane table surveying was very time consuming (indeed, the channel pattern could change before it was completed), and also photography could be undertaken without field assistance.

Photography was employed to particularly good effect at the Bas Glacier d'Arolla site, where the whole proglacial zone could be photographed in a single exposure from the valley wall above using a 35 mm wide angle lens. Photographs were either taken at the same time of day or at a time of equivalent discharge.

Rates of bank erosion were determined by repeated profiling of the same section of bank. This was done using a technique equivalent to the Blong method of slope survey (Gardiner and Dackombe, 1983) with vertical offsets measured from the horizontal. This method was used in preference to use of erosion pins, partly because the pins could not be inserted into the friable materials without causing large scale disturbance, but also because changes were so great that pins were only able to indicate minimum rates of erosion.

Tributary cross-profiles were determined from a series of depths measured from a fixed horizontal datum. A similar method was used for measuring main channel cross profiles when no field assistance was available.

At one of the bank erosion profiling sites six large boulders were marked and their position resurveyed periodically to determine the stability of these large-scale roughness/bank protection elements. Survey was by means of an angle and distance method, distance being measured using a tape and the angle using a compass.

2.4 Laboratory methods

Wherever possible, analyses was carried out in the field (e.g. field filtration of meltwater samples, preparation of samples for total dissolved solids determination, and 'large-particle' size analysis). However, certain analyses needed to be conducted under greater control in the laboratory, including the determination of suspended sediment concentration, the grain-size analysis of fine sediment samples and the determination of total dissolved solids. The techniques employed in each of these three analyses are described below.

2.4.1 Suspended sediment concentration

Fenn (1983) described a 3-phase method for determining suspended sediment concentration. Phase 1 was the careful preweighing of filter papers. This involved oven-drying the papers for 30 minutes at 105 ^{o}C, removal and storage in a desiccator for 5 minutes followed by the weighing of papers (within 30 seconds of being removed from the desiccator) at a resolution of 0.0001 g. Filters were weighed in batches of 20, the first filter being reweighed at the start and end of the cycle in order to determine whether there was a significant change during the time elapsed whilst weighing the batch of 20 papers. This procedure was used for reweighing the filter papers after filtration of meltwater. Phase 2 involved field filtration of samples of meltwater containing suspended sediment through filter papers by creating a partial vacuum of 40-50 cm Hg using a Nalgene hand pump. The filter papers were then folded whilst moist and sealed in polythene bags to avoid loss of sediment prior to reweighing. Phase 3 involved reweighing of the filter papers, by the method described for Phase 1, in order to determine the weight gain of the filters as a result of the trapped sediment. The concentration of the suspended sediment was then calculated by the weight gain on the filter paper divided by the volume of the meltwater sample with the result expressed in mg l^{-1}.

Two types of filter paper were used in the present study; Whatman 40 papers (retention 8 microns) as used previously by Fenn (1983) and Gurnell (1987b), and Whatman GFA papers (retention 1.6 microns). The latter were used for secondary filtering of the filtrate from the former in order to establish suspended sediment loss through the Whatman 40 papers for meltwaters from the Bas Glacier d'Arolla. Figure 2.8 illustrates that the loss is not consistent and can be as great as 251 mg l^{-1}. There is also no clear relationship between the suspended sediment concentration of the water sample and the amount of sediment in the residual size range 8-1.6 microns. This suggests that meltwater suspended sediment characteristics in the Bas Arolla stream vary in a non-systematic way (Figure 2.8) and may be a product of variable sources of sediment supply. However, the average error in determining total suspended sediment concentration using only Whatman 40 papers is an under estimation of 7% (of total suspended sediment concentration down to 1.6 microns).

2.4.2 Particle size analysis

Particle size analysis can be an invaluable aid to the interpretation of sediment transport hypotheses and potential sediment source areas (Bowles, 1984; Walling and Kane, 1984; Ashworth and Ferguson, 1986). Two main types of analysis were used: dry sieving (-6 to 4 phi range) and analysis by Coulter Counter (3.63 to 7.63 phi range). Individual large particles were also measured in the field using calipers and pebblometer boards, results being expressed in terms of the lengths of the three principal axes.

For dry sieving, bulk samples were air dried, physically

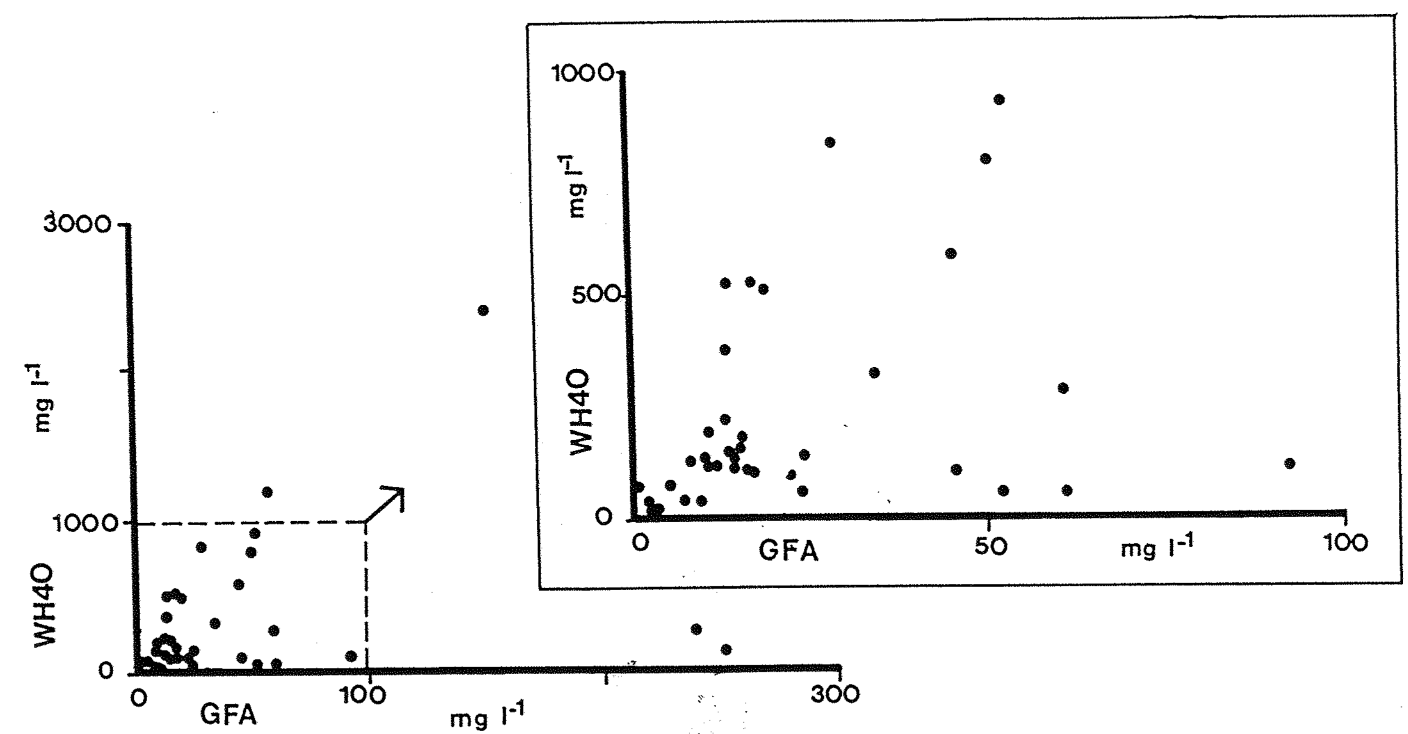


Figure 2.8 Relationship between greater than 8 micron suspended sediment concentration determined by filtering water samples through Whatman 40 papers (WH40) and 1.6 to 8 micron suspended sediment fraction (determined by filtering through Whatman GFA papers) of the filtrate from the samples filtered through the Whatman 40 papers.

disaggregated, air dried again and then sieved using a mechanical shaker. The amount of material retained on each sieve was weighed and then the sediment passing through the 4 phi sieve was retained for Coulter counter analysis.

The Coulter Counter can accurately measure small amounts of sediment using an electrical sensing zone method (see Whalley 1981, and Coulter Counter reference manual). The model used here was the TAII with population accessory PCA1. A 280 micron Coulter Counter tube was used for the analysis. The size range analysed by the Coulter Counter complemented that of dry sieving and included particles down to 7.63 microns, below which there was very little remaining in the samples analysed (Section 2.4.1), partly because clay-grade particles may remain in aggregates despite the modest physical disaggregation employed in the laboratory procedure.

Often particle size data in published studies are partitioned into the components of bedload, suspended load and dissolved load. However, the boundary between suspended sediment and bedload is difficult to determine though, based on published estimates and data collected from suspended sediment samples, a value of 2.88 mm (-1.5 phi) seems reasonable for proglacial streams. This is somewhat coarser than the 0.0625 mm (4 phi) limit proposed by Walling and Kane (1984) for streams in SW England but, in mountain streams such a value is more realistic (Stott, 1987). However, such a division is to some extent arbitrary and unless quantitatively determined is best avoided.

2.4.3 Total dissolved solids

Seventeen samples of approximately 750 ml were collected and filtered in the field (see Section 2.4.1) and were transported back to the laboratories in Southampton in sealed glass bottles for the determination of total dissolved solids concentration (Cryer and Trudgill, 1981).

2.5 Summary of Measurements

Understanding of the inter-relationships between fluvial sediment transport processes and sediment storage sites depends on the identification of relevant processes and the precision and duration of field measurements. The basic material transfer components of the fluvial sediment system are shown in Figure 2.9. This shows the main measured elements and their relationship to transport via the main stream channel. Because sediment monitoring occurred at a distance downstream from the glacier snout, inputs of glacial material to the narrow zone in the immediate vicinity of the glacier margin (moraine deposition) (Figure 2.8) were estimated. Slope inputs from hillslope failure were also unmeasured since no failures occurred within the period of study. Each of the elements shown in Figure 2.9 is considered in a separate section broadly grouped as either tributary or channel sediment sources. Sources are then placed within the context of an integrated framework for the proglacial fluvial sediment system. This framework refers to processes acting during the ablation period, but excludes the role of snow in modifying slopes and channels and must therefore be considered as a 'summer' system only.

In summary, measurements, as suggested in Figure 2.9, can be conveniently divided into four groups: 'outputs', within-channel, channel margins and valley train bluffs and side slope tributaries. Contributions from the valley train bluffs and tributaries are viewed as 'inputs' into the sediment transfer system, whilst within channel measurements are really concerned with the processes moving sediment or the 'throughputs'. Generally it is assumed, over the timescale of this study, that sediment transfers due to hillslope processes and bedrock erosion are negligible. However the instability of adjacent morainic slopes can make significant contributions to the sediment stored in the

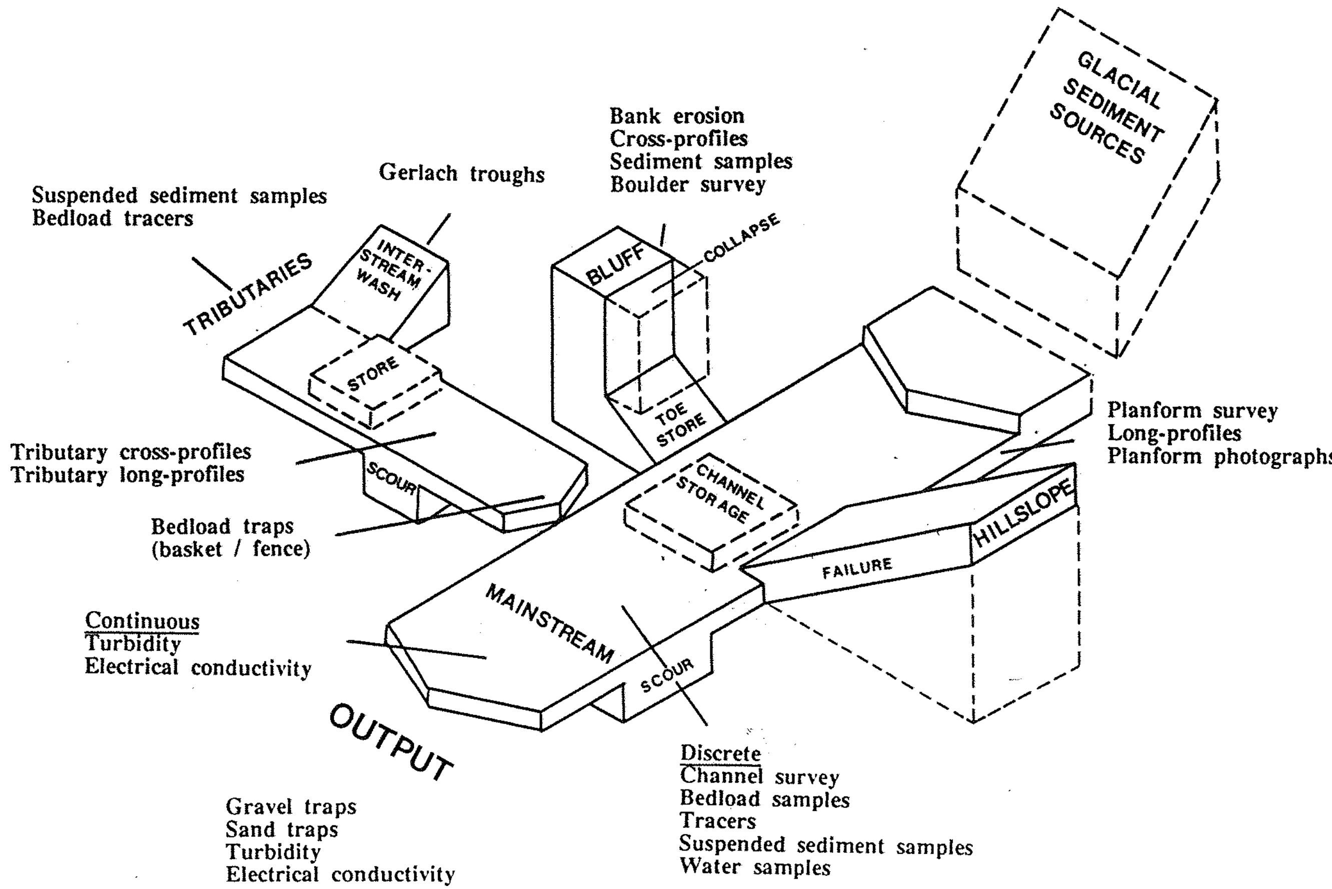


Figure 2.9 Inter-relations between fluvial sediment transport and field measurements

valley train. Although this was not measured directly, the potential importance of these processes is demonstrated in later sections.

Chapter 3.

SEDIMENT OUTPUT RESULTS

3.1 Introduction

The output of material from a catchment, by fluvial processes depends on the balance between sediment availability and the ability of the fluvial system to transport sediment. According to Jansson (1988) this is the definition of sediment yield from the basin, representing net erosion or the difference between gross erosion and the input of sediment into storage over a particular time period. The discrepancy between gross erosion and net erosion can be expressed as a ratio which is termed the sediment delivery ratio (Roehl, 192; Petts and Foster, 1985). Understanding sediment yield components can be hampered by preconceived notions of bedload, suspended load and solute loads since the boundaries between the components are arbitrary and in reality highly variable. This is especially true of proglacial streams where rapid and dramatic changes in sediment supply and stream energy, both in time and space, produce large fluctuations in the boundary between material transported as suspended or bedload. Therefore in this argument rather than 'splitting hairs' over the partitioning of load components sediment load data are expressed as a total (i.e. the summation of the measurement components) or in terms of the actual measured components e.g 'gravel trap load' or 'Helley-Smith bedload'. The main pitfall to avoid is duplicating the measurement of a proportion of the load, because the summation of the measurement components would then lead to an over-estimate of sediment yield. Alternatively techniques should be designed to sample the entire load to avoid gaps or overlap in the ranges of the components that are estimated, Whichever approach is used care is needed in reporting loads so that approximate comparisons can be made between sediment yield data from different basins and studies. Even this is a compromise since yields will only be truly comparative if all measurement and sampling techniques are standardised.

Although the fluvial geomorphological literature is replete with studies which use sediment and solute system outputs to infer spatial and temporal controls on yields (Richards, 1982) understanding of sediment and solute yield processes is still limited. This is reflected in the dominance of empirical sediment transport studies because no acceptable theoretical base exists (Walling and Webb, 1983). Emphasis on 'output' rather than the mobilisation and routing of sediment and solute-laden water has delayed understanding of sediment and solute delivery processes. The dynamics of sediment transport also complicate attempts to quantify the total yield of solutes and sediment from a basin (Richards, 1982).

Sediment output estimates provide limited information about denudation and erosion rates given the intervening role of storage in modulating the relationship between erosion and sediment yield (Meade, 1982; Phillips, 1986). Indeed outputs, as erosion estimates, are only valid if storage is zero or changes in storage occur on timescales at an order of magnitude faster than the period for which denudation is being studied (Phillips, 1986). This is probably particularly true of glacierized basins where rates of sediment production by glacial erosion (which are probably related to changes of glacier activity over many years) are unlikely to equal the annual rates of sediment transport by meltwaters. One of the few attempts to quantify the intervening effects of sediment storage in proglacial systems is the Borland Index (Borland, 1961) in which a term including the length of the glacier stream between the snout and gauging station is included. In proglacial environments (i.e valley train and sandur) this effect might be better quantified by a measure of area, which would better reflect the opportunities for storage. Accurate estimates of erosion rates require detailed close-interval sampling of meltwater sediment load over an extended period of years (Collins, 1979). This is more true of bedload and suspended load than solute load since, "in the immediate proglacial zone adjustments in solute load appear to be negligible whereas suspended sediment transport and bedload transport can vary widely" (Gurnell, 1987).

Sediment yield estimates based on the summation of streamload sediment components have often been used as a crude means of estimating denudation and erosion rates (Petts and Foster, 1985). This practice has a long history of application in glacierised catchments (e.g. Helland, 1877) where direct estimation of glacial erosion is fraught with difficulties due to the inaccessibility of the glacier bed (Collins, 1981). However, records of sediment yield from alpine glacier basins remain scarce, due to the problems of establishing measurement procedures capable of coping with the possibility of high and variable sediment transport rates (Gurnell, 1987). With the exception of work carried out in Norway (Østrem and others - e.g. Østrem, 1975) and Switzerland (Collins, 1981, 1983; Gurnell and Clark, 1987) sampling in proglacial streams has been infrequent, sampling frameworks have varied drastically (Gurnell, 1982) and the lengths of observed record have been inconsistent. The result of this is that the records which are available are difficult to use even for inter-basin comparison and rarely cover a sufficient time period to permit estimation of glacial erosion rates or long-term sediment yield rates.

Individual streamload components have received variable research attention. Suspended sediment yield has been studied most widely and in more detail than bedload and dissolved load components (Collins, 1979; Gurnell, 1987). Bedload is notoriously difficult to measure in proglacial streams because of their steep gradients, extreme variations in bed material size, large roughness elements and erratic sediment supplies. Solutes, suspended sediment and bedload undergo sucessively more discontinuous transport responding variously to changes in stream discharge and sediment supply (Richards, 1982). The bias towards suspended sediment measurement in many glacierised catchments stems from the difficulties in monitoring bedload and from the assumption that dissolved load forms an insignificant component of the total load. The limited available data suggest that in the context of glacier basin research both bedload and dissolved load are important components of total sediment yield (Church and Gilbert, 1975; Eyles et al., 1982). Further, despite the lack of data clastic sediment is known to cause problems when water is abstracted for hydro-electric use (Aegerter and Messerli, 1983; Bezinge, 1987) so further study has both scientific and applied merit.

Techniques for estimating sediment transport in proglacial streams have varied widely but one of the most common techniques has been to derive sediment rating curves (Borland, 1961; Mathews, 1964; Østrem, 1975; Collins, 1979; Fenn et al., 1985). Sediment rating curves have been most widely applied to the estimation of suspended sediment yield from glacierised basins but similar techniques have been extended to bedload and solute load components (Hammer and Smith, 1983). Sediment rating curves have been found to be in significant error when compared with actual sediment yields (Walling, 1977, 1978) although the precise magnitude is very variable because of widespread use of log transformations of the variables prior to estimation of the rating curve.

A recent analysis by Ferguson (1986a) of the effects of log transformation of the variables, indicates that suspended sediment loads may be under estimated by up to 50% if correction factors are not applied. The limitation of these techniques for use in proglacial areas has been stressed by Fenn et al. (1985) and alternative procedures involving multiple rating curves for different parts of the melt season (Østrem, 1975), multivariate regression approaches (Richards, 1984) and time series analysis (Gurnell and Fenn, 1984) have all been suggested as alternatives to the conventional approach. More recently, suspended sediment and bedload yield have been estimated for two glacier basins using information on the filling and emptying of sediment traps (Bezinge, Clark, Gurnell and Warburton, 1988; Gurnell, Warburton and Clark, 1988). This technique provides a useful data base for testing the more conventional sediment rating curve methods described above and might also be used to assess the accuracy of load estimates derived using sediment transport formulae (Church, 1972).

The aim of this chapter is to provide estimates of bedload, suspended load and solute load in order to calculate total sediment yield estimates for the Bas Arolla, Tsidjiore Nouve and Haut Arolla basins in 1986 and for the Bas Arolla catchment in 1987. Solute loads are only estimated for the Bas Arolla catchment because this is the only meltwater system for which a total dissolved solids - electrical conductivity relationship was determined.

Methods used in the present study to estimate sediment output involved the estimation of coarse sediment load (Bedload) from records of purging of sediment traps in meltwater intakes; the monitoring of suspended load using turbidity meter records calibrated to suspended sediment concentration; and the estimation of solute load from continuous electrical conductivity records calibrated to total dissolved solids. In 1986 sediment trap purge (emptying) records were collected and continuous monitoring of turbidity was undertaken on the Bas Arolla, Tsidjiore Nouve and Haut Arolla proglacial streams. In 1987 measurements were confined to the Bas Arolla basin. The strategy for using these data to estimate suspended and bedload is given in Gurnell et al. (1988) but can be summarised as follows (Figure 3.1): suspended sediment was monitored using a turbidity meter either at the entrance to the meltwater intake (A) or inside the intake structure (D) and the records were transformed into estimates of suspended sediment concentration by a calibration curve based on hand sampling. Water samples collected between these two sites allowed calibration curves to be constructed for both sites so that sediment loss between the two sites could be calculated. This calibration allowed instrumentation to be sited at locations A or D and comparable stream loads could still be calculated. The volume of bedload was estimated from surveys of sediment accumulating in the gravel trap (Site B, Figure 3.1). This was purged periodically (when full) causing a perturbation to be recorded at the stage recorder (Figure 3.1). The volume of sediment prior to purging was combined with estimates of packing density, obtained by accumulating bed material containers on the bed of the trap (B), to provide estimates of weight of sediment removed during each purge of the trap. These methods were outlined in Chapter 2 and are discussed in more detail in the first part of this Chapter (Section 3.2). Three topics are discussed: calibration of the Bas Arolla gravel trap (volumetric survey and material properties), estimating changes in sediment concentration between the stream and meltwater intake; and determining a total dissolved solids - electrical conductivity relationship. Seasonal rating curves for bedload, suspended load and total dissolved solids are also calculated.

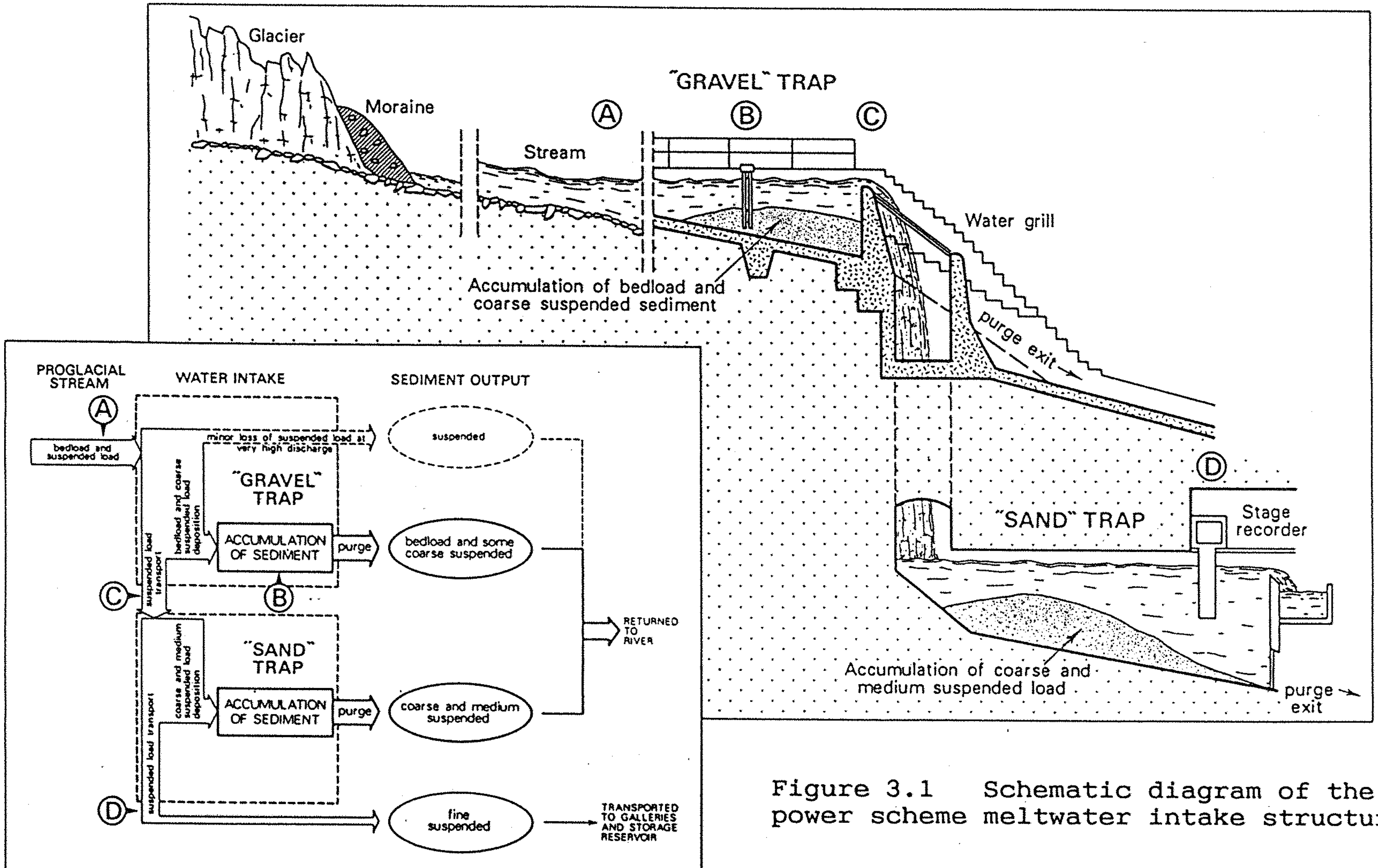


Figure 3.1 Schematic diagram of the hydro-electric power scheme meltwater intake structure.

Results of using the above methods are discussed in the following sections together with a comparison of sediment yields from the three study basins in 1986 and contrasts in yields from the Bas Arolla basin in 1986 and 1987 (Section 3.3). Sediment rating curve methods are also applied to these data so that load estimates derived by the two methods can be compared and differences or 'errors' quantified (Walling, 1977). These yields are then considered (Section 3.4) in a regional context (by using variations in sediment trap purging for 18 basins in the Val d'Hérens, Switzerland) and then in a global context by comparing sediment yields from other glacierized, mountainous and upland basins that have been reported in the literature (Section 3.5). The chapter is concluded with a summary (Section 3.6).

3.2 Discussion of calibration and measurement techniques

3.2.1 Calibration of the Bas Arolla gravel trap.

In order to convert the record of gravel trap emptying or purging into a meaningful estimate of load three important properties require measurement:
1) Volume of sediment accumulated in the trap
2) Material properties and packing density of the sediment
3) An estimate of the flushing efficiency
The procedures for calculating 1 and 2 have been outlined in Chapter 2. and estimates of flushing efficiency are based on field observations. This section provides a more detailed description of field calibration of the Bas Arolla gravel trap and discusses details, problems and results from such calibrations.

Sediment traps associated with hydro-electric schemes have been used in several studies, as a means of estimating sediment transport, (e.g. in Norway (Wold and Østrem, 1979) Austria (Lauffer and Sommer (1982) and Switzerland (Beecroft, 1983)). Problems in using traps to measure sediment transport in mountain streams have been briefly reviewed by Raemmy and Jaeggi (1981) who conclude that the best means of determining event-based sediment yields is by a sluicing system (of the type used in this study). Indeed sediment problems are common at run-of-the-river water intakes and a considerable body of research has focussed on the hydrodynamics and environmental implications of such structures (e.g. ASCE Hydraulics Division, 1988).

1) Trap Volume - Table 3.1. summarises the results of surveys carried out on the Bas Arolla gravel trap in 1987 by the weight and line method described in Chapter 2. A selection of these surveys are plotted in Figure

Table 3.1 Summary of Bas Arolla gravel trap sediment surveys 1987.

Survey date and time		Sediment volume m^3 [1]	Total period Accumulation hours [2]	Accumulation rate $m^3\ hr^{-1}$ [3]		Infill factor [4]
27.5	11.30	179.62	---	---		---
29.5	12.20	176.09	---	---		---
1.6	13.00	153.80	---	---		---
9.6	16.40	127.95	102.9	4.69	(2.85)	0.27
17.6	12.00	137.29	173.5	1.19	(0.76)	0.66
27.6	11.30	125.72	232.5	0.69	(0.42)	0.78
29.6	16.15	65.21	36.7	37.26	(8.69)	0.05
2.7	06.30	57.60	5.3	50.08	(6.61)	0.22
5.7	15.30	61.38	90.8	0.80	(0.15)	0.85
9.7	13.00	50.48	6.5	0.0	(0.0)	0.0
12.7	14.20	81.53	11.2	8.49	(3.28)	0.86
30.7	18.45	48.68	97.5	3.89	---	0.13

Notes:

1. Sediment volume is the volume at time of survey
2. Total period of accumulation = mumber of hours between last and next purge
3. Accumulation rate = volume at time of survey divided by number of hours of accumulation prior to survey. Bracketed figures = volume of survey minus $50m^3$ (volume in trap immediately after purge) divided by number of hours accumulation.
4. Infill factor = time since last purge divided by total period of accumulation between purges.

3.2. Volumes of sediment vary between 48.68 m^3 and 179.62m^3 for conditions immediately following purge (July 7th, 13.00) and times when the trap was full of sediment (May 27th, May 29th, June, 6th and June 17th, respectively). Plotting the sediment volume against an infill factor (time since last purge divided by total period between purges), as shown in Figure 3.3. yields a large scatter in points. If it was assumed that sedimentation rates were similar a linear plot would be expected, however due to the large variations in sediment transport rates, of accumulation vary enormously. This is clearly illustrated in Table 3.1 with transport rates (adjusted for post purge infill) of between 3.28 - 0.15 m^3 per hour, a 21 fold difference. Figure 3.3. also provides an illustration of The 'post purge infill effect' in that no surveyed volumes are significantly less than 50 m^3. The largest surveyed volumes were in late May when the trap was full of sediment. These values are higher than the volumes recorded prior to purging later in the season. The reason for this is that finer sediments are deposited by the very low flows early in the season so that the surface profile of the accumulated sediment is different from those developing later in the season (Figure 3.2.C). Because the early season 'humped' profile is not evident later in the season, the maximum accumulated sediment volume is best represented by the mean of the surveys for the 1st June and 17th June (145.5 m^3).

Figure 3.2. illustrates the nature of sedimentation in the trap. The three surveys shown along with the bathymetry of the trap show marked variations in sediment accumulation. Sediment tends to accumulate in a wedge on the side of the trap where the main thread of streamflow enters. Sediment is transported down-trap and deposited over an avalanche face shown by the break in slope in examples C and D. Although the trap is designed to trap all the water and hence all the suspended sediment, at discharges in excess of 4.2

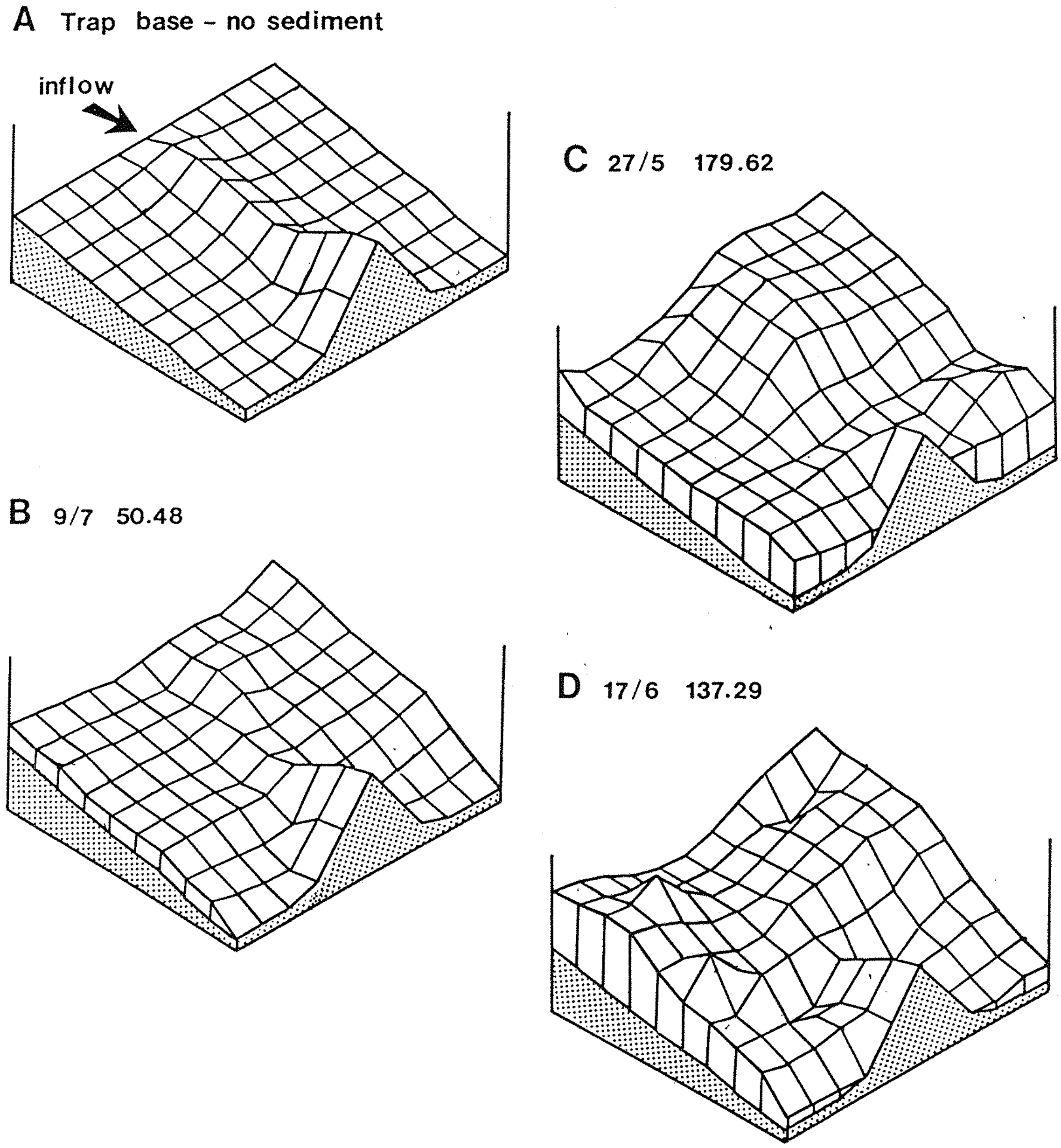


Figure 3.2 Isometric plots of sediment accumulation in the Bas Arolla gravel trap 1987.

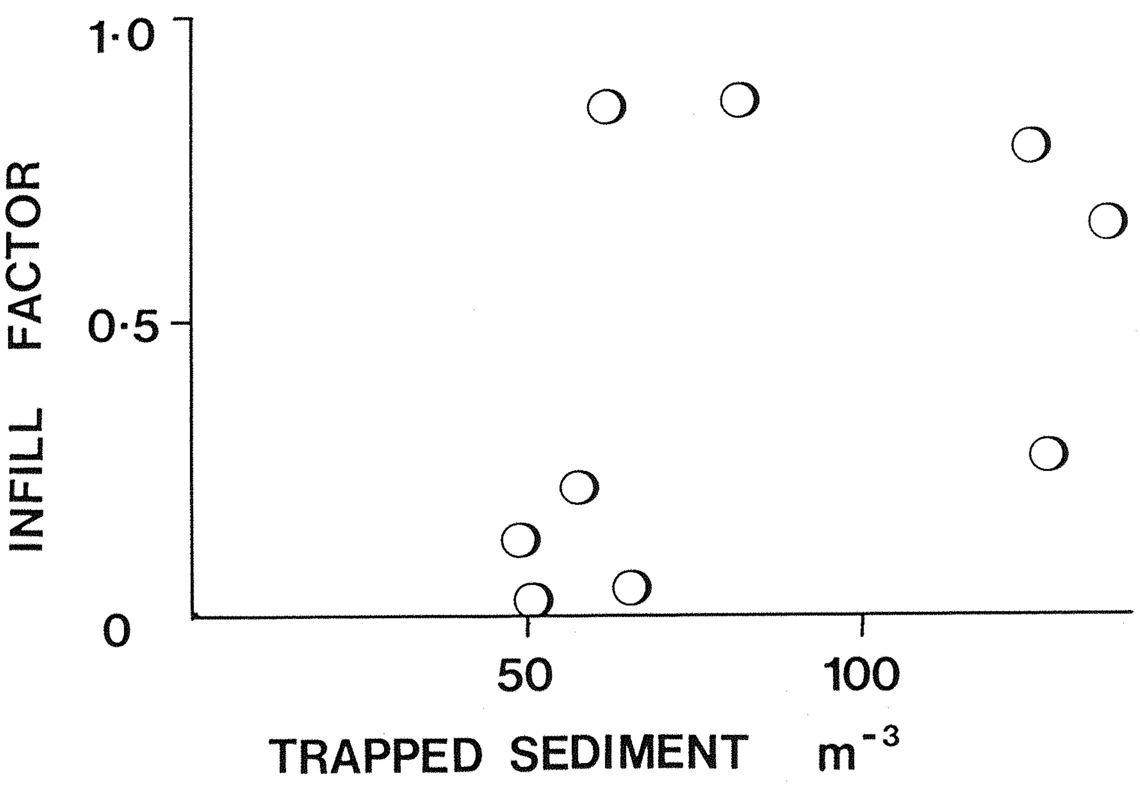


Figure 3.3 Trapped sediment volume plotted against trap infill factor for the Bas Arolla gravel trap 1987. Definition of trap infill factor given in Table 3.1.

m^3s^{-1} overspill at the front of the trap occurs and sediment and water are lost downstream.

2) Material properties - Comparison of grab samples taken in the trap on the 27th July (Figure 3.4.) show interesting variations. On this day inflow was on the eastern side of the trap and the series of down trap samples taken on this side are coarser than samples towards the western side of the trap. Although no down trap trend is discernible samples at 0 and 3 m have the coarsest elements. A sample taken at the avalanche front within the trap is the finest. Clearly this is only representative of one day and variations are bound to exist. For example Figure 3.5 shows grain size curves of samples collected for packing density estimates; these samples show contrasts in their distributions, particularly in their finer proportions.

Packing density estimates determined from sediment deposited in samplers located in the gravel trap (Figure 3.5.) gave a packing density of 1.3 tonnes.m^3 (standard deviation = 0.1 t m^{-3}, n = 9). This is lower than the packing density estimated at in the Tsidjiore Nouve gravel trap where a higher value of 1.63 tonnes.m^{-3}(standard deviation = 0.1 t m^{-3}, n = 16) was measured. These differences probably represent a difference in the grain sizes deposited in the trap, because sediment in both basins is derived from the same source rocks which have a bulk density of 2.57 t m^{-3} (standard deviation = 0.16 t m^{-3} n = 8). Variations in packing density exist (Figure 3. 5.) but the low standard deviation of the samples suggests that the average packing density gives an acceptable estimate for use in calculating the weight of sediment purged from the traps.

3) Flushing efficiency -In calculating the amount of material purged from the sediment trap no account was taken of the proportion of material that might have been

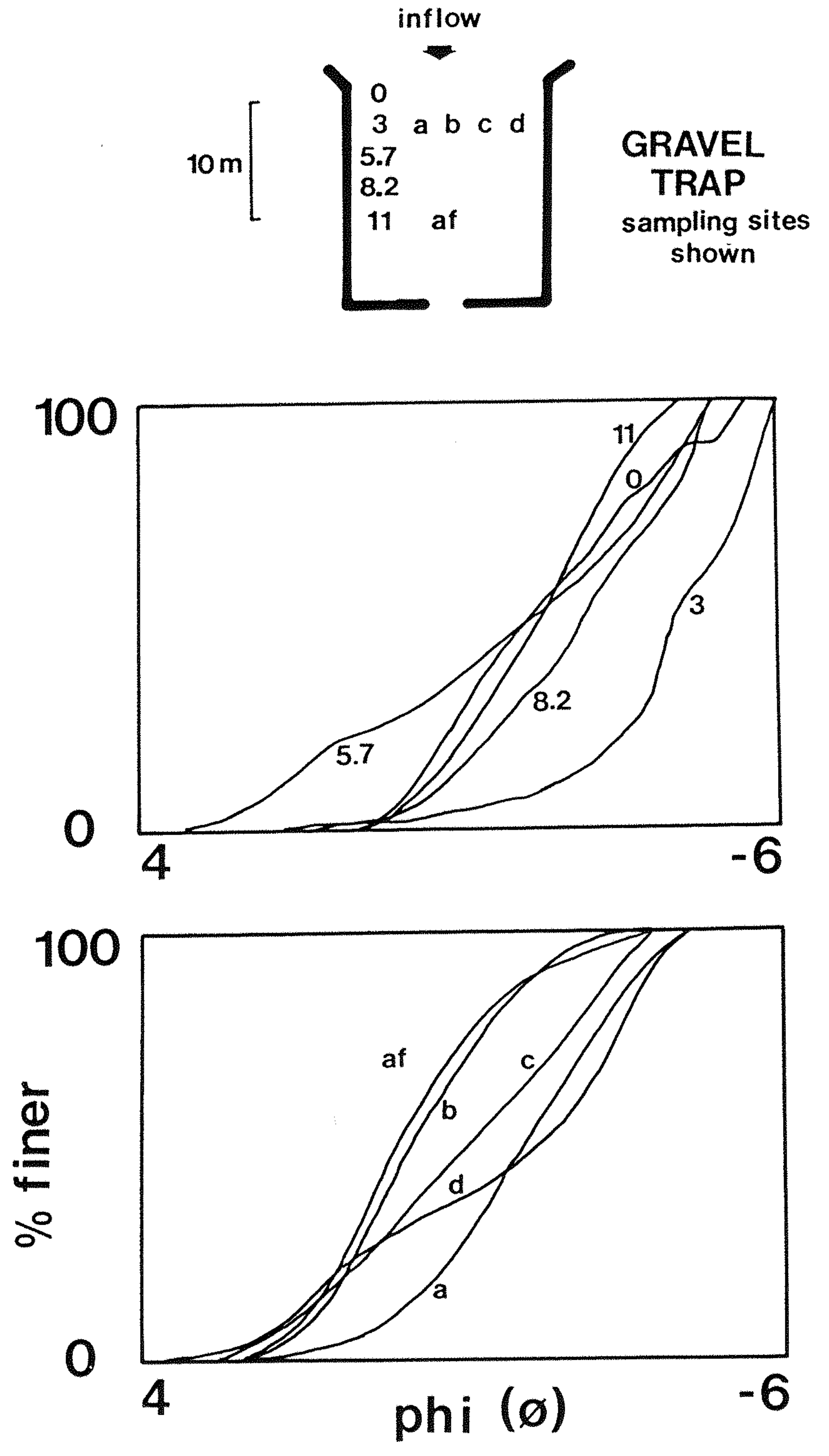


Figure 3.4 Grain size curves (by weight) of surface sediment samples from the Bas Arolla gravel trap July 27th 1987. Distances measured down from trap entrance.

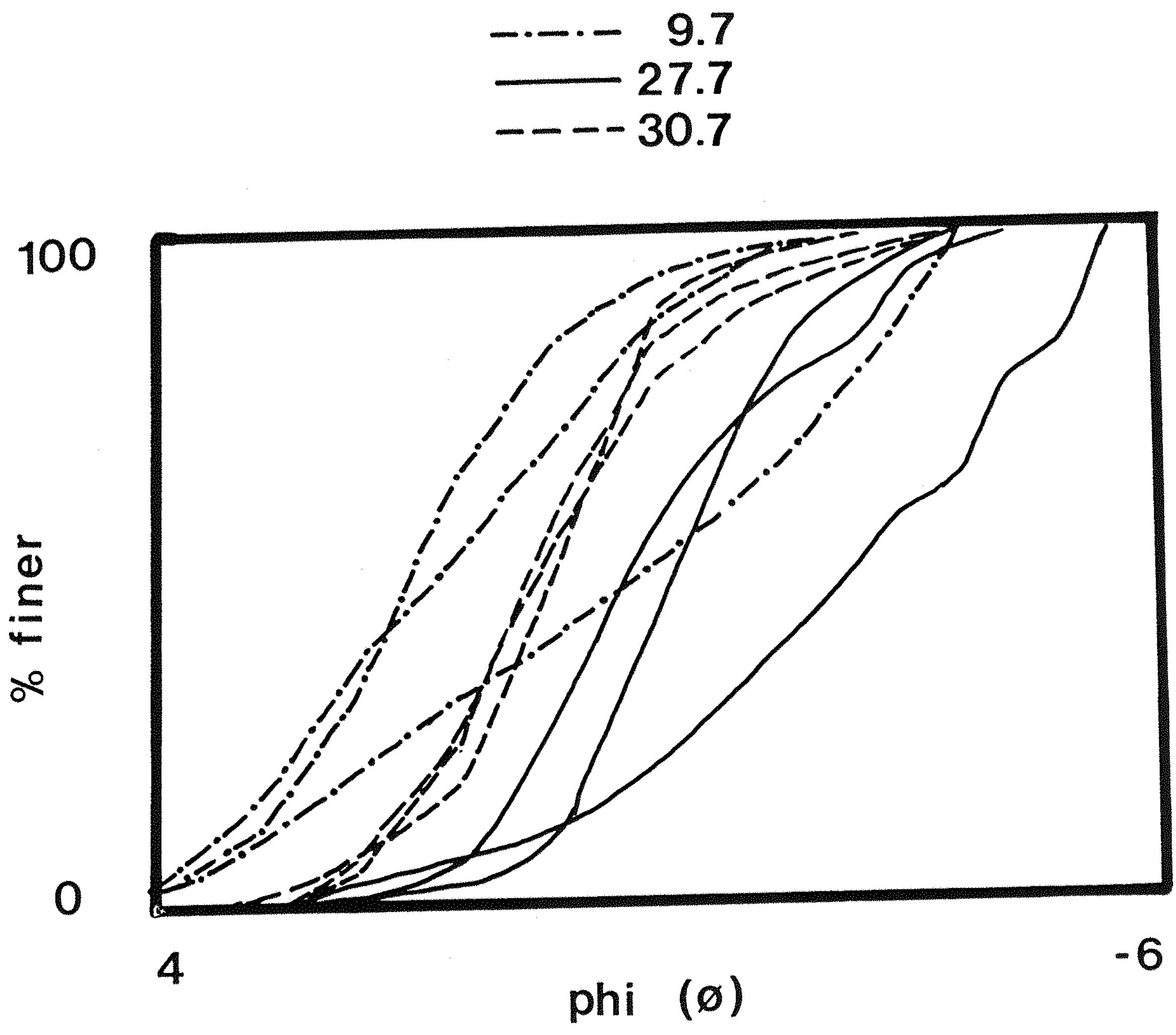


Figure 3.5 Grain size curves (by weight) of packing density trap samples from the Bas Arolla gravel trap 1987.

purged from the channel immediately above the trap due to readjustment of the stream bed/stream bed profile. The correct method for accounting for this mass of sediment would be to extend the cross-profiling upstream of the trap entrance to a point where the differences in the profile (bed elevation) are negligible before and after purging. Such bed adjustment did not occur at the Tsidjiore Nouve trap because the upper limit of the trap is controlled by a line of iron ties in the channel bed. However it becomes apparent that such adjustments did occur above the Bas Arolla trap. In order to account for this discrepancy the volume of sediment purged by this means was approximated by extrapolating both the concrete base profile of trap and the surveyed sediment wedge surface profile upstream, immediately before the purge (Figure 3.6). Downstream of the point of intersection of these two profiles is a sediment wedge which is a fair approximation of the upstream material that would be lost during purge (Figure 3.6). Based on the assumption, from field observations that the distance upstream affected by the purge is about 10m, the volume of the sediment wedge is calculated as 49 m^3 , given 0.9 m depth of sediment accumulation at the rear of the trap. In the field it is possible to see a 'high' pre-purge water line on the banks when the stream-bed and water profiles have readjusted. During purging an avalanche front extends headwards up the stream channel, calving sediment as material is excavated from the front of the trap. Therefore there is an abrupt step in the profile rather than a smooth gradual transition. Following purging the bed profile rapidly readjusts and material is quickly accumulated in the trap. The existence of this process is supported from the prise surveys which indicate that about $50m^3$ is found in the trap immediately after purge (Figure 3.2.D). This process occurs even though the actual flushing efficiency of the trap is observed to be close to 100 %. Figure 3.6 is not to scale but it illustrates these processes, which may vary in magnitude depending

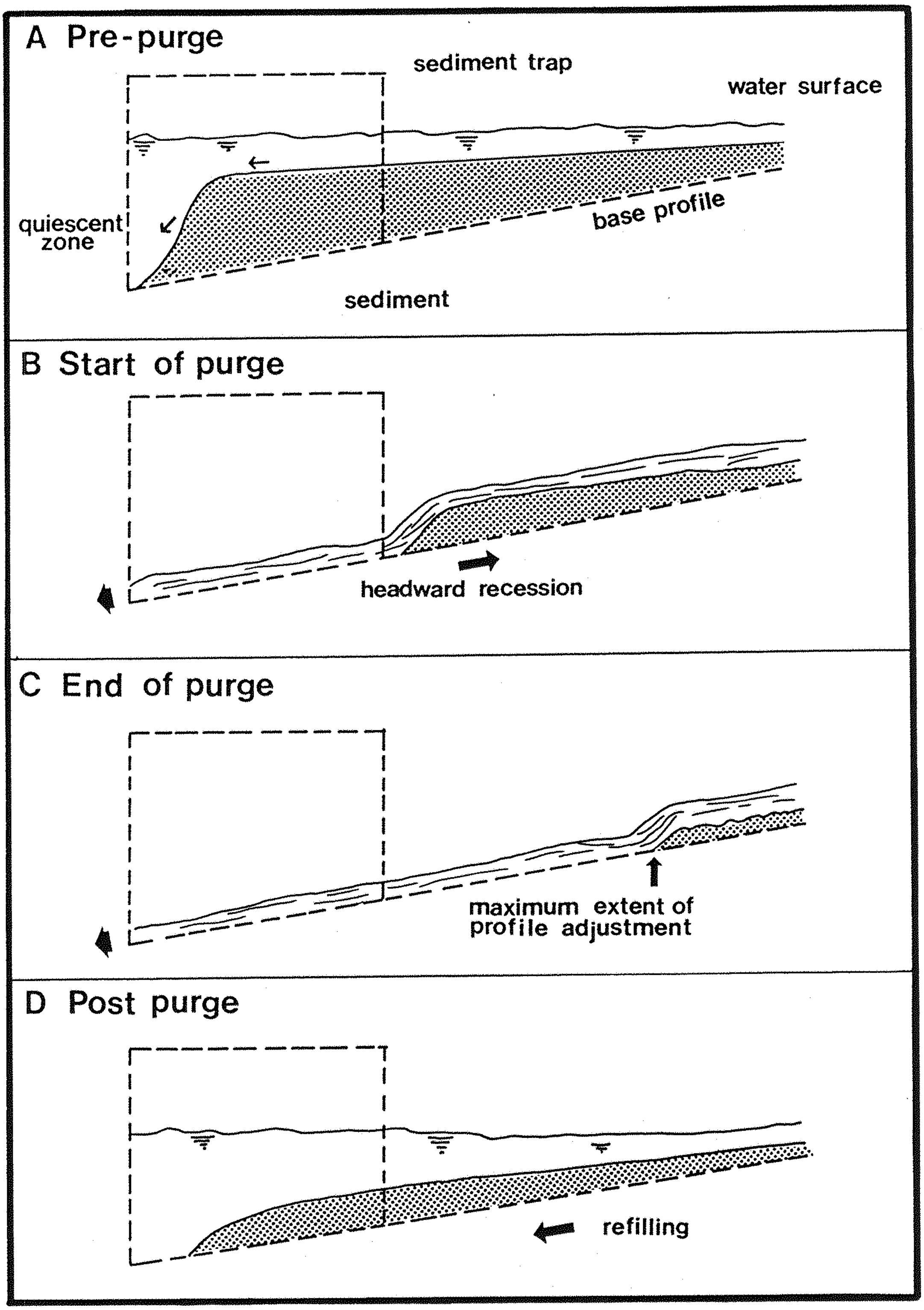


Figure 3.6 Diagram illustrating the processes acting during a gravel trap purge on the Bas Arolla proglacial stream.

on the prevailing water-sediment discharge conditions. Figure 3.6 illustrates the rapid removal of stored sediment, erosion of the channel bed upstream of the sediment trap and the almost instantaneous partial refilling of the bedload trap once the sluice gates have been closed. Following purging sediment accumulates in the quiescent confines of the trap, reducing the local bed slope which in turn causes increased deposition upstream, so establishing a positive feedback. Bedload is transported over the surface of the sediment accumulated in the trap and is deposited at an avalanche front at the leading edge of the sediment wedge (Figure 3.6.A). Because of this mode of sedimentation, a quiescent zone is created in the front of the trap which can promote the settling out of the coarse suspended sediment.

3.3.1 Suspended sediment concentration variations between the stream and meltwater intake.

Instrumentation for monitoring suspended sediment concentration was established at two sites on the Bas Arolla meltwater intake (Sites A and D, Figure 3.1). Careful field calibration was required to provide estimates of stream suspended sediment concentrations and reliable comparisons between the two locations. In 1986 a turbidity meter was located at D and in 1987 the meter was located at A from May to July and then at D from the end of July to September. Calibration was necessary between the recording sites so that all records could be corrected to suspended sediment concentration in the stream at site A.

Observations of suspended sediment entering and leaving traps at the Bas Arolla and the Tsidjiore Nouve water intakes and inspection of sediment samples collected from the bed of both the gravel traps indicated that a considerable amount of suspended sediment is deposited in the gravel trap at Tsidjiore Nouve but considerably less in the Bas Arolla trap (Gurnell, Warburton and Clark, 1988). This suggests that the pattern of purges from the gravel and sand traps at the Bas Arolla intake approximate the pattern in bedload and suspended load transport, although this does not include that proportion of suspended load transported from the basin (Which can be estimated from the turbidity record at site D).

The characteristics of suspended sediment transport through the Bas Arolla trap were further investigated in two ways, from: (1) suspended sediment grain size comparisons between sites A and D; and (2) down-trap sampling of particle size characteristics of sediments deposited in the gravel trap between sites A and C.

Figure 3.7 presents results of the calibration between sites A and D at the Bas Arolla site. The majority of observations plot close to the 1:1 relationship line indicating that there is little difference between the sediment concentrations at the two sites. However several values deviate from this trend and it is hypothesised that differences exist because of changes in the coarse grain size fraction of a water parcel as it passes through the trap, which would explain why all marked deviations plot above the 1:1 line. To investigate this hypothesis four paired sample sets were examined and their grain-size characteristics estimated using a Coulter Counter for the range 12.7 - 256 microns with the fraction greater than 25 microns also recorded. Paired samples A and B (Figure 3.8a) were selected because they showed roughly comparable sediment concentrations as opposed to C and D which were very different. The graphs (Figure 3.8b) indicate that differences in sediment concentration are produced generally by differences in the amount of coarse material in the sample because as flows are altered this is the fraction which is most easily deposited. The

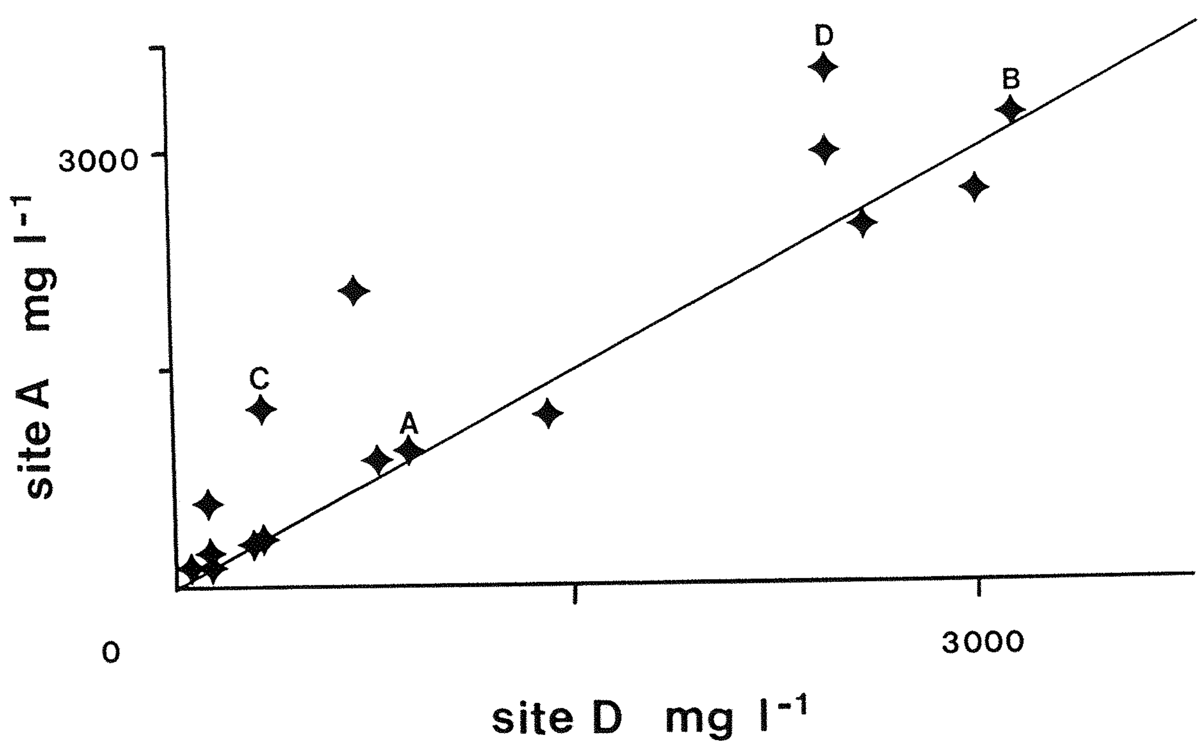


Figure 3.7 Calibration of suspended sediment concentration between sites A and D on the Bas Arolla gravel trap.

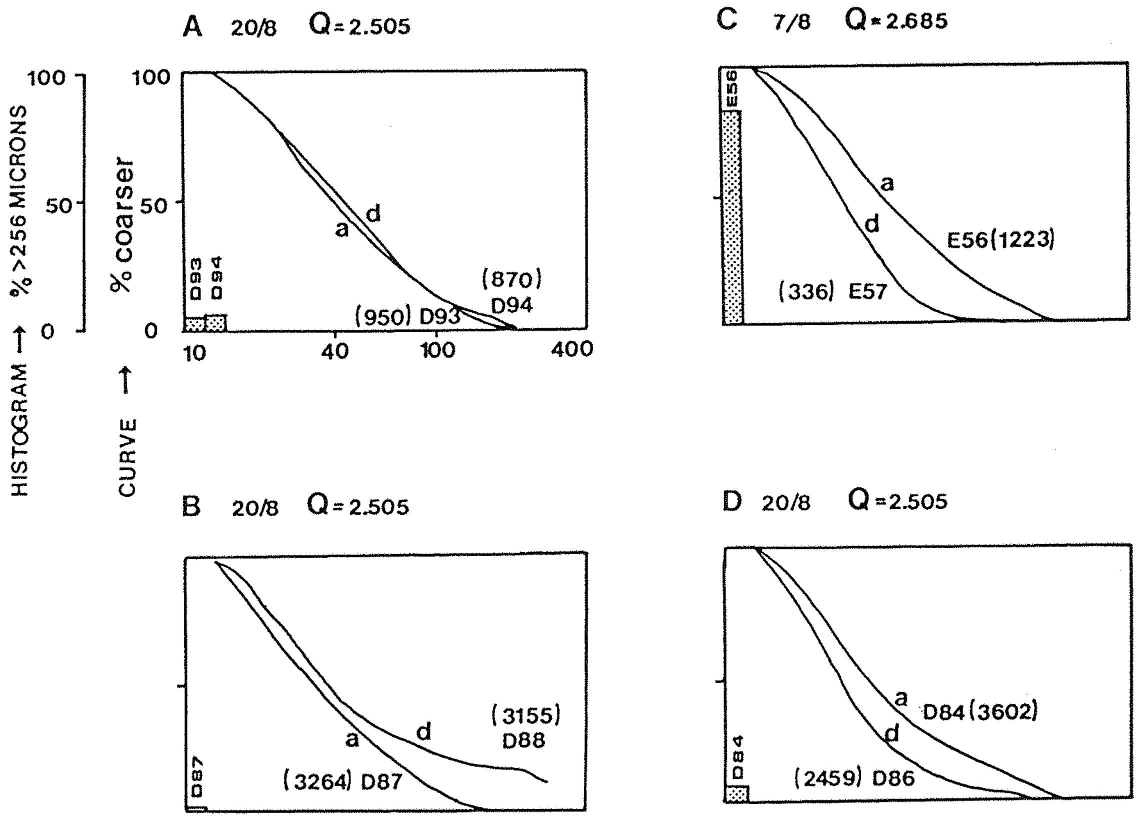


Figure 3.8 Comparison of grain size curves (range 12.7 to 256 microns) for paired samples collected at sites A (a) and D (d) on the Bas Arolla meltwater intake structure. Q = discharge at time of sampling $m^3 s^{-1}$, histograms show the residual sample (%) greater than 256 microns and suspended sediment concentrations of the samples are shown in brackets.

most marked difference in sample C results from coarse sediment (81 % by weight coarser than 256 microns) in the stream sample at site B whereas the sample collected at site D had no material coarser than 256 microns. Differences in the finer fraction also exist but their effect is less marked; the curves in B deviate in their coarse fraction but this is compensated by sample d87 having slightly more material in the fraction coarser than 256 microns. Truly comparable samples would have exactly the same distributions as shown in A (Figure 3.8).

Given that a minor proportion of the suspended load appeared to be being deposited in the gravel trap, samples were taken in a down trap direction in order to determine the nature of suspended sediment variations within the trap. Results (Figure 3.9.) show a general fining towards the front of the trap, sample D78 being the only exception. However this was sampled at the edge of the avalanche front and may have some temporarily suspended bedload deposited in it. Therefore a small proportion of the suspended load is deposited in the gravel trap but this was too small for accurate separation from the large volumes of bed material accumulating in the trap.

Because differences exist between monitoring positions, calculations of sediment load must compensate for these. At the Bas Arolla site where there is little evidence of suspended sediment deposition in the gravel trap, the best approach seemed to be to establish a calibration between suspended sediment concentration at site A and site D and then to estimate suspended sediment concentration at site d from the turbidity records.

$$\text{Site A (mg l}^{-1}\text{)} = 1.004 + 234.8 \text{ Site D (mg l}^{-1}\text{)}$$

This equation allowed suspended sediment concentration estimates to be derived for site A, above the gravel

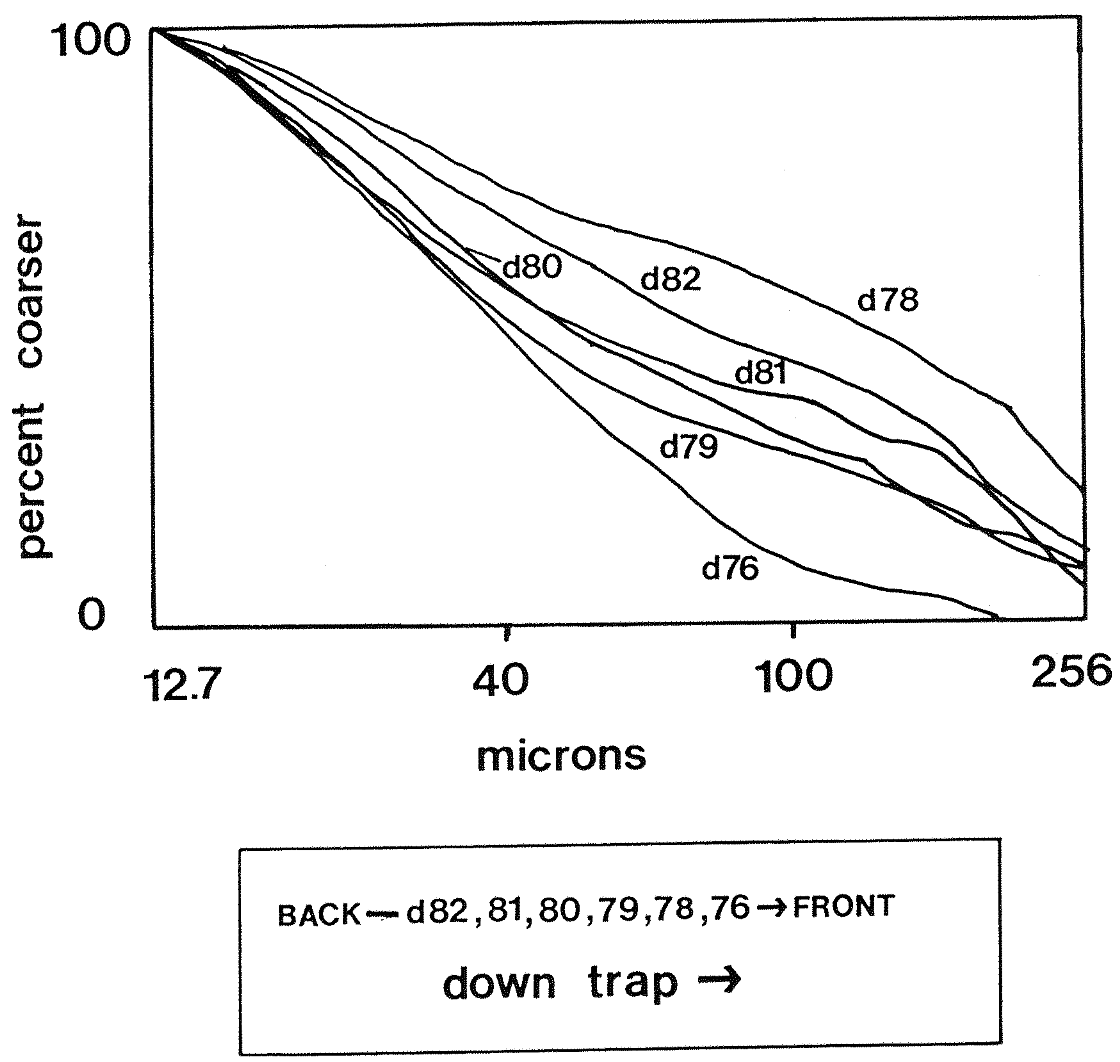


Figure 3.9 Suspended sediment samples collected down the length of the Bas Arolla gravel trap 1987.

trap, regardless of the positioning of the turbidity meter. Sediment accumulating in the gravel trap was then estimated to approximate bedload transport.

A more detailed trap calibration strategy was necessary at for the Tsidjiore Nouve intake structure because it was evident from visual inspection of material deposited in the gravel trap that a considerable amount of suspended sediment was being deposited. Calibration of this trap was undertaken by Gurnell and is reported in detail in Bezinge, Clark, Gurnell and Warburton (1988). In essence, hand samples samples were taken at three sites (A, C and D, Figure 3.1) on the Tsidjiore Nouve intake structure and were calibrated against turbidity which was recorded at site D. The three different calibration curves were used with the turbidity records to estimate the amounts of suspended sediment passing sites A, C and D or being deposited between sites A and C (in the gravel trap) and C and D (in the sand trap). Deposition rates were divided by the frequency of trap purging to show that approximately 50.5 tonnes of suspended sediment were associated with each gravel trap purge (the remainder being bedload, surveyed using the same measurement technique as used at Bas Arolla) and 19 t of suspended sediment associated with each sand trap purge. These calibrations for trap purging plus estimates of the amount of sediment passing through the intake structure were applied to two ablation seasons for which detailed (1 hourly and 4 hourly) of suspended sediment concentration, determined by pumping sampler, were available. The estimate of suspended sediment yield from the purge record were within 6% of those from the filtered pumped samples.

At Haut Arolla where there is a single sediment trap access is restricted therefore estimates of trap volume and sediment characteristics are provided by Grande Dixence.

3.2.3 Dissolved load estimation

The use of electrical conductivity as a surrogate measure of total dissolved solids is widespread in sediment yield studies (Walling and Webb, 1975). Total dissolved solids may be estimated in this way because differences in conductivity are mainly the result of differences in the concentration of charged solutes and to a lesser extent in the nature of those charged solutes and temperature, (Thomas, 1986). Quantitative estimates of solute load should therefore be possible from electrical conductivity records (Collins, 1983). This is usually achieved by relating total dissolved solids or the summation of the cations, to conductivity, with an arbitrary limit distinguishing particulates from solutes (for example Walling and Webb, 1983 used 0.045 microns). The total dissolved solid - conductivity relationship is then combined with continuous conductivity and discharge measurements to obtain an estimate of total solute load in the stream. Detailed descriptions of this method are given in Hem (1970), Cryer and Trudgill (1981) and Walling (1984). A major attraction of using electrical conductivity is the ease and economy with which measurements can be made and continuously recorded. Nevertheless, it should be acknowledged that a complex multivariate relationship exists between electrical conductivity and the summation of the ions, which is influenced by: the ionic composition of the solution, ionic concentration, water temperature, pH, suspended sediment load and dissolved organic matter (Fenn, 1987).

Because only the estimation of total solute yield was required in the present study individual ion concentrations were not determined, and electrical conductivity was simply related to total dissolved solids determinations resulting from the evaporation to dryness of prefiltered water samples. Since the main relationship sought was that between electrical conductivity and total dissolved solids.

The basis for electrical conductivity monitoring of glacial streams is that meltwater in contact with mineral grains become enriched with solutes (Collins, 1983) which enhances its conductivity. Figure 3.10 illustrates field observations of the response of supraglacial (low conductivity) water samples to mineral particles of different particle size. Similar experiments by Lemmens and Rogers (1978) illustrate the same rapid enrichment of meltwaters on exposure to glacial sediments, with maximum increases in electrical conductivity associated with exposure to fine-grained glacial sediments.

Electrical conductivity measurements have been used widely in proglacial stream studies (Fenn, 1987) to investigate temporal and spatial variations in meltwater quality (Gurnell and Fenn, 1985); to provide estimates of solute load and yield (Collins, 1983); to assess hydrochemically distinct contributions to the total meltwater discharge (Collins, 1977); and to identify solute sources and pathways allowing an indirect means of sampling glacier basal attributes (Collins, 1981). In this study electrical conductivity is used as a surrogate measure of total dissolved solids concentration and as a qualitative indicator of water sources (e.g. snowpack, tributary or glacial runoff).

Determinations of total dissolved solids, from 17 meltwater samples from the Bas Arolla proglacial stream are shown in Figure 3.11 (mean of all samples, 35.2 mg l^{-1} with a standard deviation of 6.5mg l^{-1}, coefficient of variation, 18%). These values are similar to determinations of between 10 - 50 mg l^{-1} from the Tsdijiore Nouve meltwater stream (Lorrain and Souchez, 1972) and the Gornergletscher (Collins, 1981) and fall within the range of values for rivers developed on igneous and metamorphic rocks, which exhibit low total

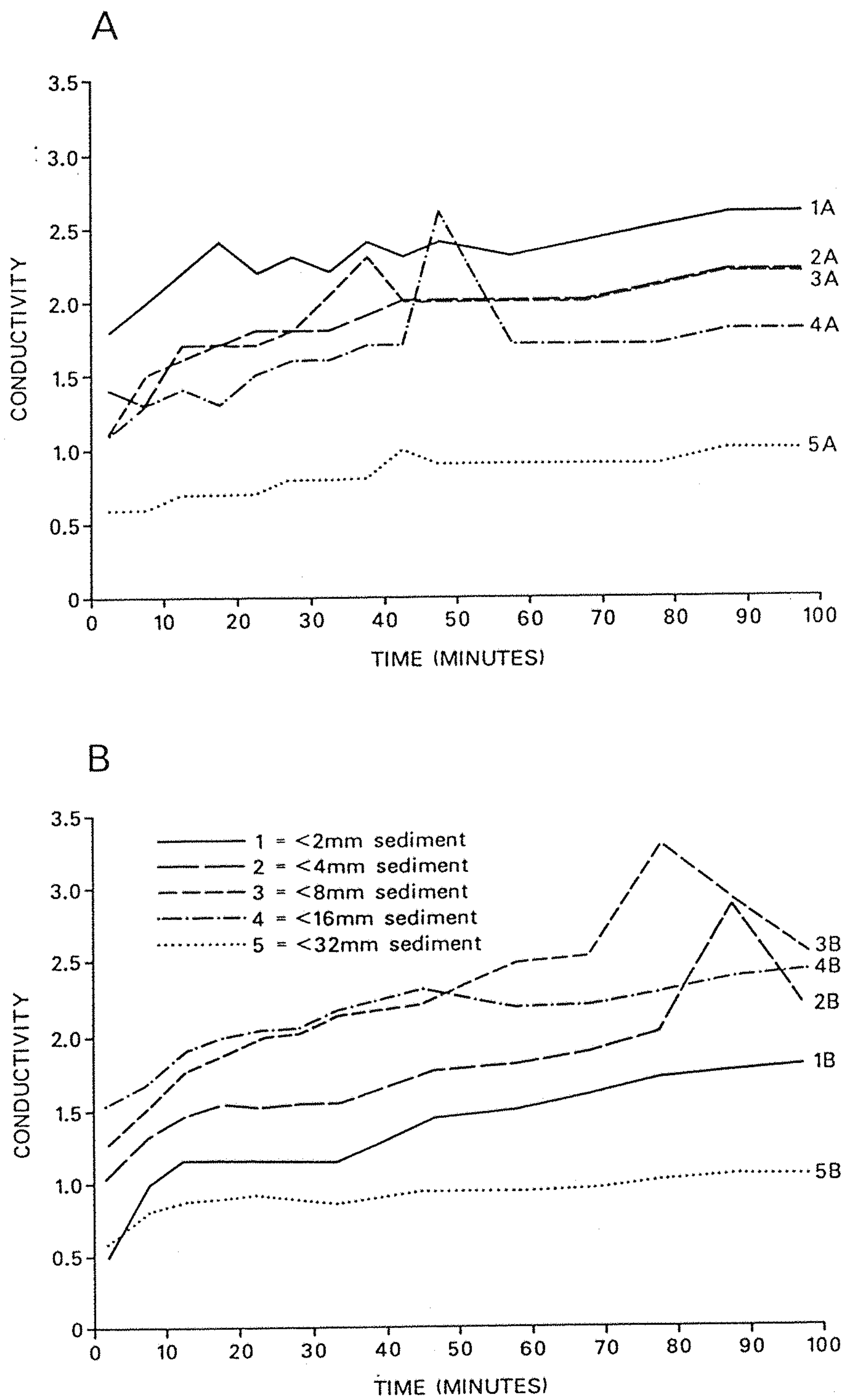
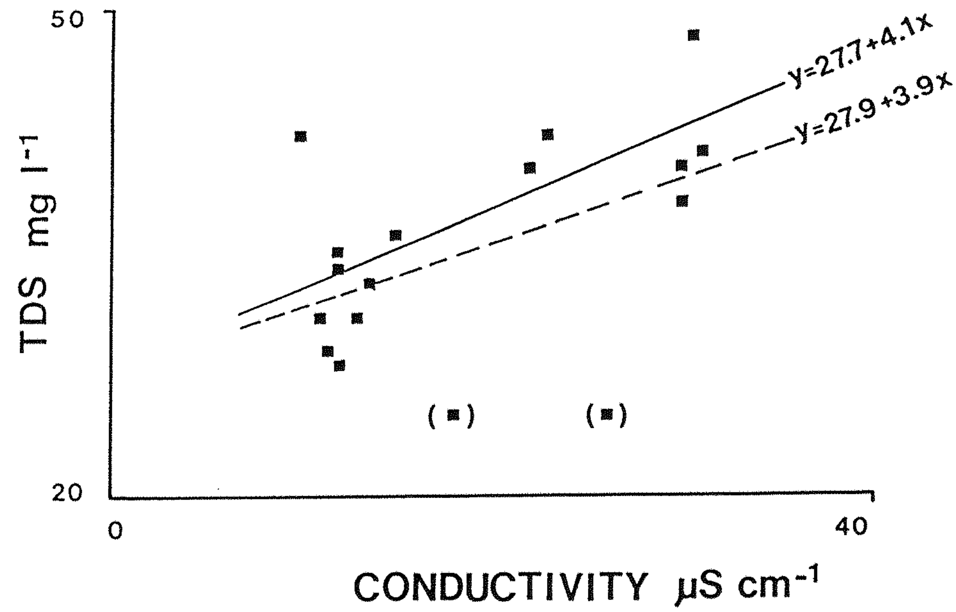


Figure 3.10 Changing conductivity with time in relation to particle size. A) 265 g sediment. B) 530 g sediment.

dissolved solids concentrations of the order of less than 50 mg l^{-1} (Walling, 1984). The large apparent scatter in the graph (17 points) yields an R^2 of 22.15% (F and T tests show this relationship to be non-significant at the 0.05 level of probability). However careful examination of the data show that two points (shown in brackets in Figure 3.11) plot low on the graph. These two points occur on the same day (25th June 1987) and appear to be lower than the general trend in predicted total dissolved solids. Streamflow and rainfall records together with field observations indicate that this is an overcast 'rainfall' day with high discharge from tributary streams of low conductivity (8 $\mu s.cm^{-1}$) water. The main tributary input is 50 m upstream of the main measurement site and the two samples on June 25th may have been affected by inadequate mixing of mainstream and tributary water sources. The impact of poor mixing could have been enhanced by the fact that conductivity at the main monitoring site was monitored with a probe mounted at 60 cm below the water surface, but meltwater samples (used in the determination of total dissolved solids) were collected at a depth of 30 cm. Variations in conductivity result from the transport of separate water parcels have been identified on the proglacial stream of Gornergletscher (Metcalf, 1986). In addition, variations in cross-section conductivities may have a pronounced effect when several water sources are contributing, particularly at low flows (Anderson 1963). Day (1975) showed the mixing length of solutes even in narrow, fast-flowing alpine streams, to be of the order of 25-250 m which is in excess of the 50 m distance to the nearest tributary encountered in this study

For these reasons the two anomalous values were omitted in estimating the total dissolved solids - electrical conductivity calibration curve. This resulted in a much stronger relationship (R^2 49.9%, F and T statistics

Figure 3.11 Relationship between total dissolved solids of Bas Arolla meltwater samples and conductivity 1987.



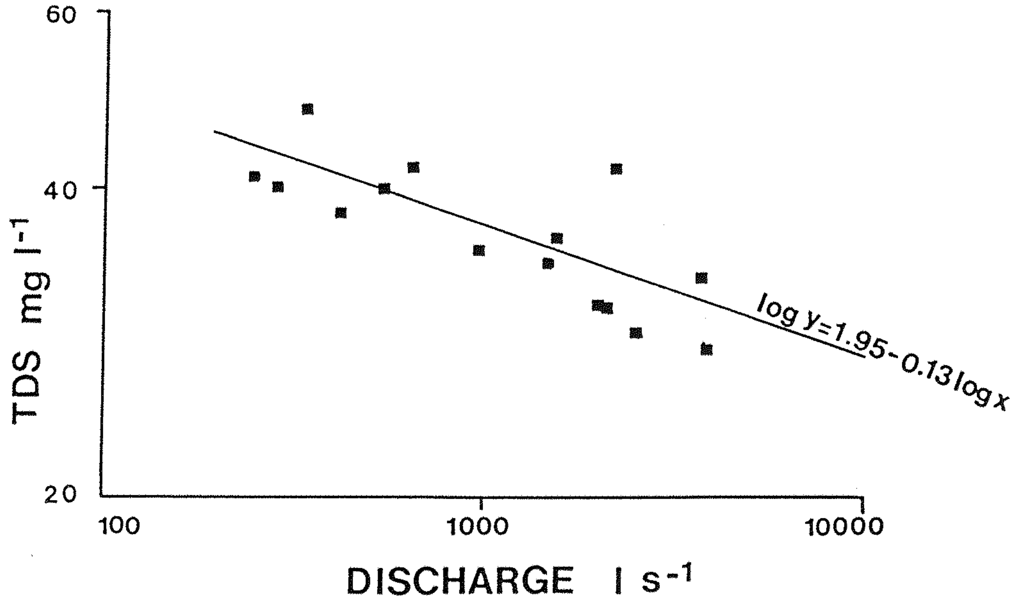


Figure 3.12 Total dissolved solids - discharge rating relationship Bas Arolla proglacial stream 1987.

significant at the 0.05 probability level) (Figure 3.11). The positive intercept in this regression relationship is not uncommon and may be due to the presence of undissolved ions or a significant silica content (Walling, 1984).

The reason for the weaker relationship defined here in comparison with studies such as that of Walling (1974) is that both electrical conductivity and total dissolved solids concentrations in proglacial streams are extremely low. The high solute loads often attributed to meltwater streams is a result of high discharge rather than high concentrations of solutes in the meltwater.

Figure 3.12 provides a dissolved solids rating relationship based on discharges at the time of sampling of the 15 meltwater samples. The estimated inverse regression relationship (Figure 3.12) between total dissolved solids and discharge indicates the impact of dilution of total dissolved solids concentration by increasing discharge.

3.3 Variations in sediment yield in the Bas Arolla, Tsidjiore Nouve and Haut Arolla basins

3.3.1 Sediment trap purges and sediment yield estimates

For the period 1977 to 1987 sediment trap purges have been identified and documented for the Haut Arolla single trap water intake and the Bas Arolla and Tsidjiore Nouve dual sediment trap (gravel and sand traps) meltwater intake structures. Years 1982 for Bas Arolla and 1978 for Tsidjiore Nouve are excluded from the analysis due to malfunctioning of the water stage recorder from which purges are identified. Also extensive works on the Haut Arolla intake during 1985 resulted in low estimated yields for that year.

Based on the frequency of purges Figure 3.13 shows a relationship between average weekly purge frequency and average weekly runoff volume which reflects a mean ablation season pattern over the years of record. The graphs indicate a general correspondence between runoff and purge frequency. In the early ablation season purge frequency is low, but it than increases to an early peak, in comparison to discharge, after which the maximum frequency is not exceeded. There appears to be a short lag between the purging of sand and gravel traps for the Bas Arolla and Tsidjiore Nouve basins which may reflect differences in the phasing of bedload and suspended load transport.

Inter-annual patterns of purge frequency (both sand and gravel traps) can be compared with annual discharge volume, and for Bas Arolla and Tsidjiore Nouve, with variations in annual glacial snout advance (Figure 3.14). Over the period 1977 to 1987 there is a generally increasing trend in all four variables at Bas Arolla, with a period of higher purge frequency and discharge from 1981 to 1983. Tsidjiore Nouve shows an

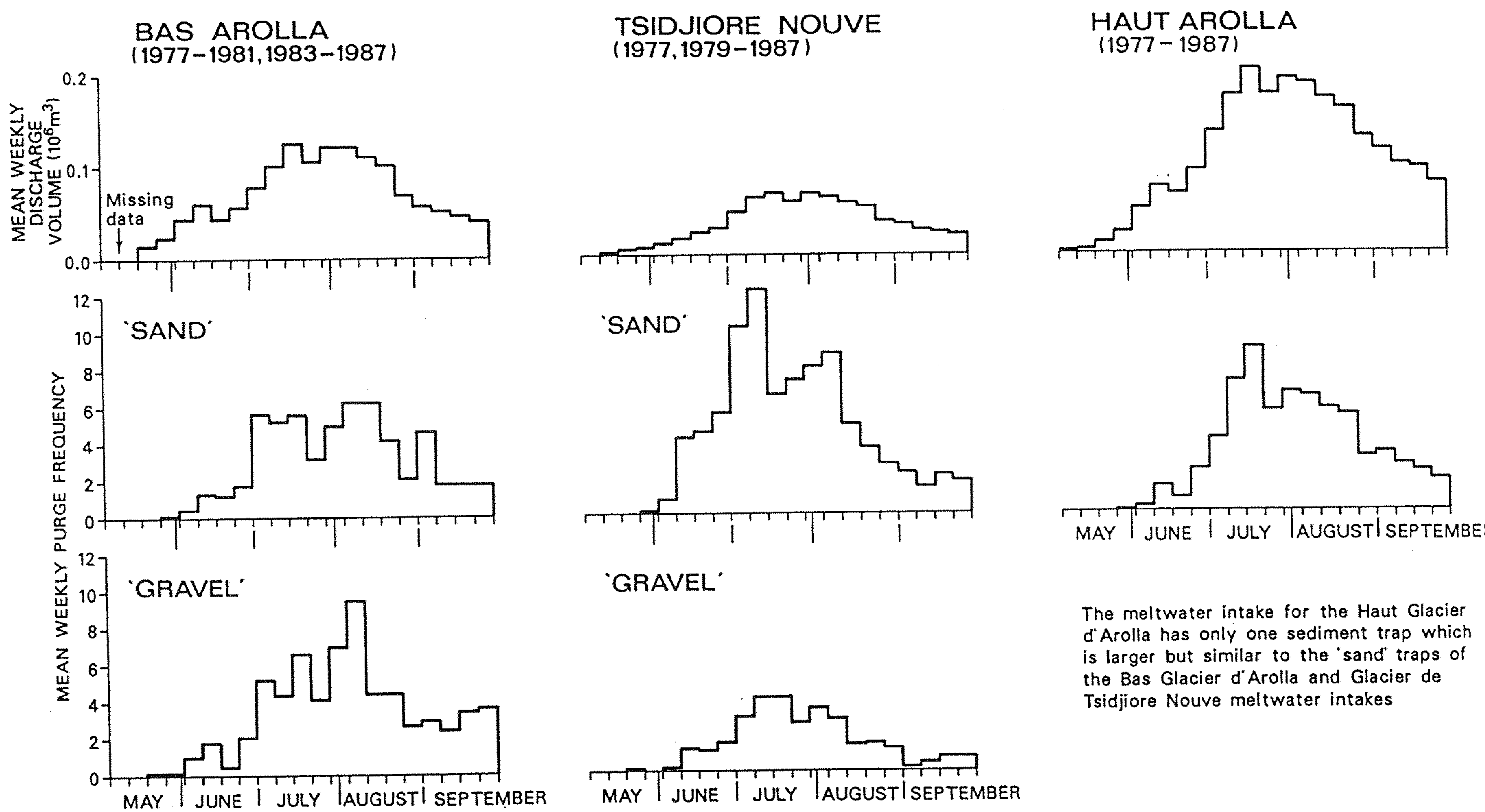


Figure 3.13 Mean weekly meltwater discharge and purge frequency, Haut Arolla, Bas Arolla and Tsidjiore Nouve basins 1986 (From: Bezinge, Clark, Gurnell and Warburton, 1988).

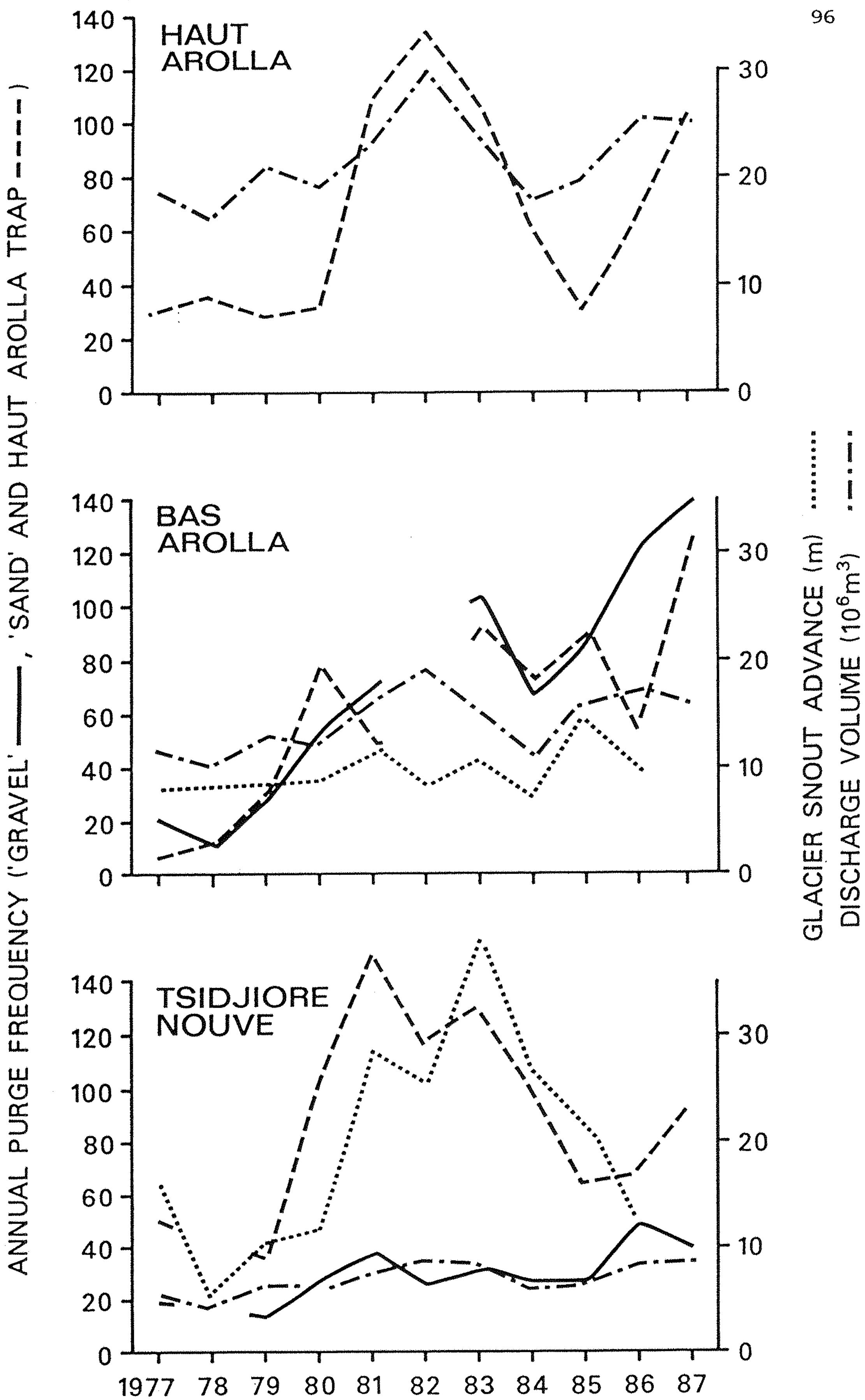


Figure 3.14 Annual purge frequency in relation to annual discharge volume and glacier snout advance, Haut Arolla, Bas Arolla and Tsidjiore Nouve basins (From: Bezinge, Clark, Gurnell and Warburton, 1988).

increase followed by a decrease in sand trap purges and glacier snout advance. Gravel trap purges and discharge show an increasing trend. For the Haut Arolla basin no glacier snout movement observations are available, but variations in discharge seem to correspond positively with purge frequency variations.

The calibration of sediment trap purges discussed in Section 3.2.2 permits the estimation of weekly sediment yields. Figure 3.15 shows the results of this analysis for the 1986 and 1987 seasons after standardising the loads for catchment area. The three basins studied in the 1986 season show similar trends in the pattern of purges. Haut Arolla has a reduced sediment output in comparison to the Bas Arolla and Tsidjiore Nouve basins. The contribution to the total clastic load (total load minus dissolved load) from sand trap purges at Tsidjiore Nouve and Bas Arolla is very low. Figure 3.15 shows a characteristic two peak pattern in the weekly sediment yield estimates for all of the basins in 1986 with an initial peak in July and a secondary peak in August. This pattern is repeated at Bas Arolla for the 1987 season.

Figure 3.16 shows the accumulated weekly transport rates of estimated suspended load and total load over the three month observation period for Tsidjiore Nouve 1986 and Bas Arolla 1986, 1987. The range in total load for the Tsidjiore Nouve basin results from uncertainty about the purge volume of bedload (Section 3.2.2). Malfunctioning of the turbidity meter at the Bas Arolla intake led to gaps in the estimated suspended sediment record which were filled using a regression relationship between purges of the sand trap and suspended sediment transport in the river. Estimates for the Haut Arolla catchment were not produced because the lack of field information on the division between suspended load and bedload from the single sediment trap. Uncertainty concerning sediment yield from the Bas Arolla catchment

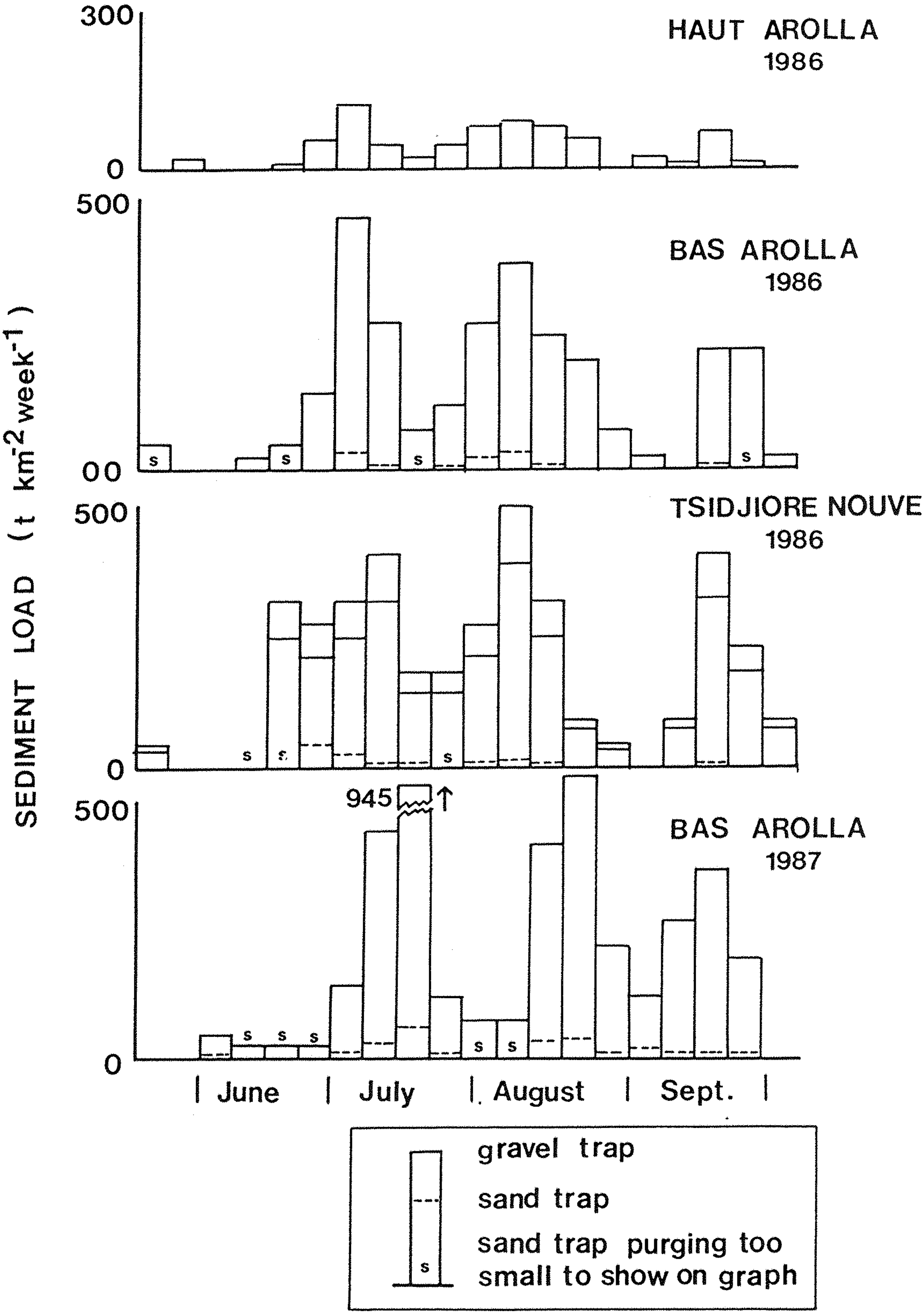


Figure 3.15 Weekly sediment yields Haut Arolla, Bas Arolla and Tsidjiore Nouve basins 1986; and Bas Arolla 1987.

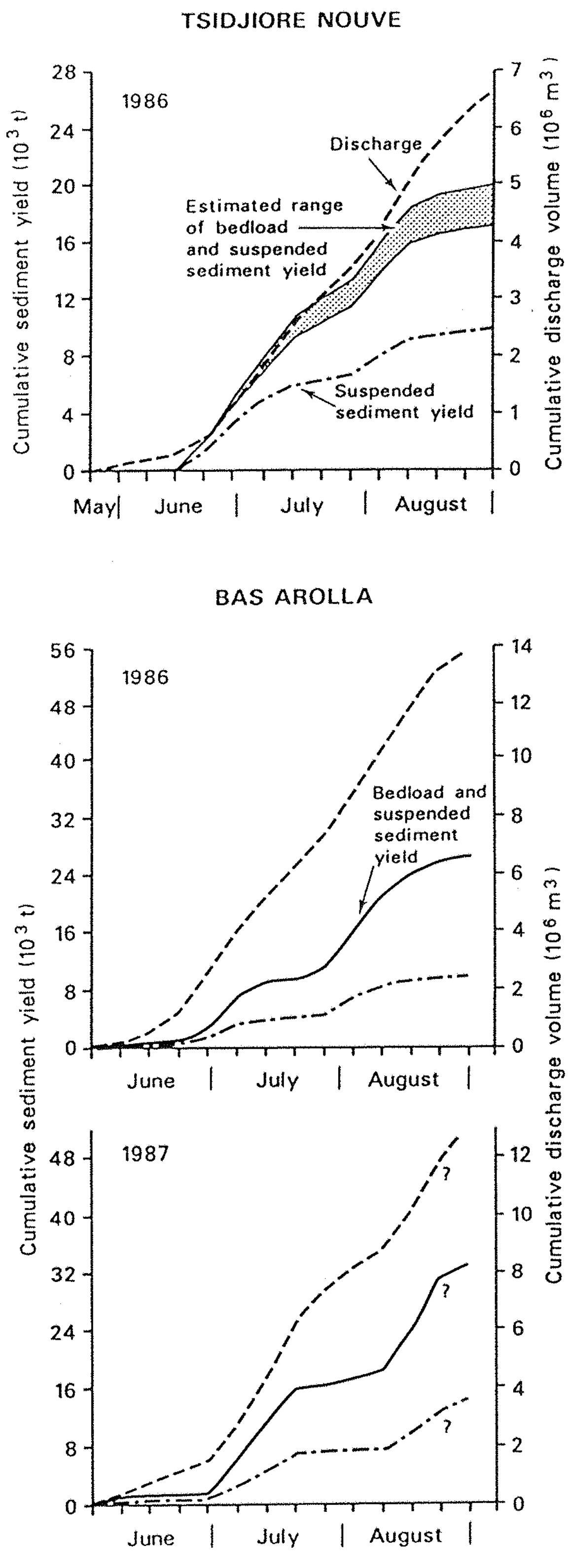


Figure 3.16 Accumulated weekly discharge, suspended sediment and bedload yield from Tsidjiore Nouve and Bas Arolla basins 1986; and from Bas Arolla basin 1987 (From: Gurnell, Warburton and Clark, 1988).

at the end of the 1987 observation period was caused by a large flood event on August 24th when prolonged and continuous purging of the intake structure was necessary.

Variations in daily discharge and suspended sediment yield (estimated from turbidity records and standardised for catchment area) are shown in Figure 3.17. In general water yields per unit area are higher from the Bas Arolla and Haut Arolla basins but sediment yields are higher from the Tsidjiore Nouve basin. The 'spikeness' of the Bas Arolla suspended sediment concentration series is notable and probably reflects flushes of suspended sediment associated with purging of upper catchment intakes or flushing of sediment caused by augmented flood peaks routed through the Bas Arolla catchment.

The pattern of sediment yield on the Tsidjiore Nouve site is unaffected by purging of high level meltwater intakes. High discharge events early in the ablation season transport large concentrations of sediment but as sediment sources are exhausted, less sediment is available for transport. This exhaustion effect is not observed at Bas Arolla. Indeed later season concentrations seem to be higher in relation to discharge. This phenomenon may be related to inputs of sediment from higher level glaciers within the natural catchment area. As a result of the higher altitudes of these basins, the commencement of the ablation season is delayed, and so sediment transport is also delayed. This is reflected in the sediment series for the Haut Arolla basin (Figure 3.15) which shows considerable sediment discharge during late July early August. Therefore, contributions from higher basins to the sediment load of the Bas Arolla glacier have an effect latter in the season, whereas the Tsidjiore Nouve basin is not affected by inputs from higher-level glaciers.

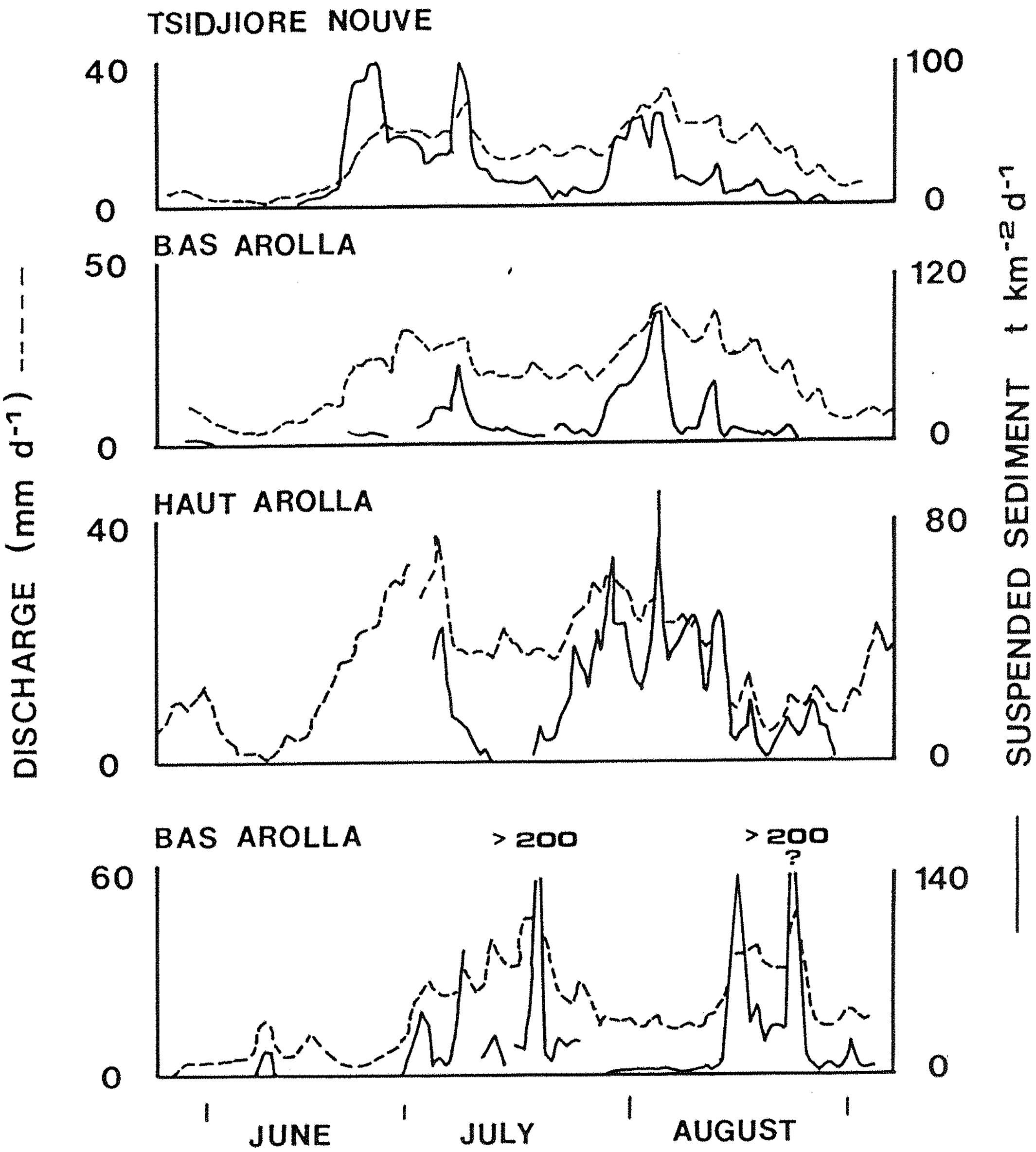


Figure 3.17 Variations in daily discharge and suspended sediment yield (standardised for catchment area) from Tsidjiore Nouve, Bas Arolla and Haut Arolla basins 1986; and Bas Arolla 1987 (Modified from: Gurnell, Warburton and Clark, 1988).

There are marked contrasts between the Bas Arolla and Tsidjiore Nouve glacier basins in both their discharge and sediment yield characteristics. The Tsidjiore Nouve proglacial stream shows high rates of sediment transport early in the ablation season followed by reduced transport latter in the season. Such differences are not as marked in the Bas Arolla proglacial stream. The Tsidjiore Nouve basin yields much more sediment in relation to its water yield than the Bas Arolla basin. Average figures for combined bedload and suspended load are approximately 2600 - 3000 mg l^{-1} for Tsidjiore Nouve and 1900 - 2500 mg l^{-1} from Bas Arolla and bedload forms a higher proportion of the total load from Bas Arolla (58 - 64 %) than from Tsidjiore Nouve (36 - 51%).

Differences in accumulated sediment yield also occur between years. The pattern of suspended sediment yield for 1986 and 1987 from the Bas Arolla basin (Figure 3.17) are initially the same with very low suspended sediment transport in June, followed by a rapid increase which levels out in July. In 1986 this is followed by another increase in suspended sediment yield which tapers off towards the middle of August. In 1987 this pattern does not occur due to a large flow in August which greatly increases sediment yield at that time.

In summary, the preliminary calibrations of the meltwater intake structures given here provide, important information on the sediment yield characteristics of these proglacial streams.

3.3.2 Sediment rating curves and sediment yield estimates

Table 3.2 summarises estimates of sediment load from the Tsidjiore Nouve, Bas Arolla and Haut Arolla basins for different time periods and for different components of the total sediment load in 1986 and 1987. The most

Table 3.2 Summary of SedimentOutput Results (tonnes)

	Bas Arolla			Tsidjiore Nouve	Haut Arolla
1986 Estimates based on sediment trap purges					
Gravel trap	22698			10248 - 13176	8840
Sand trap	832			739	
Turbidity [1]	7417 [2]			6079 [3]	19591 [4]
TOTAL	30947			17066 - 19994	28431
1986 Estimates based on sedimant rating curves [9]					
Bedload	8601	(33974)			
Suspended load	5362	(7721)			
Solutes	544	(549)	(508) [8]		
TOTAL		42244	42203		
1987 Estimates based on sediment trap purges					
Gravel trap	31210			7224 - 9288	13780
Sand trap	1487			1019	
Turbidity	19122 [5]			10440 [6]	
Conductivity	430 [7]	(386) [8]			
TOTAL	52249	52205		18683 - 20747	
1987 Estimates based on sediment rating curves [9]					
Bedload	10794	(42636)			
Suspended	9325	(14547)			
Solutes	540	(545)	(501) [8]		
TOTAL		57728	57684		

Notes

1. Calibrated to estimate sediment passing through the meltwater intake
2. Length of record 27th May - 11th September.
3. Length of record 24th May - 3rd September.
4. Length of record 5th July - 9th September.
5. Length of record 29th May - 4th September.
6. Length of record 5th June - 8th September.
7. Length of record 24th May - 26th August.
8. Solute load corrected for rainfall input.
9. Sample sizes of rating curve estimates given in Table 3.3.
10. All estimates are from 1st June until 30th September unless indicated otherwise.

comprehensive estimates are given for the Bas Arolla catchment. The purpose of this table is to present gross estimates of sediment yield for the three study basins. It is also of interest to compare these estimates with estimates derived from sediment rating curve techniques because these techniques are frequently used to estimate sediment yields in this type of stream (e.g. Hammer and Smith (1983)).

The sediment rating relationships used to estimate bedload, suspended load and total dissolved solids are detailed in Table 3.3. Samples are representative of the whole season and were weighed for bedload, filtered to determine suspended sediment concentration and evaporated to dryness to determine the amount of dissolved material in accordance with the methods outlined in Chapter 2. Sediment rating curves were estimated from log-transformed data by Ordinary Least Squares regression. Because log-transformation of the variables causes a systematic underestimation of sediment loads (Ferguson, 1986a) the appropriate correction factors were applied and the corrected loads are listed in Table 3.3. Estimates of sediment yield are for the entire ablation season. Two sediment rating relationships are presented for bedload which relate to the mean transport rate through the study section and the maximum transport rate within the section (measured using a Helley-Smith bedload sampler). The mean transport rate relationship is used for estimating bedload transport because of sediment streaming in the study section (Chapter 4). The high R^2 values associated with some of the rating relationships, particularly those for bedload, should be treated cautiously (Jansson, 1985) since previous attempts to define similar simple relationships have been less successful (e.g. Guymon, 1974)and have been primarily attributed to variations in transport rates resulting from fluctuations in sediment supply, although seasonal changes in catchment characteristics and changes in

Table 3.3 Ablation season rating curves for the Bas Arolla catchment: bedload on discharge for 1987; suspended sediment on discharge for 1986, 1987; and total dissolved solids (TDS) on discharge for1987.

	r^2	s.e	n	load correction factor
1986				
Suspended				
$\log_{10} C = 2.14 + 1.17 \log_{10} Q$	25.0	0.37	123	1.44
1987				
Bedload				
Average				
$\log_{10} C = -1.36 + 3.86 \log_{10} Q$	74.5	0.72	64	3.95
Maximum				
$\log_{10} C = -0.67 + 4.02 \log_{10} Q$	74.4	0.74	64	4.27
Suspended				
$\log_{10} C = 2.39 + 1.32 \log_{10} Q$	65.3	0.48	75	(a) 1.84
$\log_{10} C = 2.32 + 1.28 \log_{10} Q$	66.7	0.41	173	(b) 1.56
$\log_{10} C = 2.01 + 0.81 \log_{10} Q$	57.5	0.33	44	(c) 1.33
$\log_{10} C = 2.32 + 1.26 \log_{10} Q$	67.6	0.43	292	all 1.63
Total dissolved solids				
$\log_{10} C = 1.94 - 0.13 \log_{10} Q$	59.2	0.048	15	1.01

Notes:
n = sample size, a = intercept constant, b = slope, C = concentration (Bedload - kg s^{-1} of rated section; suspended - mg l^{-1}; total dissolved solids - mg l^{-1}) and Q = discharge in m^3 s^{-1}, r = correlation coefficient and s.e = standard error of estimate.
(a) Upper monitoring station, Bas Arolla proglacial stream
(b) Lower monitoring station, Bas Arolla gravel trap site A
(c) Bas Arolla site D.

runoff pathways (e.g. Walling, 1974; Collins, 1979) can also yield scatter in sediment concentration and discharge data, sediment exhaustion effects, hysteresis effects and the complex form (often multi-peak) of sedi-graphs (Walling, 1974; Collins, 1979).

A bedload rating established in 1987 for Bas Arolla and applied to the 1986 and 1987 discharge series provides load estimates which are in reasonable agreement with load estimates from both the gravel and sand traps considered together (1986 over-estimated by 44.4%, 1987 over-estimated by 30.4%). Overestimation is probably caused by sediment exhaustion effects, slight differences in the length of the observation periods due to gaps in the record and trapping a fraction of the suspended load during bedload sampling. The latter problem may lead to measurement duplication of a fraction of the suspended load. This might be overcome by simultaneously sampling bedload and suspended load together. Grain size curves could be then used to determine an appropriate correction factor. This is unlikely to be a simple relation since there is continuous interchange of particles between bedload and suspended load dependent on the prevailing streamflow and bed conditions.

Suspended sediment rating curves for the Bas Arolla approximate turbidity loads reasonably well 1986 slightly over-estimated by 4.1% and 1987 under-estimated by 34%). The largest discrepancy is for 1987 where loads are underestimated. This may again reflect the role of large events flushing-out large amounts of sediment independent of discharge. Under these circumstances sediment concentration will be poorly correlated with discharge.

The combined estimates from sediment ratings for Bas Arolla in 1986 significantly overpredict the load by 37%. In 1987 the ratings provided a better estimate of the total, within 10%, but this is due to the balancing effect of overestimating bedload and underestimating suspended load.

Comparisons between direct field measurements and indirect estimates based on sediment rating curves are important in assessing the usefulness of the sediment rating curve approach and to provide data which is comparable to the bulk of data derived in previous glacier basin sediment yield studies. In some circumstances the calculation of sediment load is fraught with difficulties (Carling, 1983). Clearly, there is not always a unique relationship between discharge and suspended sediment concentration, since relationships usually have high scatter and so the error term associated with the derived regression relationship is usually large (Grimshaw and Lewin, 1980). In the proglacial context the scatter associated with rating relationship is thought to be caused by: seasonal variations in sediment availability from the glacier; hysteresis effects; transient flushes of sediment from glacial sources without corresponding discharge variations; variations in sediment supply from local proglacial sources; and sediment dynamics associated with rainfall induced flow events (Fenn et al., 1985). In addition, Collins (1979) is particularly critical of using rating relationships for solute load estimation. Under these circumstances separate curves for sub-periods should be devised (Loughran, 1976), a technique which has been used for glacierised basins (Østrem, 1975), or more sophisticated multivariate regressions (Richards, 1984) or time series analytical approaches (Gurnell and Fenn, 1984) should be adopted.

3.3.3 Summary of sediment yield characteristics from the three study basins in 1986 and 1987

This section summarises the sediment yield characteristics for the three study basins according to the various estimates described above. Outputs of coarse sediment from gravel and sand traps varied markedly between the three basins in 1986 and between the two years 1986 and 1987 at Bas Arolla. In 1986 the Bas Arolla catchment produced approximately twice the weight of sediment yield of the Tsidjiore Nouve and Haut Arolla basins. These differences can be explained, to some extent, by the abundance of gravel sediment sources in the proglacial zone of the Bas Arolla glacier. At Tsidjiore Nouve the extremely coarse boulder bed, the lack of sediment supply from proglacial sources, lower discharges and the step pool structure of the stream inhibit active gravel transport. Haut Arolla also shows a much reduced coarse sediment load which probably reflects the greater storage potential of the low gradient and relatively long (1.0 km) proglacial area immediately upstream of the trap. The influence of such proglacial storage zones has been identified by Borland (1961). However the design of Haut Arolla trap also means that at very high discharge a substantial number of cobbles are transported over the water intake grid of the sediment trap and so are not collected in the trap.

In 1987 a similar pattern emerges, coarse sediment output is higher at Bas Arolla and Haut Arolla but lower at Tsidjiore Nouve (Table 3.2). During 1986 there was a greater incidence of rainfall events with large amounts of coarse material probably being contributed from slope and bank sources at Bas Arolla and Haut Arolla. More frequent sand trap purges at Tsidjiore Nouve during this period also suggest additional supplies of sediment to the proglacial stream, possibly from proglacial sources.

Using sediment trap calibrations bedload at Bas Arolla was estimated to be 64% for 1986 and 58% in 1987 whereas Tsidjiore Nouve estimates were between 43-51% in 1986 and 36-44% in 1987 (uncertainty being related to variations in the estimates of the volume of material accumulated in the gravel trap immediately prior to purging). Nevertheless these estimates are based on the assumption that the ratio of bedload and suspended load accumulated in the various traps is constant throughout the ablation season. It does not account for possible changes during the ablation season and for variations in the antecedent and present hydrological and hydraulic conditions.

Turbidity data for 1986 show the Haut Arolla basin to have the highest suspended sediment yields, much greater than the similar yields from Bas Arolla and Tsidjiore Nouve. This is true in spite of the fact that the observation period at Haut Arolla is of shorter duration than that for the other two basins. In 1987 differences between Bas Arolla and Tsidjiore Nouve are more marked which probably reflects the importance of large scale events contributing disproportionally to the sediment load. Østrem et al., (1967) for the Decade Glacier, Baffin Island suggests that 60% of silt transport, during a period of 66 days, occurred in one day.

In calculating the solute loads for the Bas Arolla (Table 3.2) two methods were used: a discharge - total dissolved solids rating curve (Section 3.2.3; Table 3.3) and a continuous record of conductivity calibrated for total dissolved solids. It was found that the rating curve overestimated loads by 29.8%. All totals were corrected for rainfall inputs using a mean value of rainfall total dissolved solids of 25 mg l^{-1}. The best estimate of solute load derived from geochemical weathering is given by the total load minus rainfall inputs plus or minus biological effects, (biological effects can be assumed to be unimportant in this environment, Lemmens and Roger, 1978). Bas Arolla loads, based on dissolved load rating curves and standardised for catchment area, are approximately equal in the two study years (66.8 t for 1986 and 65.9 t in 1987). The estimate based on continuous conductivity records for 1987 is somewhat smaller, 53.6 t, but this probably reflects the shorter measurement period. Values are slightly above the global average of 38.8 t suggested by Walling and Webb (1987). This is not surprising since previous studies have indicated the importance of glacierised basin for solute yields. Eyles et al. (1982), for example, demonstrated chemical weathering to be three times the world denudation rate in the Berendon Glacier basin, British Columbia. The 1986 estimate of solute load is 1.2% of the total load and the 1987 estimates 0.9% (rating curve estimate) and 0.7% (continuous conductivity record estimate). Therefore, the solute load estimates represent a small proportion of the total load and are approximately equal for the two years.

Total clastic sediment load estimates for 1986 show the Bas Arolla basin to produce the greatest absolute volume of sediment with Tsidjiore Nouve and Haut Arolla approximately equal. In 1987 Bas Arolla yielded much greater total load than the Tsidjiore Nouve basin. This difference is thought to be due to the importance of large events, both natural and artificial, which occurred in the Bas Arolla basin during this period.

The data in Table 3.2 should be regarded as minimum estimates of sediment transport since the measurements do not extend over the entire ablation season. The determination of total sediment yield and its components is fraught with difficulty. Sediment rating curves have been estimated to provide a basis for valid comparison with other studies and to extend load estimates beyond the limits of the observation period. In spite of the difficulties of trap calibration it is felt that the data from this source combined with calibration curves and continuous monitoring of turbidity and electrical conductivity provide the most reliable estimates of sediment yield.

3.4 Comparison of purge sediment yields from glacierised basins in Val d'Hérens, Switzerland.

As shown in Table 3.2 sediment yields from the three study basins show quite substantial differences. The purpose of this section is to compare, or at least discuss, the nature of these loads in comparison with of other glacier basins. To this end useful preliminary summaries of sediment yield data are provided by Gurnell (1987); Walling and Kleo (1979); Walling and Webb (1987); and Jansson (1988). However the later review articles (Walling and Webb, 1987; Jansson, 1988) are largely inappropriate to the present discussion since they are only concerned with suspended sediment yields and large basins (greater than 350 km^2). Catchment area is important because there is a small but definite scale dependence of sediment load per unit area as a results of increased depositional opportunities in larger basins (Ferguson, 1984). Nevertheless, glacierised basin suspended sediment yields are generally above the global average when standardised for catchment area, in spite of the temporally restricted runoff period (Gurnell, 1987).

Firstly it is appropriate that the three study basins should be set in the context of other glacierised catchments in the Val d'Hérens (Bezinge, Clark, Gurnell and Warburton, 1988). Table 3.4 summarises average sediment yield data for 18 basins over a 5 year period. Yields are calculated on the basis of the design volume and an average packing density of 1.45 t m^{-3}. Results reflect the very large differences in yield but several limitations must be borne in mind:

a) Manual traps are purged infrequently when the trap is visited. Although this generally occurs when the trap is near capacity, purging before the trap is full can lead to over-estimation of sediment yield (or

Table 3.4 Provisional estimates of mean sediment yield from basins in the Val d'Herens, Switzerland: based on purging of water intake sediment traps (Developed from: Bezinge, Clark, Gurnell and Warburton 1988).

WATER INTAKE / CATCHMENT	CATCHMENT AREA km² (A)	GLACIERISED AREA km² (B)	TOTAL SEDIMENT YIELD t	SEDIMENT YIELD t. km² (A)	(B)
Les Rosses	2.34	0.2	290	124	1450
Bricola	4.05	1.32	203	50	154
Dent Blanche	2.54	1.51	145	57	96
Rocs Rouge	1.61	0.73	218	135	299
Manzettes	2.24	1.61	297	133	184
Ferpecle	32.29	21.46	145	5	7
Mourti	0.78	0.15	44	56	294
Piece	3.27	2.14	247	76	115
Douves Blanche	1.22	0.34	174	143	580
Bertol Sup.	2.29	0.63	218	95	346
Haut Arolla	11.7	6.3	8410	718	1335
Vuibe	1.41	1.07	290	206	271
Ignes	2.95	1.39	218	74	157
Aiguilles Rouge	2.89	1.33	464	161	349
Bas Arolla	7.6 (25.1)	5.32	23983	3156	4509
Tsidjiore Nouve	4.8	3.41	6192	1290	1816
Vouasson	3.18	2.36	870	274	369
Fontainesses	4.97	0.0	negligible yield		
Total		approximately	43500		

All yields in metric tonnes based on an average volume weight conversion of 1 : 1.45

under-estimation if the trap is over full).

b) Some manual purging is undertaken during maintenance of automatic traps but generally this can be identified in the water stage record.

c) Packing densities depend on the variable calibre of the trapped sediment.

d) Marked discrepancies may exist between the design volume of the trap and the actual volume of sediment accumulated at purge (as is the case at Tsidjiore Nouve).

e) Trap efficiencies will vary not only between different styles of trap but with the same type of trap at different sites and even with the same trap over time. For example there is overspill of bedload at the Haut Arolla trap and overspill of suspended sediment charged water at Bas Arolla when discharge is high. This is more likely at manual traps where purging is less frequent.

f) Flushing efficiency may not be 100% (i.e. some residual amount of sediment is left in the trap following purge, if this is not accounted for the actual yield will be over-estimated.

g) The siting of the trap also has important implications since natural intervening sediment stores in the proglacial zone may have a profound effect on yield. This is graphically illustrated in Table 3.4 since Ferpecle with the greatest catchment and glaciated area has an extremely low yield. This is because the sediment trap is sited at a considerable distance away from the snout and a large proglacial lake acts as a sediment store in the intervening distance. Mathews (1964) in his study of sediment yield from the Athabasca glacier, Canada, demonstrated that a proglacial lake trapped 99% of the total load, all of the bedload and all of the sand fraction of the suspended load.

h) A large proportion of the load may be transported in suspension right through the sediment trap This explains some of the differences between load estimates presented in this study (Table 3.2) and estimated in Table 3.4 for the three study basins.

Given these limitations comparisons seem difficult even on this relatively small local scale. This emphasises the need for detailed calibration of sediment traps coupled with a detailed knowledge of the catchments being studied. However, one final point to be made about Table 3.4 is that the three study basins considered here are the three basins with the highest sediment yield within the Val d'Hérens (approximately 86 % of the 'measured' sediment output). The Bas Arolla basin accounts for 53 % of this value whereas Haut Arolla and Tsidjiore Nouve account for 19 % and 14 % respectively. These general trends are in agreement with the conclusions drawn from the detailed estimation of loads for these three basins in 1986.

3.5 Sediment yields from glacierized, mountainous and upland basins

At a somewhat larger scale Table 3.5 attempts to assess how glacier basin sediment yields derived in this study compare with other glacierized, mountainous and upland catchments (many of which have been previously glaciated). Firstly, data are presented without standardisation. The most common form is to report loads in tons Km^{2} yr^{-1}, but where this is not the case attempts to standardise estimates are not recommended because the history of the data are unknown. If standardisation is attempted an important question is whether yield data should be standardised for catchment area or glacier area? Clearly estimates will vary greatly depending on which convention is used. The Norwegians (e.g. Kjeldsen and Østrem, 1980) tend to use glacier area for presenting specific sediment yields and this can be justified by the assumption that most of the sediment is derived from the products of glacier erosion. This makes an interesting assumption regarding the role of the proglacial zone since it suggests little sediment is eroded or deposited in this area. Some of the data is also derived from sediment rating curves which have not been corrected for logarithmic transformation bias (e.g. Hammer and Smith (1983)). Furthermore estimates of bedload are based on a variety of sources including purging of sediment traps (Gurnell et al., 1988; this study); fence traps and delta surveys (Kjeldsen and Østrem, 1980); bulk accumulation traps averaged over many years (Newson and Leeks, 1985); sediment rating curves (Hammer and Smith, 1982); and bedload transport equations (Church, 1972; Hasholt, 1976; Lewin, 1977). The proliferation of non-standardised methods is caused by each study being approached as a separate problem and the method of solving it (the technique of measurement) being different in each case. There is a real need for appropriate and proven equipment for monitoring sediment dynamics in mountain streams, since established sediment transport equations significantly over-estimate rates and sediment rating curves are of limited use given variations in transport rate related to fluctuations in sediment supply (Bathurst et al, 1986a). Because of measurement errors and differences in the proportions of the load being measured comparisons are fraught with difficulties. The net effect of these inconsistencies is load estimates which can only be regarded very generally (Walling and Webb, 1983). Solutes load are also prone to error because unless samples are prefiltered before storage, estimates can be greatly in error (Eyles et al., 1982) which makes many estimates incompatible. However in general solute data tend to be more reliable than those for suspended load which may be in error of 60% for annual loads (Walling, 1978).

Total yields vary greatly within glacier basins and even between years. This is clearly illustrated in the studies of Hammer and Smith (1983) and Beecroft (In Gurnell, 1987). Often a large proportion of the load is transported in a single event (Beecroft, 1983). Kjeldsen and Østrem (1980) estimate between 20-50% of annual sediment transport is during the first flash flood event of the season. If this does not occur, sediment will accumulate in storage until the following ablation season. This explains to some extent the large variations in annual sediment production and suggests the need for long runs of sediment yield data if annual sediment totals are to be interpreted meaningfully and storage effects averaged out (Phillips, 1986). This may be possible in Norway where sediment yields have been studied since 1967 (Kjeldsen and Østrem, 1980) and in Switzerland where purging data from sediment traps is available (Gurnell, Warburton and Clark, 1988). Proglacial streamloads tend to be at least one order of magnitude greater than temperate, non-glacierised mountainous basins, even when standardised for catchment area.

STUDY	RIVER / AREA	YEAR	BASIN AREA Km2	BEDLOAD	SUSPENDED	DISSOLVED	TOTAL	BEDLOAD: SUSPENDED LOAD RATIO
PROGLACIAL-GLACIERIZED								
Mathews (1964)	Athabasca, Alberta	1957	28.4 (13.5)		778 (37) b	1311 (63) b	2089 b	
Church (1972)	South River, Baffin Island,	1968						0.27
Church and Gilbert (1975)	South River, Baffin Island,	1967		570 (90) a	(10)			9
	Lewis River, Baffin	1963		1049 (82) a	(18)			4.5
		1964		236 (77) a	(23)			3.5
Ostrem (1975)	Norwegian meltwater streams			(25)				
Hasholt (1976)	Sermilikarea, East Greenland,	1972	38 (30)	(5-6)	(72-78)	(17-22)		0.07
Kjeldsen and Ostrem (1980)	Engabreen, Norway,	1979	50 (38)	7300 (37) b	12200 (63) b			0.58
		1980		8600 (36) b	15500 (64) b			0.56
	Nigardsbreen, Norway	1979	65 (48)	14000 (43) b	18400 (57) b			0.75
		1980		5200 (30) b	11900 (70) b			0.42
Shcheglova and Chizhov (1981)	Lyanga, Asia				323			
	Issykata, Asia				463			
Hammer and Smith (1983)	Hilda, Alberta, Canada	1977	2.24	784 (57) b	548 (39.9) b	42 (3) b	1374 b	1.43
		1978		981 (54) b	808 (44.5) b	28 (1.5) b	1817 b	1.21
		1978		905 b				
Beecroft (1983)	Tsidjiore Nouve, Switzerland	1981	4.8 (3.4)	3840 (70) b	1674 (30) b		5514 b	2.29
Beecroft (In Gurnell, 1987)	Tsidjiore Nouve, Switzerland	1981	4.8 (3.4)	46400 (40) b	9500 (60) b		15900 b	0.66
		1982		4300 (33) b	8700 (67) b		13000 b	0.49
THIS STUDY	Bas Arolla	1986	5.31(70.2)	23530 (74) b	7417 (24) b	508 (2) b	31455 b	3.17
		1987		32697 (63) b	19122 (36) b	386 (1) b	52205 b	1.71
	Tsidjiore Nouve	1986	3.4 (70.6)	13915 (70) b	6079 (30) b		19994 b	2.28
	Haut Arolla	1986	6.33(54.0)	8840 (31) b	19591 (69) b		28431 b	0.45
MOUNTAINOUS AND UPLAND STREAMS								4.0
Lewin et al (1974)	Maesant, Wales		0.54	1.1 c				
Lewin (1977)	Maesant, Wales	1971-72	0.54	0.93 (20) a	3.72 (80) a		4.65 a	
	Ystwyth, Wales	1973-74	170	255.5 (90) a	28.4 (10) a		283.9 a	
Lewin and Wolfenden (1978)	Iago, Wales			1.2 c				
Newson (1981)	Peny Banc, Wales			9.9 c				
Carling (1983)	Carl beck, Pennines		2.18	0.33 (0.2)a	24.7 (18.3) a	110 (81) a		
	Eggleshope beck		11.68	0.55 (4.4)a	12.07(95.6) a	0.0 (0) a		
Emmett (1984)	Tanana, Snake and Clearwater, USA			(>10)				
Ferguson (1984)	Hunza, Karakoram		13200	(2-10)	(88-96)	(2)		0.02
Richards and McCaig (1985)	Allt a Mhuillin, Scotland		6.19	26 a				
Newson and Leeks (1985)	English Lake District mountain streams		0.5-6.0	103-144 a				
				29-77 a				
Stott et al (1986)	Monachyle, Scotland	1982-85	7.7	0.12 c				0.003
Bathurst et al (1986)	Roaring River, Colorado	1984-85	33	(5-48)				0.05-0.9
GENERAL								
Lane and Borland (1951)	General estimate			(2-26)				
McCann and Cogley (1973)	Mecham, Canada			21.1 a		30-40 a		

Notes: Basin area - ( ) values = glacier area
Load components - ( ) = % of total load
a = t Km-2 yr-1
b = t yr-1
c = m3 km-2 yr-1

Table 3.5 Sediment load measurements in small catchment proglacial and steep, coarse bed streams.

Fluvially transported material is often divided into bedload, suspended and solute components. These divisions are largely arbitrary and should be treated with caution since the desire to pigeon-hole process can lead to bias in the results. Furthermore because sediment is studied in the relative quiescence of the trap environment the load components are divided in a phase of deposition but conditions of deposition greatly differ from those of active transport and entrainment. Load components, measured in the stream channel, vary erratically, since sediment transport in proglacial streams is closely related to local conditions. Clearly such limitations are more related to the bedload - suspended load transition rather than the dissolved load - fine particulate boundary.

In temperate rivers the dominance of suspended load over bedload has been frequently reported (Gregory and Walling, 1973; Duijsings, 1985) so much so that bedload is often excluded from measurement programs. However, Alpine proglacial streams are steep, highly active, coarse-bed channels. The combination of variable flows, abundant sediment sources and easily erodible non-cohesive banks provide a high potential for bedload transport (Church and Gilbert, 1975; Hammer and Smith, 1983). This is verified by the data in Table 3.5. These data emphasise the wide variability in bedload transport potential. Church (1972) and Church and Gilbert (1975) with data from the Lewis River, Baffin Island (1963, 1964) demonstrate the large variability in annual yields and the proportion of the total load. In addition the length of observations also has an effect. For example the bedload contribution to total load measured by Beecroft (1983, and In Gurnell, 1987) for Tsidjiore Nouve, Switzerland in the 1981 melt season show a considerable difference for a 4 day outburst event (2.29 : 1, Bedload : Suspended) and for the entire season, May - September (0.66 : 1). The other main factor brought out by Table 3.5 is that proglacial streams appear to transport more bedload than coarse-bed mountain and piedmont streams. This contradicts some authors (e.g. Lewin and Weir, 1977) who have suggested some similarities between the two environments. In reality both types of stream show a wide variability in the amount of sediment transported (e.g. Alpine streams in the Colorado Front Range transport hardly any coarse clastic sediment (Caine, 1974)). Research into bedload movement is in general under investigated (Newson, 1981) but there have been even fewer studies in proglacial streams (Gurnell, 1987; Gomez, 1987).

Differences in the proportion of suspended load in the total load vary between 10 and 78% for proglacial streams and between 10 and 96% for mountain catchments. Absolute yields tend to be high but not as high as yields from some of the mountainous basins of the world with glaciers in their basins heads (e.g. Shcheglova and Chizhov (1981) quote a specific suspended sediment yield for the Vakhsh basin in Pamir of 14900 t km^{2} yr^{-1}). In fact as distance from the glacial source increases (as basin area increases) the importance of the suspended load component increases (Ferguson, 1984).

Measurements of dissolved load are few because this load poses little problem for river engineering (Walling, 1984). However this should not be ignored since solute load from small glacierised basins represent disproportionately high contributions to the load of the river (Eyles et al., 1982). Furthermore solutes interact with the suspended load (Richards, 1982) and may be transported along with the suspended load. Dissolved loads in glacier basins tend to be above the global average but the fraction of the total load is small (section 3.5). Annual loads have been characterised in terms of mequiv.m^{2} yr^{-1}. These chemical yields are comparable for the Gornergletscher - 454 mequiv.m^{2} yr^{-1} (Collins, 1983), Tsidjiore Nouve

- 508 $mequiv.m^{2}$ yr^{-1} (Souchez and Lemmens, 1987), S. Cascade (USA) - 960 $mequiv.m^{2}$ yr^{-1} (Reynolds and Johnson, 1972) and Berendon (Canada) 947 $mequiv.m^{2}$ yr-1 (Eyles et al., 1982). Values are above the global mean of 390 $mequiv.m^{2}$ yr^{-1}. It is generally accepted that solute loads are an important load component in Arctic basins (McCann and Cogley, 1973), previously glaciated catchments (Rapp, 1960) and in basins of calcareous bedrock (Carling, 1983b) as well as Alpine glacier basins.

In summary, glacierised basins appear to have high sediment and solute yields, with clastic sediments being responsible for the major load components of which bedload form a very significant part. Further comparisons of the data are difficult because yield data are not clearly documented and methods used to determine sediment movement differ widely. This lack of a standardised methodology is clearly demonstrated in Table 3.5 and largely restricts the opportunities to draw more detailed conclusions.

3.6 Summary

Sediment traps located in meltwater intake structures offer a valuable opportunity to estimate sediment yields. Once surveyed and the material properties characterised the purge records from these traps can be converted into meaningful estimates of load. Combined with continuous monitoring of suspended sediment concentration and electrical conductivity a total load can be estimated (Section 3.2).

Accepting the preliminary calibration of meltwater intake sediment traps, the sediment load of a proglacial stream can be characterised. Results illustrate differences in total loads between basins and the nature of the load transported.

Comparison of the three study basins suggests that Bas Arolla has the greatest sediment yield, second is the Haut Arolla basin and the smallest yields are from the smallest Tsidjiore basin Nouve. When yields are standardised for catchment area Tsidjiore Nouve yields 4174 t km^{-2}, Bas Arolla 4093 t km^{-2} and Haut Arolla has the smallest yield of 2422t km^{-2}. Calibration of the sediment traps allows estimates of load components at Bas Arolla and Tsidjiore Nouve: Bas Arolla carries a higher proportion of bedload (58-64 %, 1986-87 estimates) than Tsidjiore Nouve (36-51%, 1986-87 estimates). Similar estimates are not available for Haut Arolla.

Despite some sediment exhaustion later in the season, Tsidjiore Nouve sediment yield is more closely related to discharge than that from Bas Arolla, which has more severe supply limitations except in the late season when sediment supply is augmented from upper catchment sediment stores of the higher level glaciers. Delay in sediment release is clearly demonstrated from the sediment series of the Haut Arolla.

Solute yields appear to be above the global average but when expressed as a fraction of the total load are only a small proportion (at Bas Arolla about 1%).

Contrasts in sediment yield can be caused by a multiplicity of factors related to conditions in the proglacial zone, within the catchment and at the glacier bed. Especially important is the recent mass balance of the glacier (Kamb et al., 1985; Humphrey et al., 1986) and the accessibility to sediment sources in the immediate glacier foreland (Gurnell et al., 1988). The role of the proglacial zone as a source of sediment and as a store (sediment regulator) will be examined in the following chapters.

Compared with regional estimates, the three study basins account for 86 % of the measured sediment output from the Val d'Hérens. In a global context glacierized basins have high sediment and solute loads with clastic sediment accounting for the major part of this. Attempts to standardise these estimates are fraught with difficulty due to uncertainties surrounding the derivation of the sediment yield data. For example, it would be pointless to compare sediment yields between sediment rating curve estimates, when all estimates are not corrected for logarithmic transformation bias (Ferguson, 1986).

Chapter 4.

SEDIMENT INPUT - OUTPUT FROM THE PROGLACIAL ZONE

4.1 Introduction

Channel processes control the balance between sediment transport, deposition and erosion (Petts and Foster, 1987) in the valley train. Therefore, recognition of the channel as a sediment store (Coldwell, 1957; Kelsey et al., 1986) is fundamental in understanding the conveyance of sediment through river systems (Walling, 1988). The degree to which channel processes control storage depends on the interaction of various channel hydraulic, geometric and sedimentary characteristics with the streamload and discharge regime. This effect (i.e. sediment delivery, Roehl (1962)) varies considerably: Trimble (1981) documented large increases in storage following urbanisation of Coon Creek, Wisconsin; whereas Lambert and Walling (1986) working on the River Exe catchment, Devon demonstrate storage is of minimal importance and suggest that the channel acts as an efficient conveyer of suspended sediment. Clearly the timescale over which storage changes are identified is important since losses and gains in storage occur over storm periods (Meade et al., 1981), over many years (Kelsey et al., 1986) and over decades (Trimble, 1981). In smaller basins Duijsings (1985) suggests sediment delivery can be 100% at a seasonal scale but varies markedly, due to channel storage, with season.

Few studies have attempted to investigate changes in sediment load over a reach in mountain streams: Kelsey et al.(1986) and Bathurst et al.(1986a) being notable exceptions. The approach adopted in this chapter is to determine variations in suspended and bedload storage over the Bas Arolla proglacial channel using both direct and indirect methods. Direct methods are applied to the suspended load where a direct comparison of sediment loads entering and leaving the reach is made by referring to observations from suspended sediment monitoring stations at the upper and lower ends of the reach. This is appropriate because the transport of suspended load is quasi-continuous. For bedload, where transport is inherently discontinuous (Lewin et al., 1988), indirect methods are used. In this case a paired station approach was inapplicable to the Bas Arolla proglacial stream because of the inherent lack of continuity in the transport of bed material, the complex nature of the channel bed structure and the difficulty in finding a second site (at the proximal end of the reach) suitable for Helley - Smith bedload sampling. In addition, as bedload could not be measured continuously and rating curve methods did not offer the accuracy required for detailed investigations, due to the influence of variable bed conditions on entrainment thresholds (Laronne and Carson, 1976), a paired station approach was rejected. As an alternative, the intermittency and discontinuity in bedload transport was best studied by examining inter-relations between within-reach flow characteristics (discharge) and bed material mobility. The choice of different methods is justified on the grounds that bedload moves at velocities slower than the surrounding flow and suspended sediment, since grains move intermittently. This is a response to: fluctuating flow characteristics such as flow velocity at the bed, shear stress or stream power (Richards, 1982); variations in bed materials and bed structure; and limitations in sediment supply.

This Chapter examines the importance of channel erosion as a source of sediment for fluvial transport. Two load components are distinguished: suspended sediment (sediment supported by the vertical velocity component of turbulent eddies (Richards, 1982)) and bedload (that part of the stream sediment load which is transported by rolling , sliding or saltation (jumping) in close proximity to the bed (Abbott and Francis, 1977)). Suspended sediment includes washload which is fine sediment that remains in suspension even at low flow velocities, and the finer fraction of the bed material which is temporarily suspended during high flows. The distinction between bedload and suspended load is never clear-cut since it is highly variable both temporally and spatially.

This Chapter begins with a description of within channel sediment sources. This is followed by a two part discussion of sediment tranport: suspended sediment transport through the Bas Arolla proglacial stream channel, and bed material mobility in the reach.

4.2 Within-channel sources of sediment

Material arranged on the channel bed represents a source of sediment for sediment transport. The size range of the Bas Arolla channel bed materials ranged from silts up to boulders of a metre or more in diameter.

In coarse bed channels large samples are required to accurately characterise channel sediments (Church et al., 1987) Extensive systematic sampling of valley train sediments was not undertaken (e.g. Ballantyne, 1978) A random sampling design was used, which may be considered inappropriate in proglacial areas where downstream trends in sediment characteristics are notable (McDonald and Banerjee, 1971). However, given the short length of the Bas Arolla proglacial zone (0.3 km) it is unlikely that any downstream trends would be apparent since changes in downstream sediment characteristics in mountain streams show great variance (McPherson, 1971)

Four main types of sediment sample were collected:
1) Grab samples with from within-channel storage sites

on the channel margins (M), bar tops within the channel (BI), and bar tops adjacent to the channel (BO) (Figure 4.1, 12 samples).

2) Grab samples from the channel bed (Figure 4.2, 3 samples).

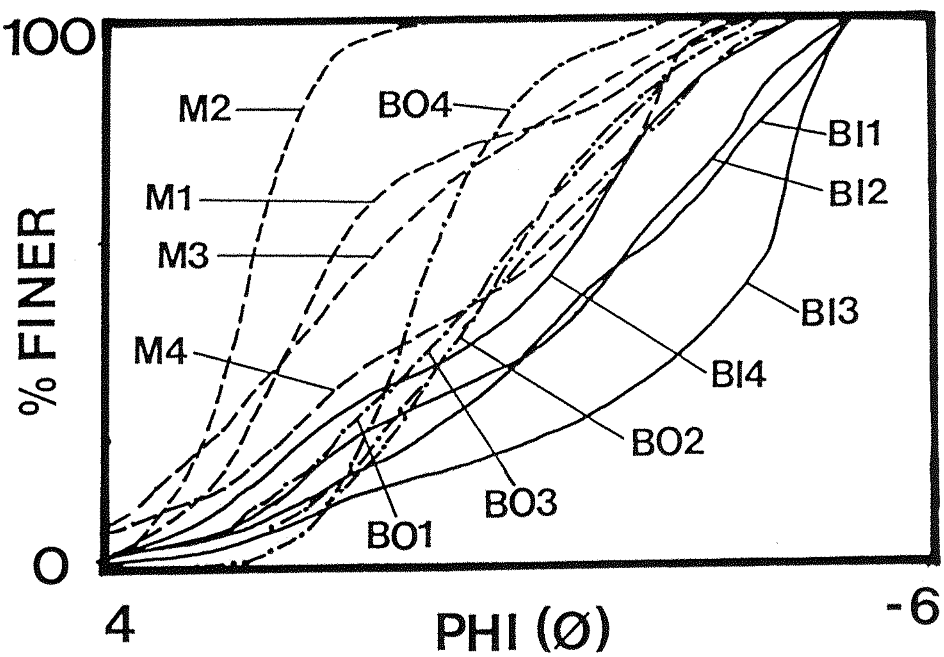
3) Paired surface and sub-surface samples at 9 different locations along the proglacial stream channel (Figure 4.3).

4) Grid sampling was used to determine surface particle sizes (Wolman, 1954) of bed sample sites and particle clusters - boulder steps.

Grab sampling of within-channel sediments (June 2nd 1987) provided an assessment of the sediment sizes available in the channel bed early in the season. The size distributions (Figure 4.1) of all 12 samples show considerable variety in bed sediment sizes. Generally speaking, the marginal samples (M) tend to be finer than the out-of-channel bar-top sediments which in turn are finer than the bar-top samples within the channel. This is to be expected since marginal deposits consist of fine material deposited in small pockets or shadows related to areas of quiescence in the flow (e.g. in the lee of boulders (Carling and Reader, 1982). Bar-top samples on the other hand are actively deposited often in an imbricated arrangement forming a coarse surface armour. The distinction between in and out of channel bar sediments is useful since out-of-channel bars, which constitute the bulk of the valley train deposits immediately adjacent to the channel, store a greater proportion of the fine material. The fine component in these deposits may be the result of overbank flows, filtering through from the sub-surface or deposition of fines from ablating snow. These deposits are a rich source of fine sediment available for transport during the initial melt-season flows.

When compared with bed material grab samples (Figure 4.2) the bed material was coarser than the sediments

Figure 4.1 Grain size distribution curves of within channel sediment samples collected from the Bas Arolla proglacial stream, June 2nd 1987.



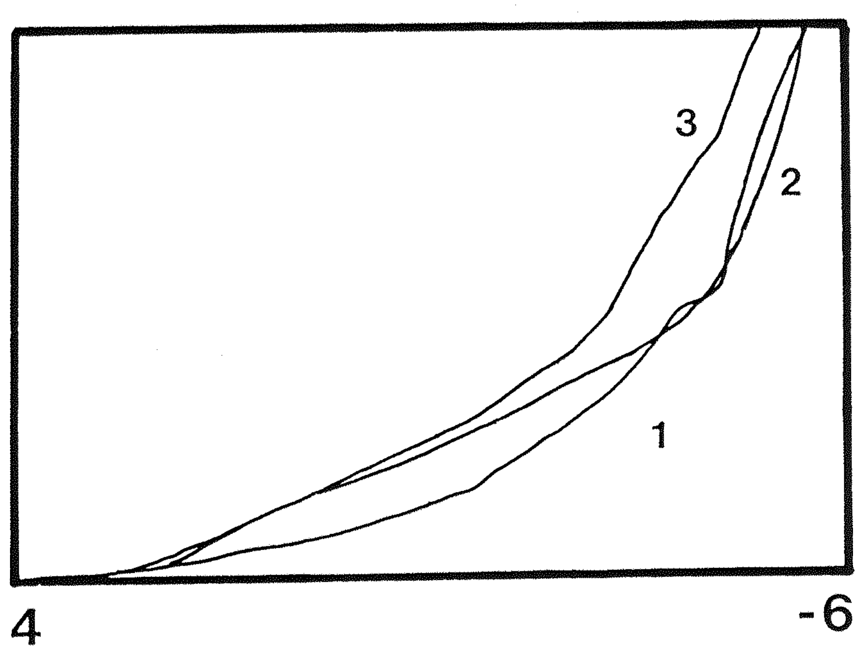


Figure 4.2 Grain size distribution curves of bed material samples collected from the main lower cross-section site, June 1987.

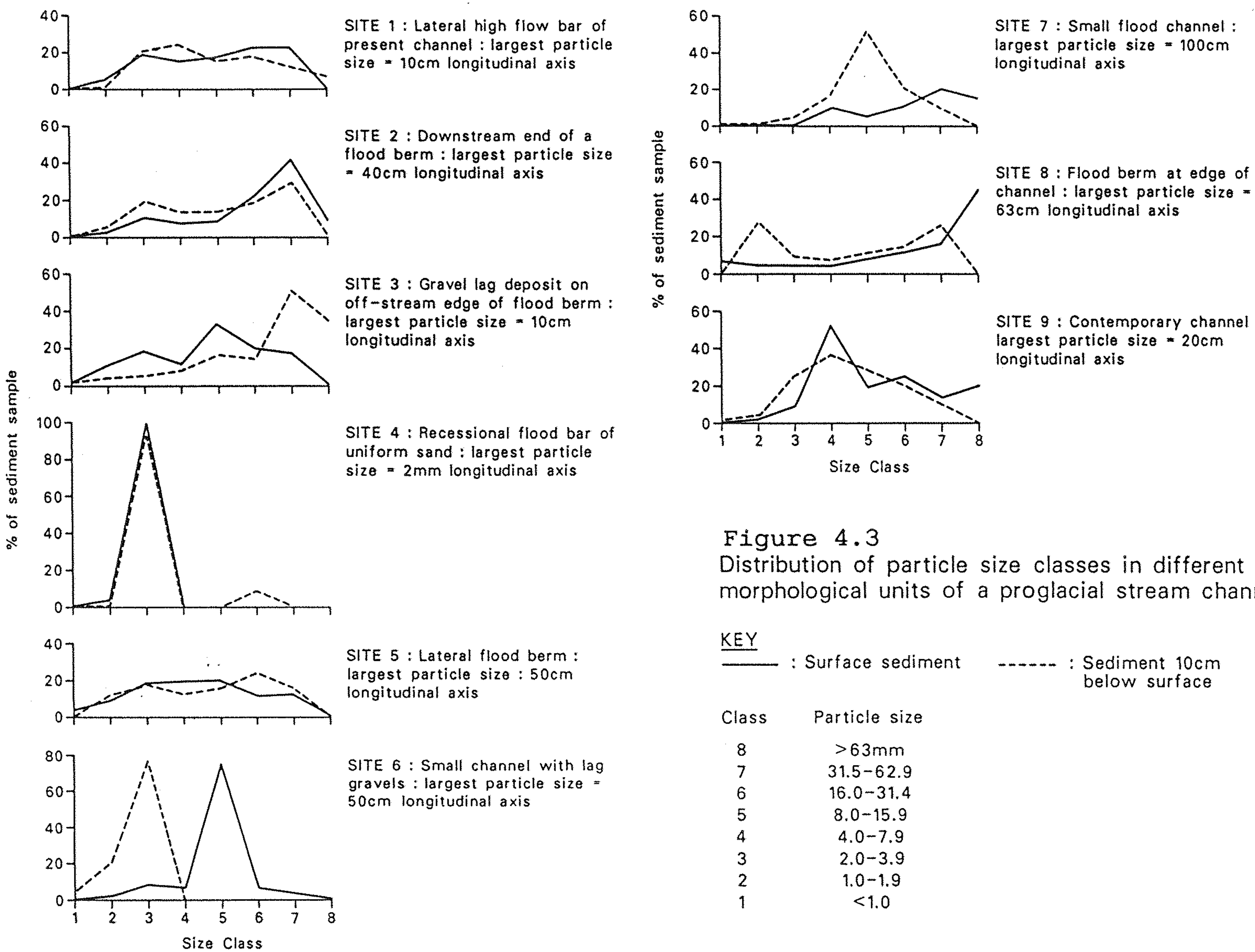


Figure 4.3
Distribution of particle size classes in different morphological units of a proglacial stream channel

KEY
——— : Surface sediment   ------ : Sediment 10cm below surface

Class	Particle size
8	>63mm
7	31.5–62.9
6	16.0–31.4
5	8.0–15.9
4	4.0–7.9
3	2.0–3.9
2	1.0–1.9
1	<1.0

described above. However, there is some overlap in the two groups of particle size distributions suggesting that the two are drawn from the same population of bed sediment. These results suggest that, in early June the channel and adjacent valley train contain a lot of stored sediment, which is fine enough to be removed from storage by early season flows.

In order to compare variations in channel sediment characteristics over the ablation season an investigation into the composition of surface and sub-surface bed sediment at 9 locations on the Bas Arolla proglacial channel was undertaken on September 9th, 1987. By comparing surface samples with sub-surface samples taken at a depth of 10 cm or at a depth greater than the depth of penetration of the largest surface clast, it was possible to test whether coarse surface sediment shielded finer sub-surface sediment from fluvial entrainment. This effect can be produced by processes of armouring, paving or censorship which are commonplace in gravel and boulder bed channels (Carling and Reader, 1982; Church et al., 1987; Andrews and Parker, 1987). Sampling areas of approximately 1 m^2 were excavated, large pebble axes measured and sieved into 8 size classes, ranging from 1 to 63 mm and weighed. The results (Figure 4.3) clearly demonstrate that a crude armouring was present. This was particularly well represented at sites 2 and 6 and to a lesser extent at sites 1, 7, 8, and 9. These sites represent a range of valley train fluvial environments including flood berms, small subsidiary channels and the contemporary channel. Site 4 showed very similar distributions between the surface and sub-surface with a slight fining upwards. Site 3 represented the only real anomaly, the sub-surface deposit was coarser than the surface. This is thought to represent a waning flow deposit. It appears there was less fine sediment available at the surface at the end of the melt-season than there was at the beginning of the season (June

2nd), which suggests most of the fine sediment has been removed or is stored sub-surface.

Surface sediment size distributions were also characterised by sampling with a 1 m^2 grid divided at 10 cm intervals. The b-axis of each pebble was measured at the grid-intersections and the median of all the pebbles was used to approximate a representative pebble size for the surface. This techniques was applied to the channel bed and to boulder cluster forms alike. Median diameters of the channel bed varied between 60 to 80 mm whereas those for the cluster diameters ranged from 128 to 430 mm. Both bed sediment and clusters were sampled, since the 'break-up' of either would have an influence on the bedload transport rate. Cluster break-up was thought to be particularly important in releasing sediment from the step-pool sequences (Reid et al., 1984) of the channel and for bed rupture of armouring in the braided sections.

In summary, at the start of the melt-season there was an abundance of relatively fine sediment at the surface in the main channel, whereas late in the melt-season crude armouring was widely developed. Grain sizes were highly variable and various bed elements (e.g. bars and clusters) have different characteristic grain sizes.

4.3 Suspended sediment transport - a comparison of paired suspended sediment series.

In 1987 two suspended sediment monitoring stations were set up at proximal and distal ends of the Bas Arolla proglacial zone (Chapter 2, Figure 2.1). The upper station was located 40 m downstream from the snout and the lower station was sited on the Grande Dixence water intake structure at the outlet from the proglacial channel reach (Figure 2.1). Both sites were located so that, at the point of measurement, flow was routed through a single cross-section. The upper station was positioned so as to be clear of the moraine accumulation zone but near enough to the glacier snout so that no tributary streams flowed between this station and the glacier snout (i.e. all tributary contributions were sampled) (Figure 2.1). Hourly suspended sediment concentration was estimated for the two sites using continuously recording turbidity meters and suspended sediment concentration/turbidity rating curves (Chapter 2). The R^2 for the upper and lower stations were 76% and 88% respectively. Discharge was gauged at the meltwater intake structure at the outlet from the proglacial zone. This does not account for differences in discharge between the two sites and the implications this has for the construction of the sediment rating curves. However, it is assumed that discharge is approximately equal at the two sites. Flow was turbulent and suspended sediment was well mixed in the flow at both sites.

Estimates of total load, based on turbidity records from the two sites, for the period May 29th to July 19th provide estimates of 5646 tonnes from the upper site and 6226 tonnes for the lower site which suggests a net downstream increase in load of 580 tonnes. The true differences would be greater because discharge is over-estimated at the upper site when tributaries are flowing. This net gain in sediment suggests inputs from within-channel or tributary sources. Inputs from tributaries (Chapter 5) account for between 103 and 319 tonnes, but not all of this will be transported as suspended load. A large proportion will be deposited as bedload in the main channel and along channel margins. For example measurements of tributary sediment discharge on June 15th (Chapter 5) produced a gross yield of 53.4 tonnes but only 7.8% (4.15 tonnes) was transported as suspended load past the downstream main channel monitoring station.

Cumulative daily suspended sediment load plotted against time for the two main stream monitoring sites (Figure 4.4) gives an indication of the flux in sediment storage in the channel between the two monitoring stations. The two curves have a similar form up until June 29th. After this date suspended sediment load increased more markedly at the lower site until mid-July. This increase was not a smooth rise but consisted of two distinct steps, one at the beginning of July and a second between July 5th and July 7th. By July 13th cumulative suspended sediment transport at the two sites had equalised. Between the 15th and 18th of July both sites showed a rapid rise in suspended sediment load which was associated with a major discharge event. Suspended sediment load at the lower site soon exceeded that at the upper site with approximately one third of the suspended sediment transport for the entire study period at the lower site being discharged during these 4 days. The stepped form of cumulative suspended sediment transport at the lower site suggests that sediment is being eroded from or input to the proglacial zone between the two sites and is then, during periods of rising discharge, being partially replaced by deposition between the two sites. There is a rapid rise in the suspended sediment load at the lower site when stored sediment is evacuated, which is then followed by a period of reduced transport rates

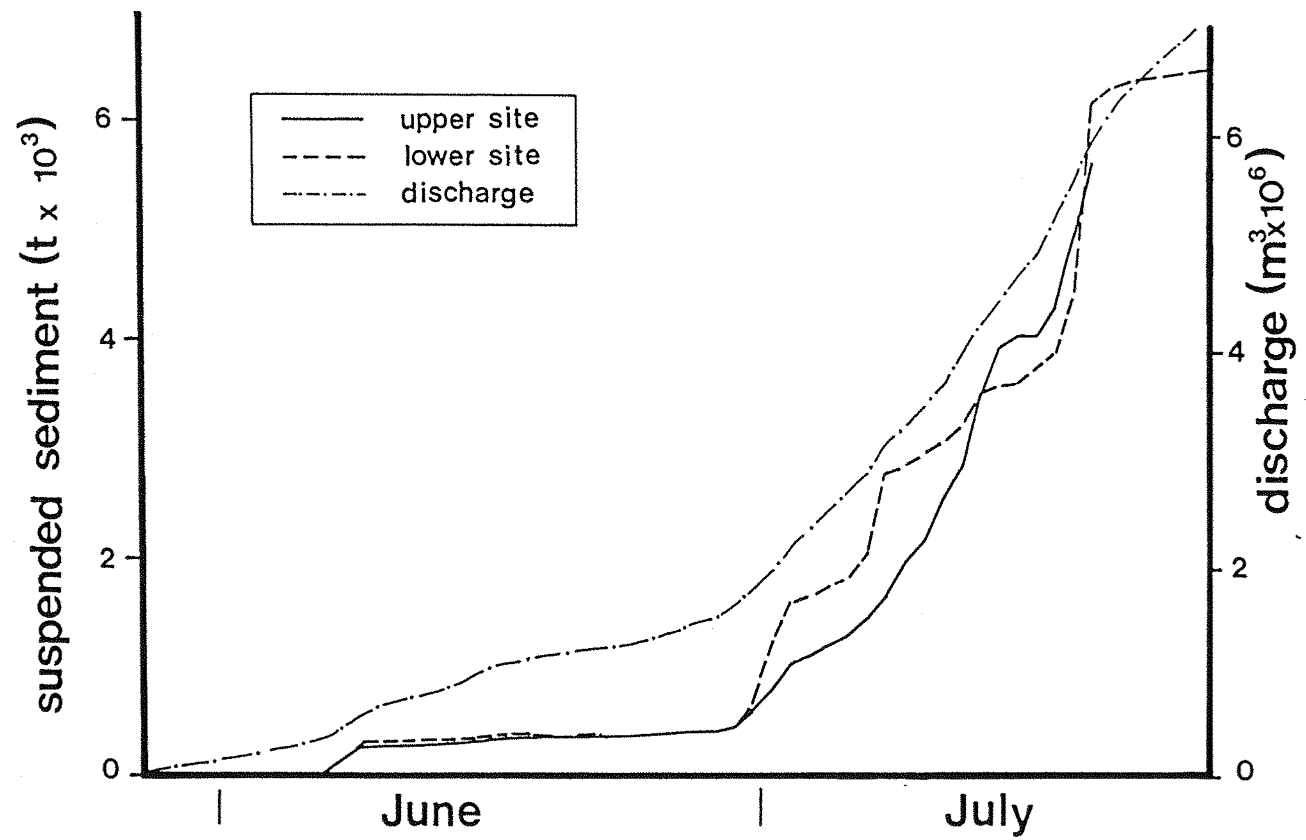


Figure 4.4 Comparison of cumulative suspended sediment loads monitored at upper and lower sites on the Bas Arolla proglacial stream (1987). Site locations shown in Figure 2.1.

over the observation period (Figure 4.4). Following the high discharge event of July 15th-18th suspended sediment discharge at the lower station is considerably reduced (records are not available for the upper site). This implies that most of the available sediment in both the glacial and proglacial sediment systems has been depleted.

In summary, from late May until the end of June sediment loads were very low with the only major sediment release on June 8th-9th. From the end of June to the end of July sediment load increased dramatically with the lower station loads initially exceeding the upper site loads implying some suspended sediment supply from the proglacial zone. From July 13th more sediment was contributed from upstream (glacial) sources than passed the lower station implying that suspended sediment was entering temporary proglacial storage. From July 15th to 18th there were very high flows and the suspended sediment load at the lower station increased dramatically possibly as a result of 'flushing-out' proglacial, mainly channel, sediment stores. It appears that proglacial sediment sources play a significant role in regulating suspended sediment discharge.

Suspended sediment concentration series were analysed further to assess the degree and the pattern of interdependence between records from the two sites. Suspended sediment time series can be compared using cross correlation analysis (Chatfield, 1987). This involves the estimation of a cross correlation function (CCF) from the two time series. The two sequences of observations are compared at successive time lags Cross correlations are calculated for both positive and negative lags and are plotted against lag (match position) in the form of a correlogram. Positive lags describe the impact of the record at the upper or the lower site. Negative lags are assumed to be related to feedback effects.

Three data sets were analysed:

1) The full series - based on hourly values of suspended sediment concentration for the period May 29th to July 19th 1987 (Figure 4.5).

2) Partial series - two days of 5 minute interval suspended sediment concentrations for a period characterised by high glacier meltwater discharge from July 4th (1200h) to July 6th (1200h) (Figure 4.6).

3) Partial series - 24 hours of 5 minute interval suspended sediment observations on June 15th, a day of prolonged and heavy rainfall and very high tributary suspended sediment discharge (Chapter 5) (Figure 4.7).

The correlograms for the hourly series (May 29th to July 19th) is shown in Figure 4.5. Correlograms for both the raw data and first differenced series clearly show that the best match position is at lag zero This suggests that the two series are directly and positively related. The sediment 'signature' from the upper site is faithfully reproduced (cross correlation after first differencing = 0.8) at the lower station within one hour (or between 0 and 2 hours given the nature of the sample stratagem) The physical significance of this result is that tributaries contribute insufficient sediment to interfere with the mainstream transport pattern over the reach and that channel processes also do little to alter the suspended sediment time series pattern.

The inter-relationships between suspended sediment concentration series at the upper and lower sites can be investigated in more detail by examining partial series of 5 minute determinations for two short periods: July 4th to 6th and June 15th. The July series (Figure 4.6) show a close match between the two series with lag 1 having the highest cross correlation. This suggests an average lag of 5 minutes between the two series which corresponds to the mean travel time for a salt pulse travelling between the two sites. The two raw data

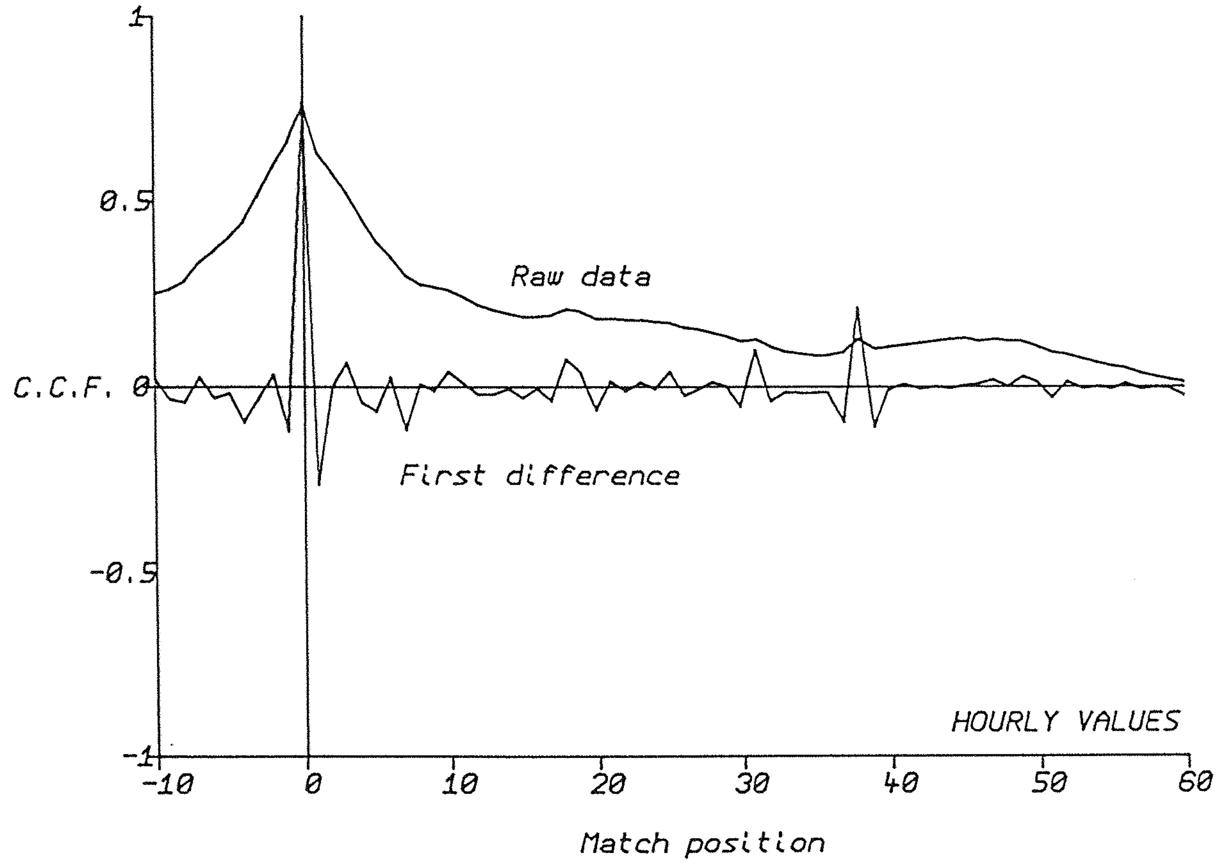
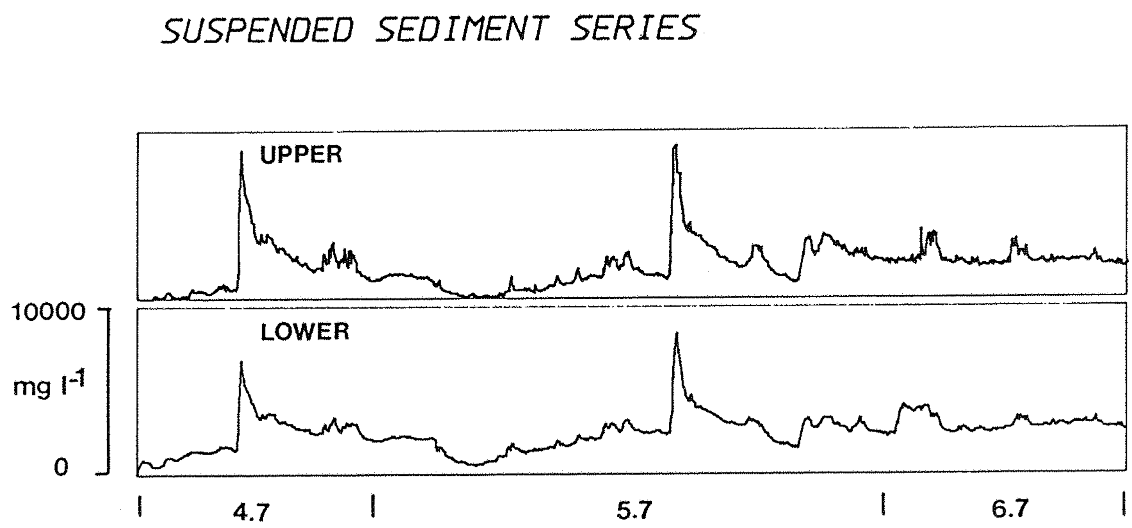


Figure 4.5 Cross correlogram of Bas Arolla upper and lower site suspended sediment series, May 29th to July 19th, 1987. Hourly values.



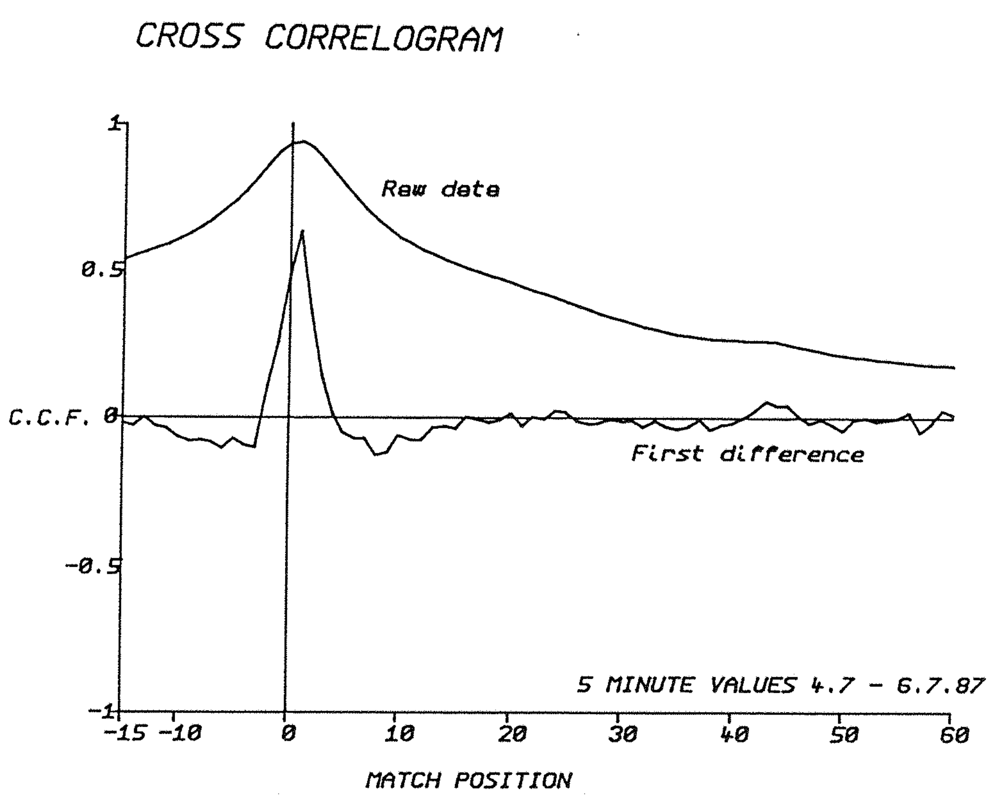


Figure 4.6 Cross correlogram of Bas Arolla upper and lower site suspended sediment series July 4th, 1200h to July 6th, 1987. 5 minute values. Upper part of diagram shows the two suspended sediment series used in the analysis.

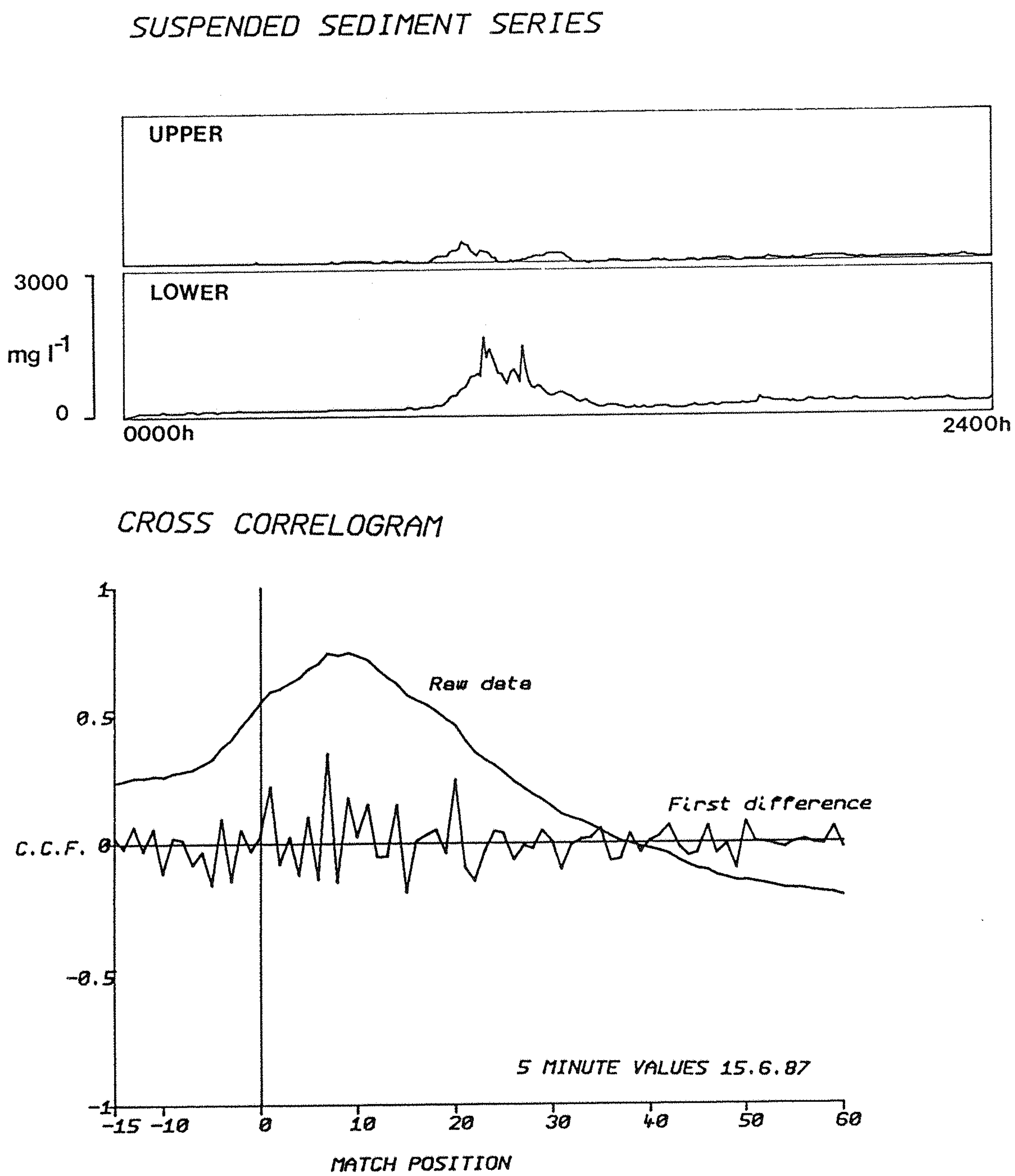


Figure 4.7 Cross correlogram of Bas Arolla upper and lower site suspended sediment series June 15th, 1987. 5 minute values. Upper part of the diagram shows the two suspended sediment series used in the analysis.

series (Figure 4.6) appear to be identical in virtually every respect. This suggests very little modification along the proglacial reach (although there is a slight increase in concentration between the two sites) and reinforces the view that suspended sediment generated from the glacier was transmitted through the stream channel with little addition of sediment.

On the 15th of June the pattern was markedly different, although the timing of the period of highest sediment discharge was roughly synchronous (Figure 4.7). Suspended sediment concentrations were lower than for July 4th to 6th, and large differences are obvious between the records from the upper and lower sites. At the lower site suspended sediment discharge was much higher and more 'spikey' than at the upper site. Statistical comparison shows very little correspondence between the two series (Figure 4.7). The best match position is around 36 minutes but little physical significance can be attached to this. It may be that this lag reflects differences in response time, of sub-catchment areas, to rainfall. For example, the glacier may respond more rapidly to rainfall than the tributaries which may lag behind until streamflow is generated. The role of channel sources in contributing sediment in this event is ruled out because it was observed that only tributaries contributed large quantities of turbid water and survey of the channel bed immediately below the main tributary input, and at two other cross sections in the reach, showed a net increase in bed storage. This implies that under the circumstances encountered in this June event tributaries can contribute an important element to the main channel suspended load. These 'tributary contribution' events are few because extreme rainfall days, which 'trigger' them, are not very common and, on the day in question, intense rainfall was accompanied by small scale debris flows and slope failures in the upper reaches of tributary channels, which greatly contributed to the sediment load. These events are unusual and therefore their contribution to the overall relationship between the stations (Figure 4.5) was probably only at the level of a 'noise' component.

The shape and size of the sediment pulses in the two examples may be diagnostic of source since the meltwater dominated series (July 4 to 6th) (Figure 4.6) was characterised by large asymmetrical pulses whilst the rainfall day (June 15th) had pulses which were much smaller, spikey and symmetrical (Figure 4.7).

4.4 Bedload transport within the proglacial reach

Recent reviews by Bathurst (1987), Lisle (1987) and Whittaker (1987) have examined bedload movement in steep coarse-bed channels and have all reached similar conclusions: 'bedload movement in coarse-bed channels is an extremely complex process, (Bathurst, 1987); 'field conditions for studying sediment transport in steepland channels are formidable and their behaviour is complex and episodic' (Lisle, 1987) and; 'considerable difficulties stand in the way of predicting sediment transport in steep mountain streams' (Whittaker, 1987). With this prospect it is not surprising that bedload, as a component of the total load of a stream, is rarely measured in mountain streams (Thorne et al., 1987). Field measurements are therefore essential since existing transport equations and sediment discharge rating curves provide poor estimates of transport rates. Bedload discharge can be estimated using a number of existing theoretical approaches (see the use of the Schoklitsch equation applied later in this section), but these have been shown to offer only an order of magnitude accuracy (Bathurst et al., 1987). Therefore, there is no substitute for direct measurement.

This section uses field measurements to describe the characteristics of bedload transport through the Bas Arolla proglacial channel system and attempts to identify the timing and magnitude of bedload transport. The basic theme adhered to is the interaction between bedload transport, discharge and changes in bed storage. This relates to suspended sediment transport because bed storage is important in contributing sediment to material in suspension (Lambert and Walling, 1986). Considerations of this kind are important for the sediment budget because it is important to estimate when and how much material is derived from the proglacial stream bed. This section is sub-divided into 2 parts: bedload transport - discharge relationships; and bedload transport variations. The first part discusses thresholds of bed material entrainment and the prediction of bed material transport and the second part examines the variations in observed transport rates. The measurements of bedload transport discussed here were collected using a Helley-Smith bedload sampler (Chapter 2) and should not be confused with bedload transport estimates based on purging of sediment traps at the meltwater intake structures (Chapter 3).

4.4.1 Bedload - Discharge relationships

Because of the large range of sediment sizes represented at the channel boundary, conventional bedload transport theory, epitomised in the Shields equation, might suggest a definite dependency between the size and amount of material in motion and the water discharge (Bathurst, 1987). Bedload transport measurements collected from the Bas Arolla proglacial stream (Figure 2.1, lower cross section) with a Helley-Smith sampler between May 25th and July 30th 1987 (Chapter 2) do not show a clear relationship with discharge (Figure 4.8). The raw data (Figure 4.8a) show that below 2.5 $m^3 s^{-1}$ bedload movement is negligible but, above 2.5 $m^3 s^{-1}$ there is a very rapid increase in transport rate. Plotting the Bas Arolla data after log transforming both axes (e.g. Leopold and Emmett, 1976) produces a near-vertical increase in bedload transport rate with discharges greater than 2.5 $m^3 s^{-1}$. (Figure 4.8b). This pattern only applies to a restricted discharge range up to 3 $m^3 s^{-1}$ but includes channel-full discharge which is between 2.6 and 2.8 $m^3 s^{-1}$. This suggests the may be a critical discharge (threshold) for initiation of bedload transport.

This apparent threshold value can be compared with critical conditions for the initiation of bedload movement calculated using the Schoklitsch equation (Schoklitsch, 1962; Bathurst et al., 1987). The

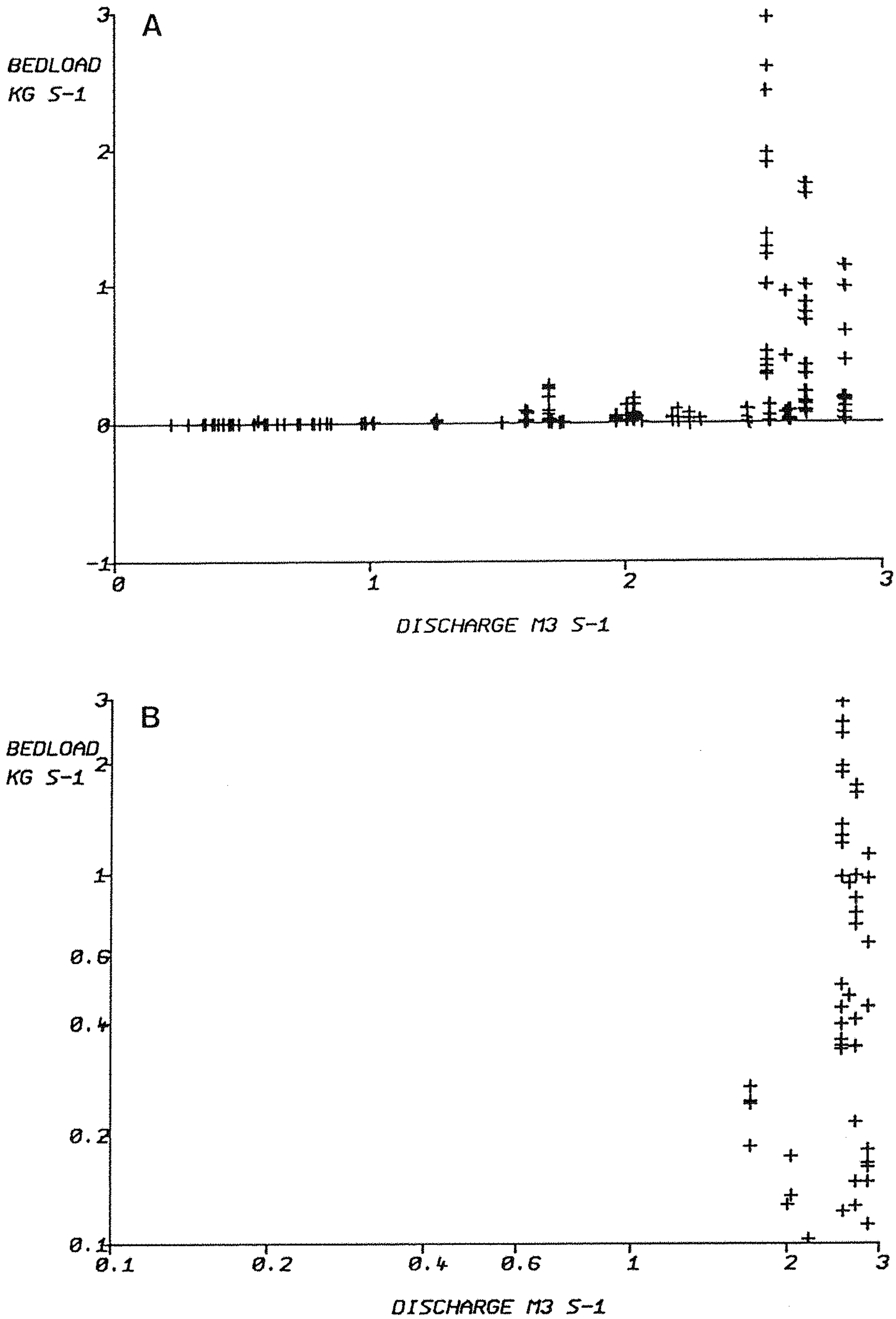


Figure 4.8 (A) Bedload transport against discharge (B) Log transformed bedload transport against discharge.

equation used was:

$$q_{cr} = 0.26 \left(\frac{\rho s}{\rho} - 1 \right)^{5/3} \frac{D_{40}^{3/4}}{S^{7/6}}$$

where: q_{cr} = critical water discharge for initial movement; ρs = sediment density; ρ = water density; D_{40} = size of particle median axis for which 40 per cent of the sediment is finer; and S = channel slope (0.077 m m^{-1}). Calculations were based on the D_{40} of a bulk bed sample collected from the channel bed in June. The estimated unit critical discharge (and critical discharge calculated for the measurement cross section) was:

	D40 mm	Unit discharge $m^2 s^{-1}$	Discharge range $m^3 s^{-1}$
Bed sample	11.3	0.095	0.478 - 0.669

In addition, unit critical discharges can be calculated for the D_{40} of bed material grid samples; and for boulder and cobble cluster bedforms:

	D40 mm	Unit discharge m^2 s-1	Discharge range $m^3 s^{-1}$
Grid sample	50.0	0.142	0.709 - 0.992
Clusters	263.0	1.710	8.548 - 11.967

The range in critical discharge estimates is based on the variation in the active bedload transport width in the cross section studied here.

Estimating bedload discharge from a theoretical relationship between flow conditions and bedload transport rate seems an attractive proposition (Walling and Webb, 1983). However, the application of these equations in field situations is fraught with difficulties (Leopold and Emmett, 1976). Sediment transport functions or sediment discharge give the rate of discharge of sediment in terms of the sediment properties and the hydraulic properties of flow. Equations rely on the experimental determination of coefficients often in small-scale laboratory flumes. These empirical constants demand testing in natural streams (Church and Gilbert, 1975; Richards, 1982; Dawdy and Vanoni, 1986).

Testing or calibration of bedload transport formulae is a daunting prospect since most of the relationships used to estimate sediment transport are too complex for ready evaluation (Lewin, 1981). Furthermore the true processes are too complex to be accurately modelled and available measurement techniques are too simple to adequately characterise process (although vortex trap samplers, etc hold considerable promise (Tacconi and Billi, 1987)). This disparity means that calibrations of bedload discharge equations can only be approximate. In any event, complete agreement between predicted and measured rates will never be achieved because bedload discharge is not a direct function of stream discharge. It is more a function of flow strength and the physical arrangement of the bed.

Nevertheless, bedload transport equations should still be tested since they can provide insight into bedload movement. Several equations are available for predicting sediment transport in steep mountain rivers (Bathurst et al., 1987; Jaeggi and Rickenmann, 1987; Wiberg and Smith, 1987). The choice of equation is somewhat arbitrary as long as the equation was developed for flow conditions and sediments typical of mountain streams. Therefore, the Schoklitsch equation (Schoklitsch, 1962) is suitable, given conditions of unlimited sediment availability (Bathurst et al.,1986b), but application of this formula has met with limited success. For example, transport rates calculated using this equation were 3 orders of magnitude greater than actual measured transport rates during a period of snowmelt discharge on the Roaring River, Colorado (Bathurst et al., 1986a). Only under conditions of near unlimited sediment availability will the Schoklitsch equation predict the correct order of magnitude of bedload transport (Bathurst et al., 1987). Hean and Nanson (1987) tested 8 bedload transport equations, including Schoklitsch, at 11 gauging sites on eastern Australian rivers and found the formulae inherently unreliable under field conditions, concluding that using formulae to determine absolute bedload yields was a futile exercise.

Since the question of which equation is most appropriate for field application remains open (Jaeggi and Rickenmann, 1987) further bedload transport data are needed to test existing equations. Of the available equations application of the Schoklitsch equation is attractive since it is based on discharge rather than shear stress or stream power which are considerably more difficult to measure in mountain streams (Wiberg and Smith, 1987). To some extent this is unimportant since the choice of flow variable selected to predict bedload transport is arbitrary (Carson, 1988). But practically, a discharge-based approach is important, because discharge is frequently measured by hydro-electric companies and could therefore, be used in assessing the propensity for bedload transport.

The calculated critical discharges (given above) can be used to calculate characteristic bedload transport curves using the Schoklitsch transport equation (Graf, 1971; Grimshaw and Lewin, 1981; Bathurst, 1987):

$$qb = 2500 \; S^{3/2} \; (q - q_{cr})$$

where qb = bedload discharge (kg s^{-1}); S = channel slope; q = water discharge ($m^3 s^{-1}$); and q_{cr} = critical water discharge for the initiation of movement. Plotting these relationships against the field data (Figure 4.9) shows no correspondence between the calculated thresholds of motion and the measured movements but, using a critical discharge of 2.4 $m^3 s^{-1}$, derived from Figure 4.8a, the Schoklitsch curve mirrors the initial rapid rise in bedload transport very well. However for higher discharges, which occurred later in the season the predictions are poor implying that, sediment supplies were limiting transport. seems limiting. This emphasises the history-dependent nature of the transport process (Østrem et al., 1971; Bathurst et al., 1986a) and the very rapid depletion of sediment sources.

The relationship calculated for the movement of clusters plots well off the range of the Helley - Smith measurements and is not shown. The predicted initiation of motion for clusters is not reached until a critical discharge of approximately 10 $m^3 s^{-1}$ and discharges of this magnitude only occur during extreme floods. This is not surprising since Whittaker (1987) suggests bed steps (clustered bedforms) of step-pool streams are relatively stable over the usual flow conditions. Indeed it is only after large floods that the channel pattern is substantially altered (Chapter 6).

The lack of correspondence between observed movement rates and the calculated thresholds may be partly caused by sampling methods. The grid sample is more representation of surface bed conditions than the grab sample, since it better reflects the conditions at the channel boundary/flow interface. Also the curves shown are 'mean' curves based on one characteristic grain diameter. Bed materials have been shown to be widely variable in mountain streams so the threshold would be better represented by an envelope of curves.

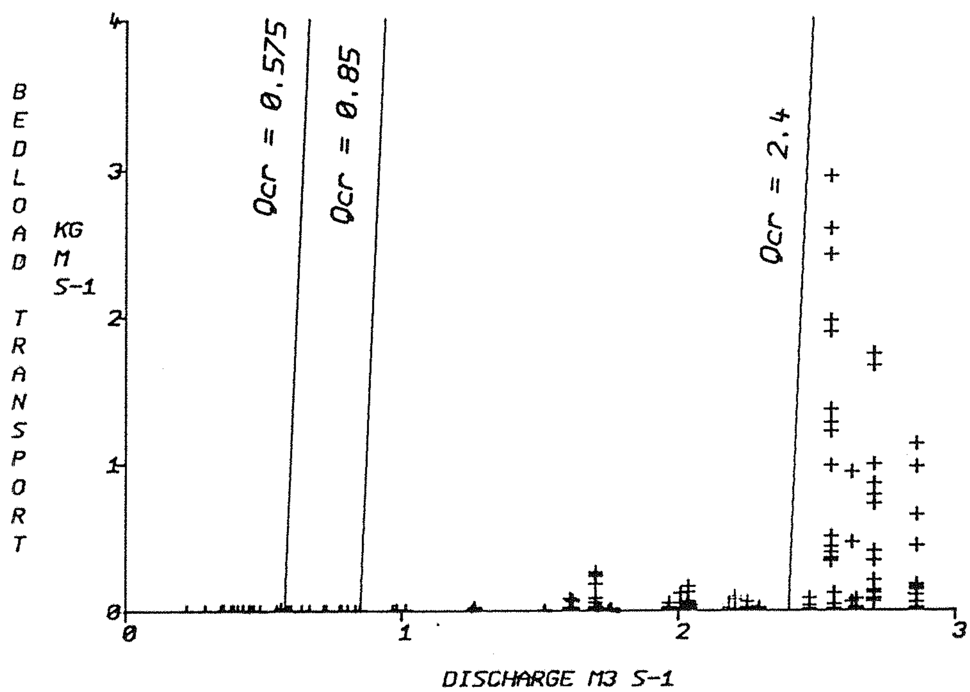


Figure 4.9 Comparison between Bas Arolla proglacial stream bedload transport data and critical water discharge for the initiation of bedload movement predicted by the Schoklitsch equation. a = calculated on the basis of grain size D40 of bed material grab samples, b = calculated on the basis of grain size D40 of bed material grid samples, c = determined from Figure 4.8a.

Further, calculations are based on on one representative grain size, however, both the conditions for the initiation of motion and the transport rate itself for a given size fraction are known to be affected by the presence of other sizes in the mixture. In other words the transport equation cannot be applied to each size fraction as if to a material of uniform size distribution. Small particles are shielded from the flow by larger particles and are more difficult to move than if in a uniform size distribution. This hiding/exposure effect has been quantified with some success for the initiation of motion (Andrews, 1983), although further work is still needed. However rather less progress has been made in accounting for the effect in transport equations (e.g. Misrii et al., 1984; Samaga et al., 1986). This emphasises the need for detailed representative measurements of non uniform size distributions of bed materials and considerations of relative and absolute particle sizes.

For gravel bed rivers Jackson and Beschta (1982) divide the sediment transport process into two phases. Phase 1 movement involves the flushing out of sands deposited in the channel during low discharges. The bed structure is undisturbed and thus, initial transport rates are low (Figure 4.10a). With increased flow bottom velocities and associated shear stresses increase and the armour layer is disrupted as gravels are entrained and erosion of bed sediments proceeds rapidly as smaller relative particle sizes are exposed. This disruption of the armour layer is phase 2 transport (Figure 4.10b). In mountain streams a third phase may be envisaged since the bed structure of mountain streams alternates between reaches of armoured bed (described in Jackson and Beschta (1982)) and step pool sequences of clustered boulders. This will have the effect of producing a third threshold for bed movement which involves the break up of clustered bedforms (Whittaker, 1987) which would produce massive sediment transport in the stream

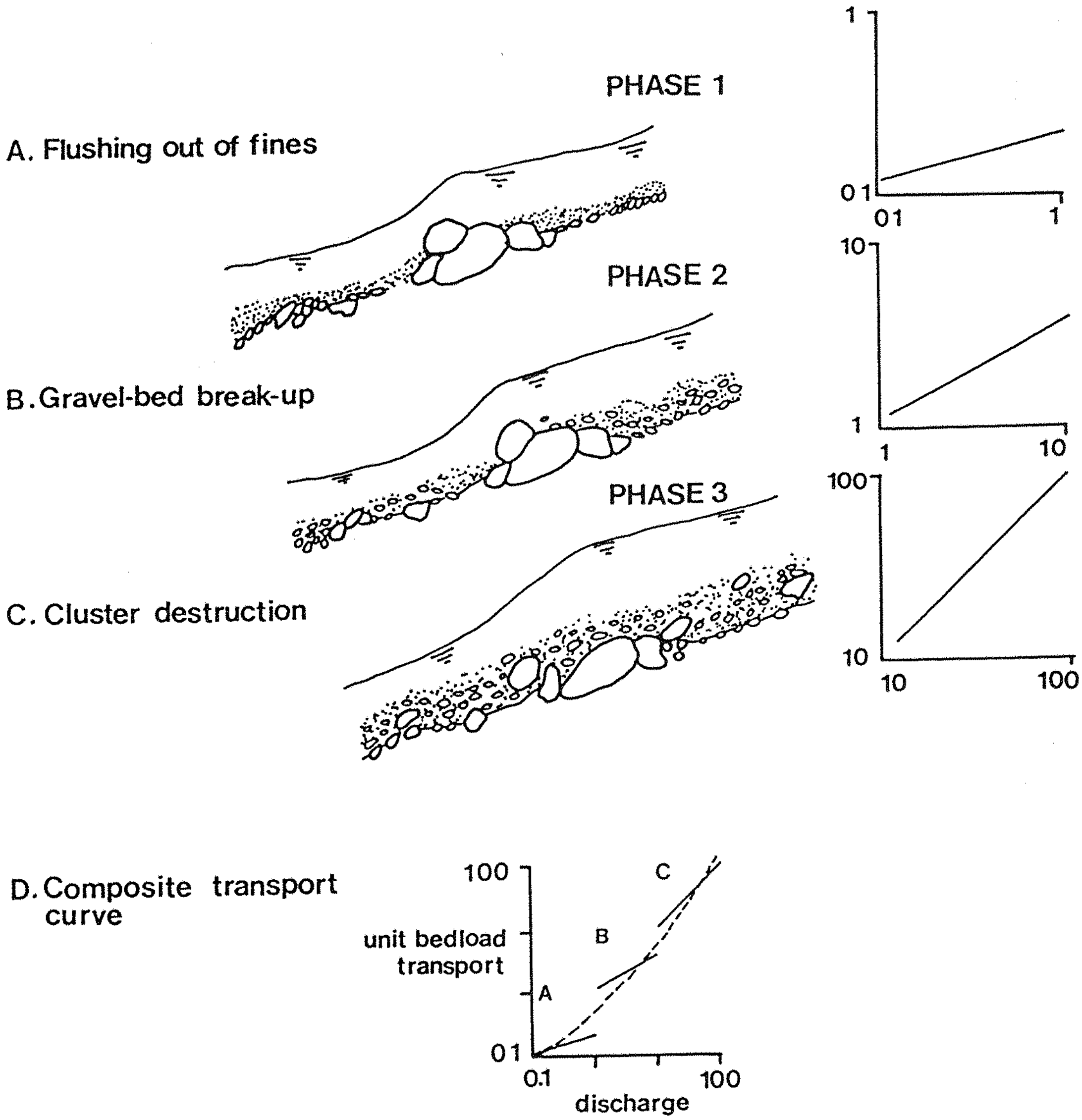


Figure 4.10 Three phase model for bedload transport in mountain streams.

channel (Figure 4.10c). This is important since the identification of a hierarchy of bedforms suggests that transport will vary between extreme discontinuity in phase 1 to almost continuous transport in phase 3, (i.e. a mobile particle in phase 1 has two bed hierarchies to negotiate (interstices of the armour and cluster forms), a mobile bed particle in phase 2 has one obstacle (Cluster forms) and in phase 3 all the bed is in motion). This means that transport rates will be successively greater from phases 1 through 3 (Figure 4.10). As a result the composite sediment transport curve (Figure 4.10d) will probably have a curvilinear form. As no Helley Smith transport measurements were taken in phase 3 (or could be !) this is unproven but the model does help to explain the discontinuity and in the bedload transport relationships. The very rapid increase in transport rate at discharges greater than 2.5 $m^3 s^{-1}$ could also be interpreted as being indicative of equal mobility transport where above a certain threshold all the bed is mobilised (Andrews and Parker, 1987).

4.4.2 Field measurement of bedload transport variations

Given that the prediction of bedload movement is not easily modelled, field observations are essential in describing the general nature of the transport processes and will provide the ultimate test of bedload transport hypotheses and models (Thorne et al., 1987a). Variations in bedload transport have been demonstrated in both flume (Kuhnle and Sutherland, 1988) and field conditions (Pitlick, 1988). The variations in bedload transport at various spatial and temporal scales are considered in this section in order to characterise the general nature of movement. The results relate to conditions measured at the lower channel cross section (Figure 2.1). All measurements of bedload transport were collected using a 152 mm orifice Helley-Smith bedload sampler.

Figure 4.11 shows the relationship between discharge, Helley-Smith bedload transport measurements and changes in bed sediment storage (determined from successive cross section surveys). Bedload was sampled at between 1300 to 1400h each day (days where longer periods of sampling occurred are also shown which explains the range in values for some days). At the seasonal scale there seems to be some general agreement between discharge fluctuations and the Helley-Smith bedload transport measurements (Figure 4.11), even though the samples are relatively infrequent and there are large variations in transport rates (note log scale). Particularly important was the large flood between July 15th to 18th but, because of its magnitude, the flow was not fully gauged and Helley-Smith bedload measurements could not be collected. Survey of the channel bed (before and after the flood), and the frequency of sediment trap purges revealed major sediment losses during this period. In fact, the purge records show that bedload transport during the flood were the highest recorded, until the end of July. These data imply a

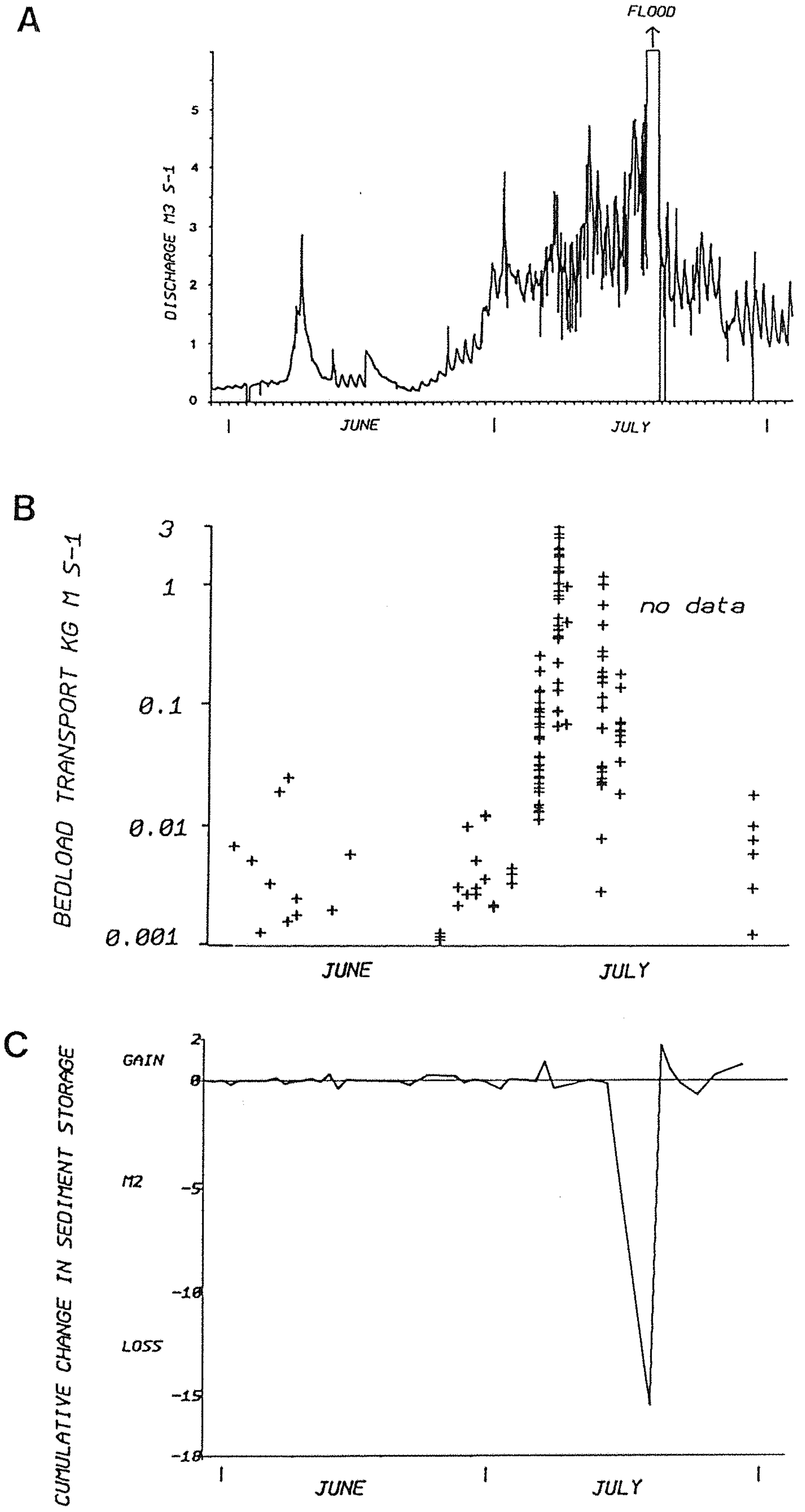


Figure 4.11 Comparison between 1987 seasonal discharge, Helley Smith bedload sampling and cumulative change in sediment storage at the lower 'daily' cross-section.

crude relationship between bedload discharge and water discharge (Figure 4.11) even though bedload discharge is extremely variable and the bed remains relatively stable except during flood. The variation in bed sediment storage following the major flood of July 15th to 18th suggests that channel banks and bedforms need to re-adjust to attain a stable structure.

Variations in transport rate also occur on shorter timescales. Diurnal measurement of bedload discharge at hourly intervals on July 6th, between 0920h and 2010h, reveal marked fluctuations in the maximum transport rate even though the minimum transport rate is relatively stable (Figure 4.12). Variations in peak transport rates did not correspond to discharge peaks, rather variations in bedload transport seemed to respond to variations in discharge whether increasing or decreasing. This suggests that flow variability was more important than discharge magnitude in regulating bedload discharge. Indeed over very short timescales (seconds), Ashmore (1987) has shown that fluctuations in bedload transport can be related to turbulent velocity bursts. Alternatively, supply factors such as bank erosion along the channel margin or sporadic inputs of sediment from outside the channel may have produced the observed variability. Indeed during this period of bedload measurement the local bed elevation changed little (Figure 4.12) which suggests that supply was probably not from the immediate channel bed but from upstream. The general lack of any locally definable relationship between bedload transport rates and discharge is illustrated in the relationship between the grain size distributions of Helley-Smith bedload samples and discharge (Figure 4.13). In this instance bed materials were initially relatively fine (9.25) and became coarser with falling discharge (1115h and 1310h) became fine again (1510h) and later tended to become coarser (1710h and 1910h). The discharge pattern was initially high (0925h), declined (1000h-1300h) and

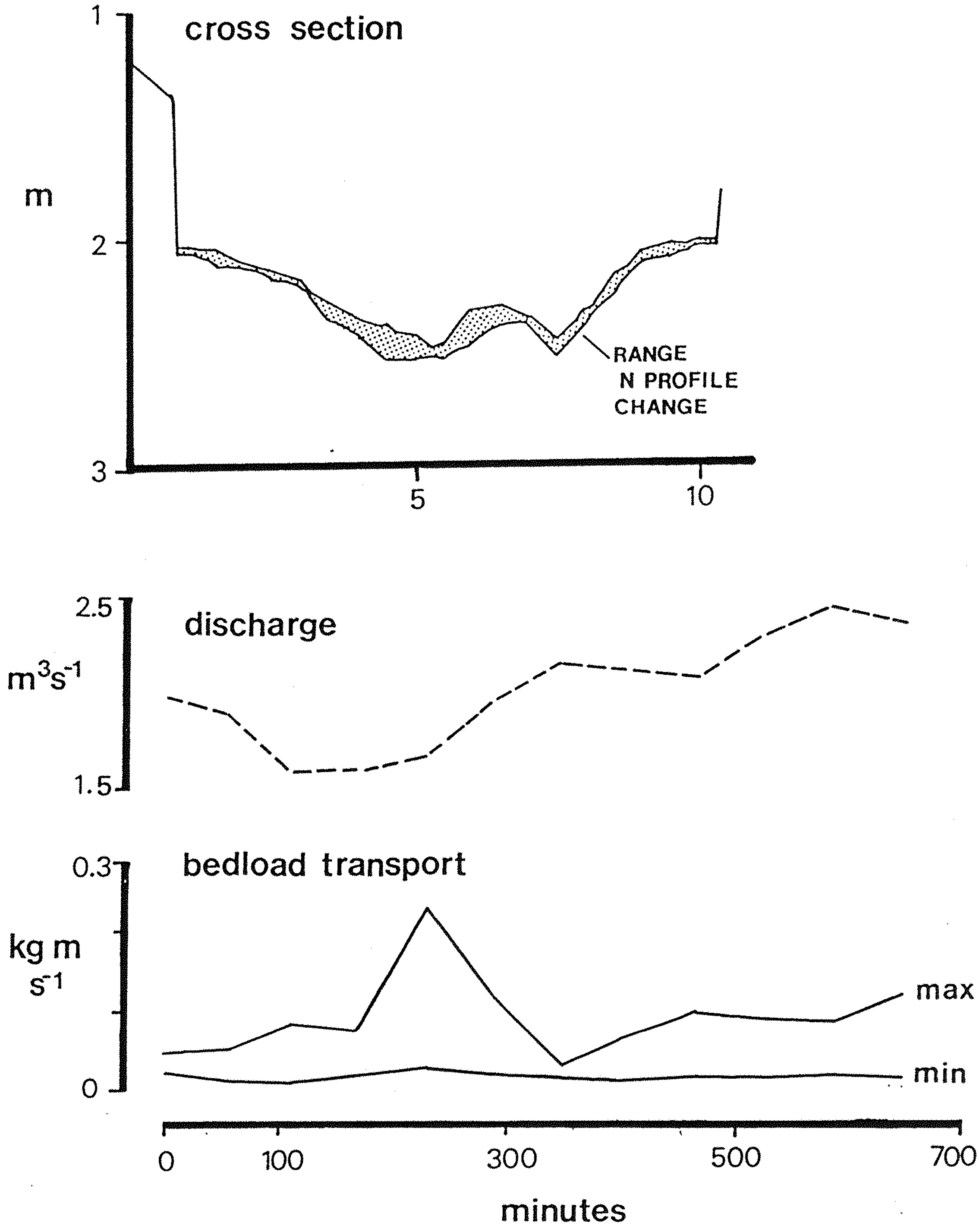
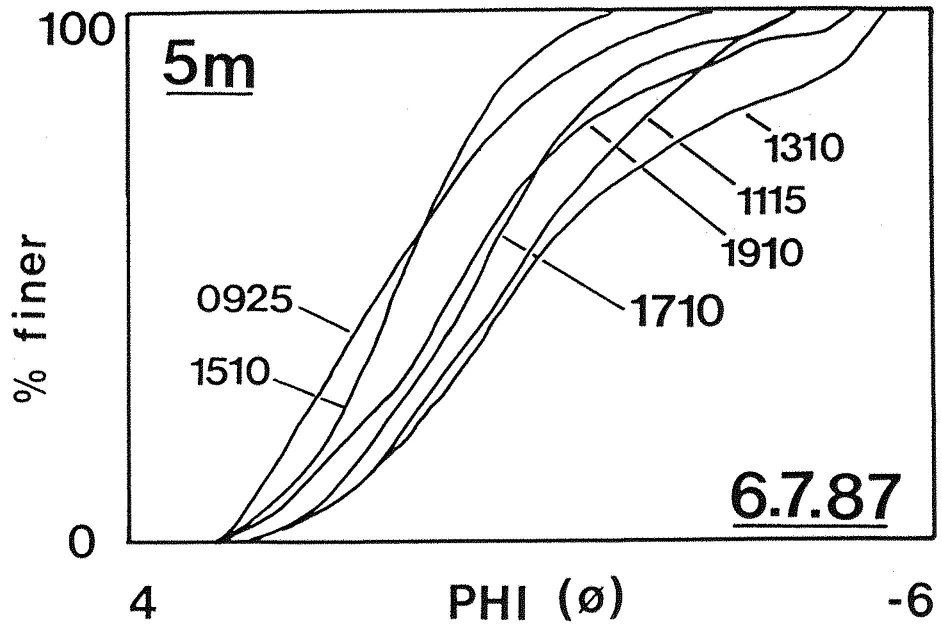


Figure 4.12 Comparison between hourly surveys of channel cross-section, bedload transport (Helley Smith sampling) and discharge on July 6th 1987. Cross-section shows the envelope of bed elevation changes during the sampling period.



TIME h	*DISCHARGE m^3 s^{-1}*	*BEDLOAD TRANSPORT kg ms^{-1}*
0925	*2.181*	*0.025*
1115	*1.605*	*0.091*
1310	*1.699*	*0.187*
1510	*2.287*	*0.027*
1710	*2.202*	*0.085*
1910	*2.640*	*0.086*

Figure 4.13 Grain-size characteristics of Helley Smith bedload samples in relation to sampling on July 6th 1987.

gradually increased from 1300h onwards.

Intensive bedload sampling at 10 minute intervals on July 8th showed even more pronounced variations in transport rates during a period of almost constant discharge (Figure 4.14) (e.g. Beschta, 1983). During the 2 hour sampling period both minimum and maximum transport rates generally increased although fluctuations were very marked. Survey of the channel bed before, during and after the sequence of measurements shows the bed elevation to increase between 1020h and 1120h and decrease from 1120h to 1200h which suggests the possible passage of a bed sediment wave or load pulse because by 1150h bedload transport rate was declining.

The above example (Figure 4.14) illustrates short duration changes in bed elevation associated with sediment transport. Such changes in bed storage have been previously associated with the passage of much larger bed megaforms in alluvial channels (Ashmore, 1987; Whiting et al., 1988) and are a useful indicator of changes in sediment load (Grant, 1977). In mountain streams it is still possible to get changes of this kind associated with the migration of bedforms (Klingemann and Emmett, 1982) and in a similar fashion bulk inputs of sediment move as long low waves through the channel (Whittaker, 1987). In order to assess whether such features were responsible for sediment transport in the Bas Arolla proglacial stream 22-day series (May 27th to June 17th) of daily bed sediment storage measurements were compared for three sites along the proglacial channel reach (Figure 4.15). The Bas Arolla stream was thought susceptible to this kind of movement given the relative abundance of fine sediment stored in the channel early in the season (Section 4.2). Such fluctuations in bedload transport can be produced by longitudinal sediment sorting (Iseya and Ikeda, 1987), periodic surges of discharge causing sediment

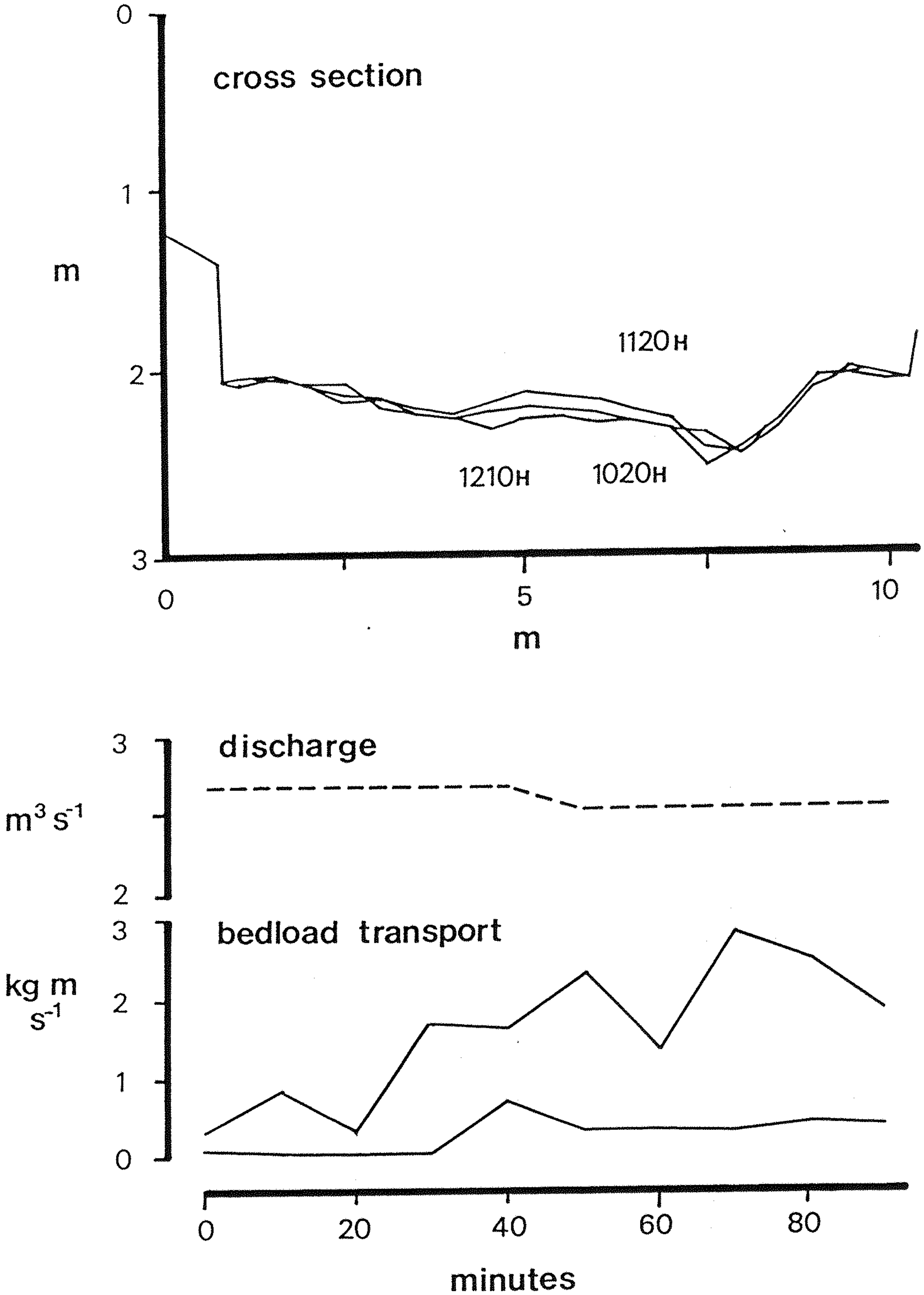


Figure 4.14 Intensive sampling of bedload transport (Helley Smith samples) at 10 minute intervals on the Bas Arolla proglacial stream. 1020h to 1220 July 8th 1987.

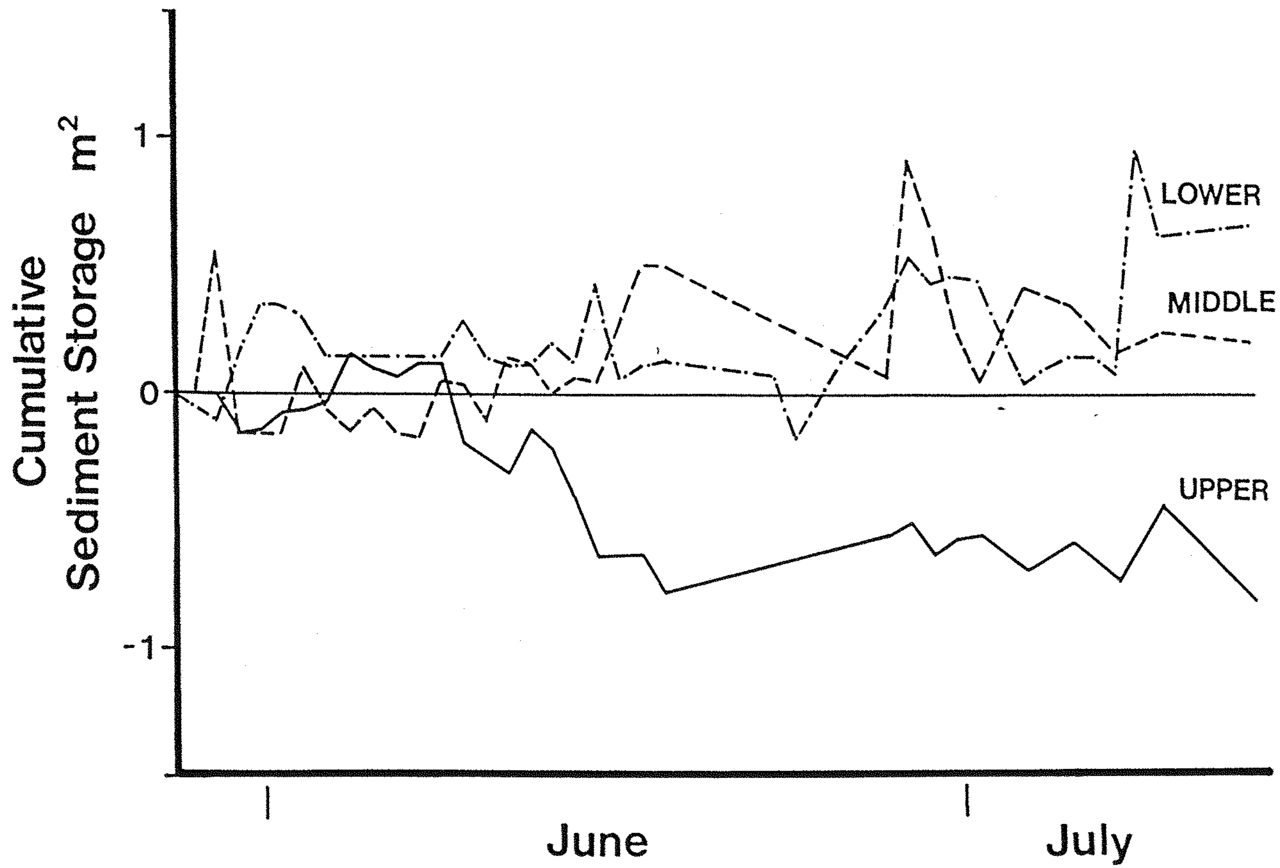


Figure 4.15 Cumulative changes in cross-section sediment storage at the three 'daily' mainstream sections (Figure 2.1). May 27th to July 13th 1987.

mobilization (Heggen, 1986) or may simply reflect local fluctuations around sediment transport thresholds at the channel boundary (Maizels et al., 1987).

Cross correlation and auto correlation techniques (Chatfield, 1987) were used to determine whether bed changes could be cross correlated between sites and if so to estimate relationships in their movement. There was no significant cross correlation between the 'upper and middle' and 'upper and lower' sites. The only significant cross correlation was between the 'middle and lower' sites (distance apart 40 m). The best match position was at lag zero (i.e. within the same day) and therefore probably reflects general bed changes associated with reach-wide transport events since a lag of greater than zero would be expected if large-scale bed megaforms were being transported slowly along the channel. In addition no significant periodicities (lags) were found in any of the auto correlation functions of the 3 series which implies no systematic change in bed storage. Cumulative changes in bed elevation at the three sites are very different (Figure 4.15) the lowest site showing overall net loss in storage and the upper sites showing net gain up until July 13th. Contrasts in channel storage between the 3 sites probably reflect differences in the type of bed materials (Bathurst, 1987) or channel morphology (Lisle, 1987) (e.g. contrasts between braided and step - pool channel segments). In this the three sites studied here were all single-thread.

Substantial variations in transport also exist in the cross-channel dimension with bedload moving as threads along the channel bed (Bathurst et al., 1986b). This streaming effect (Figure 4.16) varies markedly between days. The three examples shown in Figure 4.16 all show a central channel bedload peak which roughly corresponds to the central thread of flow. Figure 4.16a shows variations in transport rate of up to 10 times in the

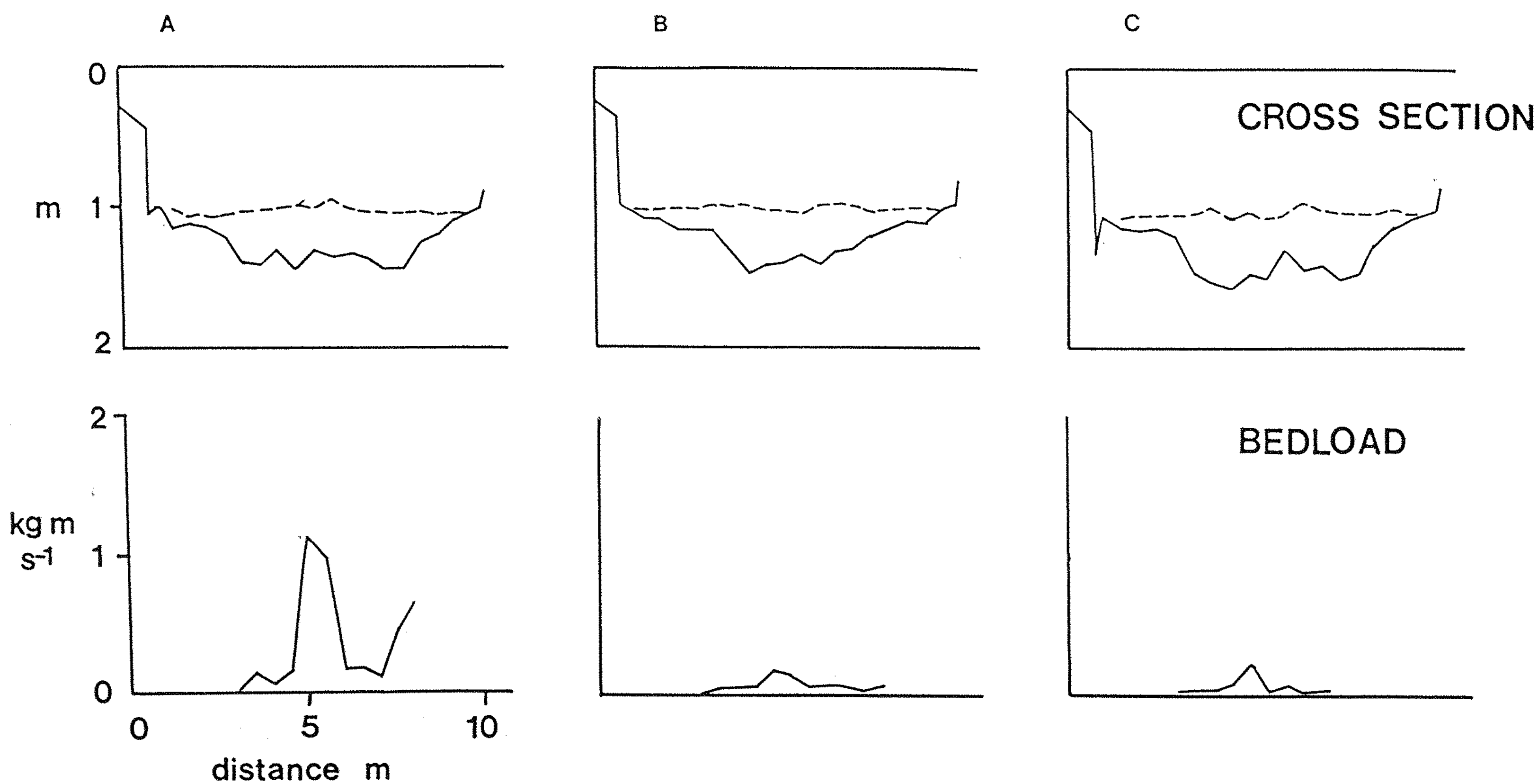


Figure 4.16 Cross-section variations in bedload transport at the Bas Arolla proglacial stream lower section 1987. Variations show a central channel thread of maximum transport.

space of 1 metre (collected over a time period of 10 minutes). Furthermore measurements at three mid-channel sites spaced 1 metre apart on the 6th September 1987 showed both synchronous and out of phase fluctuations in transport rate over time (Figure 4.17).

Such varied observations have implications for sampling and for the calculation of mean transport rates and may also help to explain the variance in observed transport rates which are based only on a few cross channel samples. For example the relationship between grain size and discharge presented in Figure 4.13 was only measured in one vertical so variations may reflect the shifting thread of maximum transport. Pitlick (1988) presents bedload transport rates where cross-section mean values are the same order of magnitude as the standard deviation of the measurements. Plots of cross-section variation in transport rate are also useful in determining the 'active' bed width over which movement takes place. In the Bas Arolla channel the active width can be approximated by 0.6-0.7 channel width.

In addition to examining variations in bedload transport at a cross section, variations along the channel need to be evaluated. In coarse bedload channels the use of tagged particles can provide useful information about the nature of bedload transport. The most popular and cheapest way of tagging clasts is by painting or numbering (Thorne and Lewin, 1979). This method has its disadvantages since paint is easily abraded from the surface of the clasts and tagged particles cannot be traced when buried. Both these factors contribute to low recovery rates; on average about 30 % (Hassen et al, 1984). In this respect magnetic tracers, either naturally enhanced (Oldfield et al, 1981) or artificial (Hassen et al, 1984), provide more detailed and reliable information, especially when used with electromagnetic sensors in the stream bed (Ergenzinger and Custer,

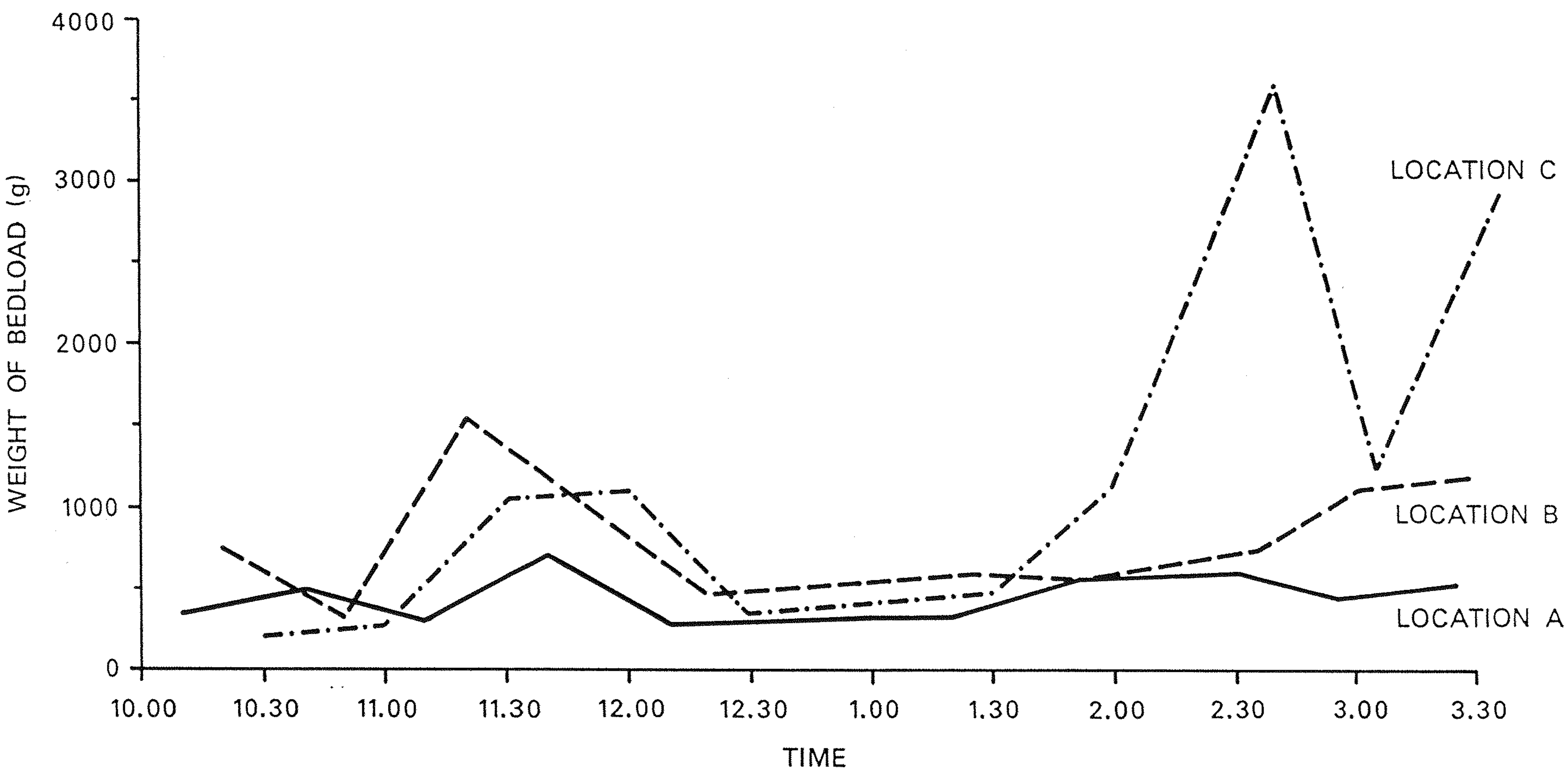


Figure 4.17 Variations in bedload transport over time at 3 mid-channel sampling locations in the same cross-section on the Bas Arolla proglacial stream September 8th 1987.

1983). Radioactively labelled clasts have also been used (Crickmore, 1967) but have not proved popular.

Tracers were selected as a suitable means of investigating bedload movement since they have been proven in a range of sediment transport studies, in coarse-bed streams, (Thorne and Lewin, 1979; Arkell et al, 1983) and are well-suited to studies where only a general picture of bedload dynamics was required. Tracers were used to: (1) Determine the cumulative distance of tracer movement over the study period; (2) Relate travel distance of marked particles, to particle size and shape; (3) To see whether mobile pebbles accumulate at certain points along the long-profile, (4) To assess the overall effectiveness of using pebble tracers in proglacial streams, as well as suggesting improvements on the specific methods used in this project.

Five experiments were carried out on the Bas Arolla (Table 4.1) of between 6 and 15 days duration and over a range of maximum recorded discharges of between 0.61 and 2.74 $m^3 s^{-1}$, using painted and numbered clasts. Tracers were 'seeded' in the channel bed, left for several days, were relocated and their size, distance travelled and location recorded. It is important to seed clasts in the bed because particle protrusion, from the bed, is fundamental to the entrainment process (Andrews, 1983). Together with bedload tracer experiments stream long profiles were surveyed to see if movements could be related to the morphology of the stream gradient.

The results of the experiments indicate tracer recovery was very variable and compared to other studies overall recovery rates are low (14.1%) (Hassen et al.(1984) quote 30% as being an average rate). The reasons for this are a combination of the presence of step pool reaches, high suspended sediment concentrations and

Table 4.1 Description of Bas Arolla proglacial stream tracer studies.

INTRODUCTION POINT	DATE IN	NUMBER CLASTS	COLOUR	RECOVERED	MAXIMUM DISTANCE MOVED (m)	A AXIS OF LARGEST CLAST MOVED (mm)	PERIOD (days)	MAXIMUM DISCHARGE (m3 s-1)
T1 upper main stream	28.5.86	228	red	29 (12.7)	104.0	160	6	1.07
T1 upper main stream	6.9.86	269	red	17 (6.3)	150.9	111	7	1.04
T2 lower main stream	6.9.86	205	yellow	12 (5.9)	123.0	97	7	1.04
T1 lower main stream (A)	28.5.87	115	red	24 (20.1)	234.7	60	13	2.74
T1 lower main stream (B)	28.5.87	115	red	57 (49.6)	258.6	60	15	0.61
	TOTAL	817		115 (14.1)				

Bracketed values are percentages.
Maximum distance moved = maximum cummulative distance moved by an individual tracer clast.

highly turbulent flow; factors which all contribute to hiding of clasts. The apparent clustering of clasts in various locations along the long profile during the 1986 experiments (Figure 4.18) seems related to a local reduction in stream gradient around pools. Given the disruptive influence of storage sites on downstream dispersal, trends in the data are still apparent. Size sorting downstream, based on b axes dimensions, is crudely developed (Figure 4.19b) for the two 1987 surveys. The two surveys were carried out two days apart and show a marked contrast in recovery rate 20.1% as opposed to 49.6%. This is interesting since the greatest flow (2.74 $m^3 s^{-1}$) occurred before the first survey, whereas between the two surveys peak flow was only 0.61 $m^3 s^{-1}$. It is suggested therefore that the bulk of the transport was accomplished by the maximum flow and then intervening flows have removed finer gravels to expose the tracer clasts. This highlights the problem that buried clasts are lost clasts and may leave the experiment for unknown periods (Bathurst et al., 1986b). In this respect magnetic tracing (Arkell et al., 1983) offers a better alternative.

Only limited evidence is available for longer-term movements of tracer clasts because recovery rates are so low and abrasion of clasts very rapid. However re-survey on June 1st 1987 searching for 1986 tracers revealed 10 of the original clasts in the stream bed and another 14 were found up to 1.7 km downstream of the meltwater intake structure. This movement occurred between September 1986 and June 1987 during a period of relatively low flows. Virtually all the clasts were heavily abraded and all buried or imbricated into the bed.

Whilst the experiments produced some useful results there is still considerable scope for improvement in the methodology. The method is cheap and has proved

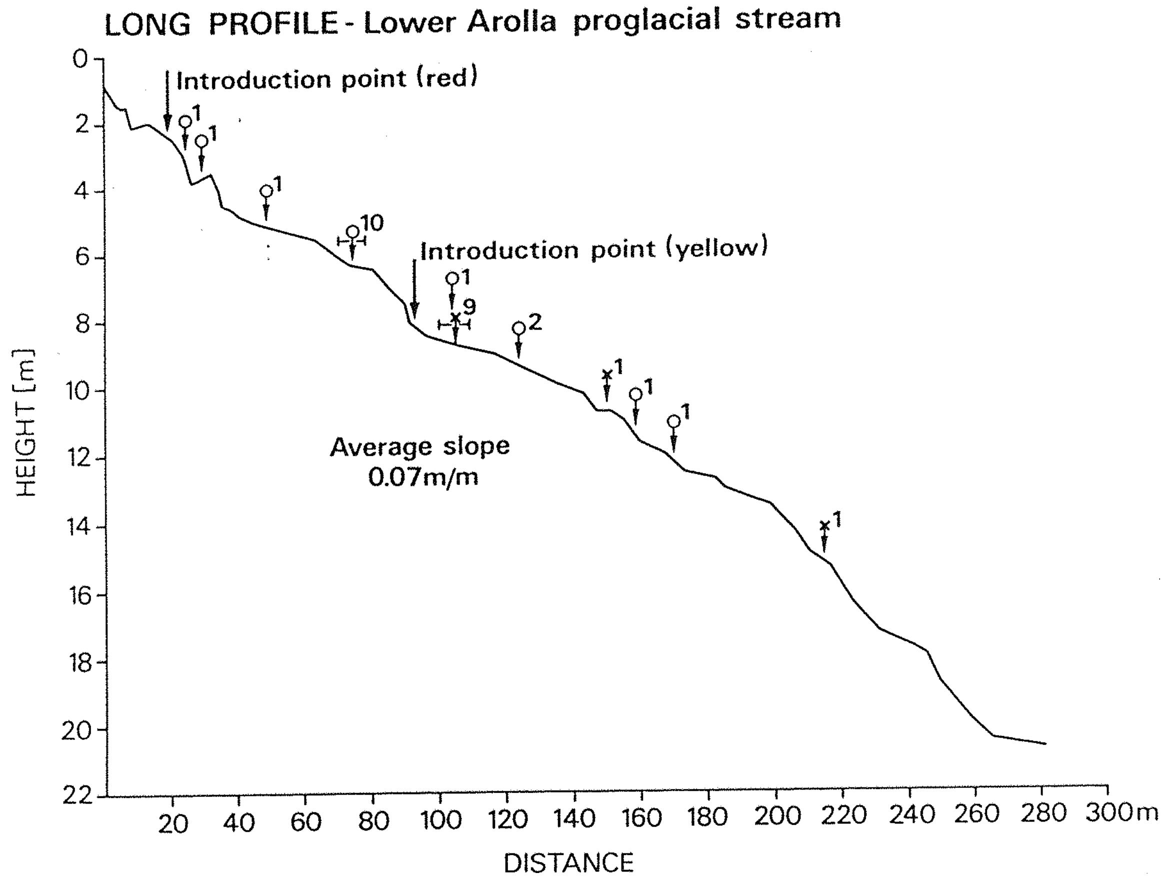
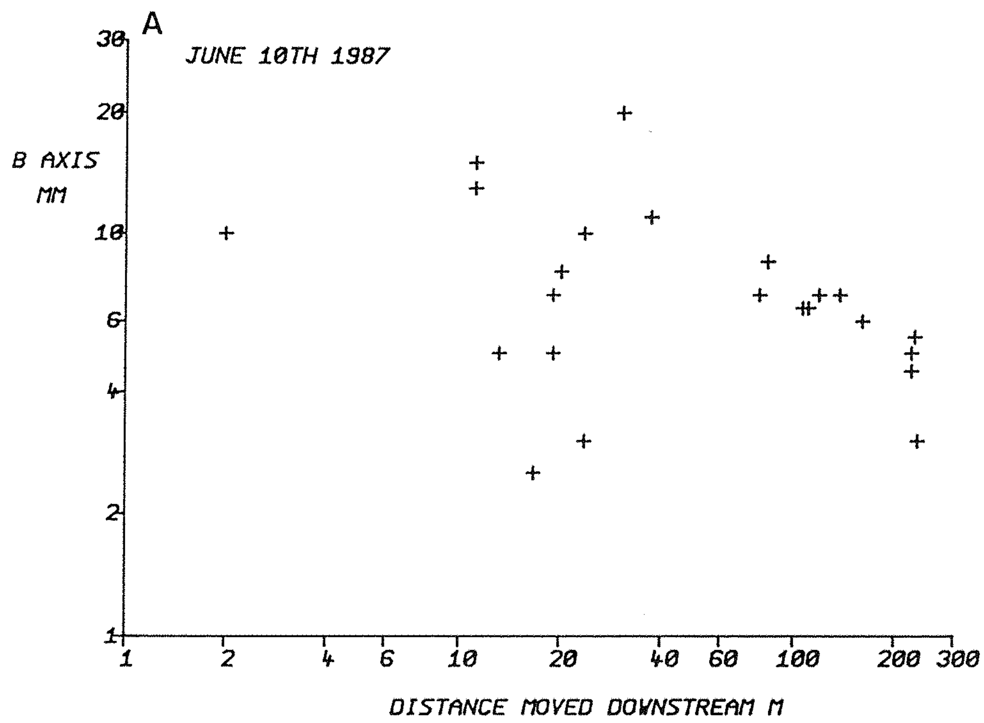


Figure 4.18 Dispersion of tracer clasts along the long profile of the Bas Arolla proglacial stream September 6th 1986. Circles = red tracers, crosses are yellow tracers and the number refers to the number of tracers found at each locality.



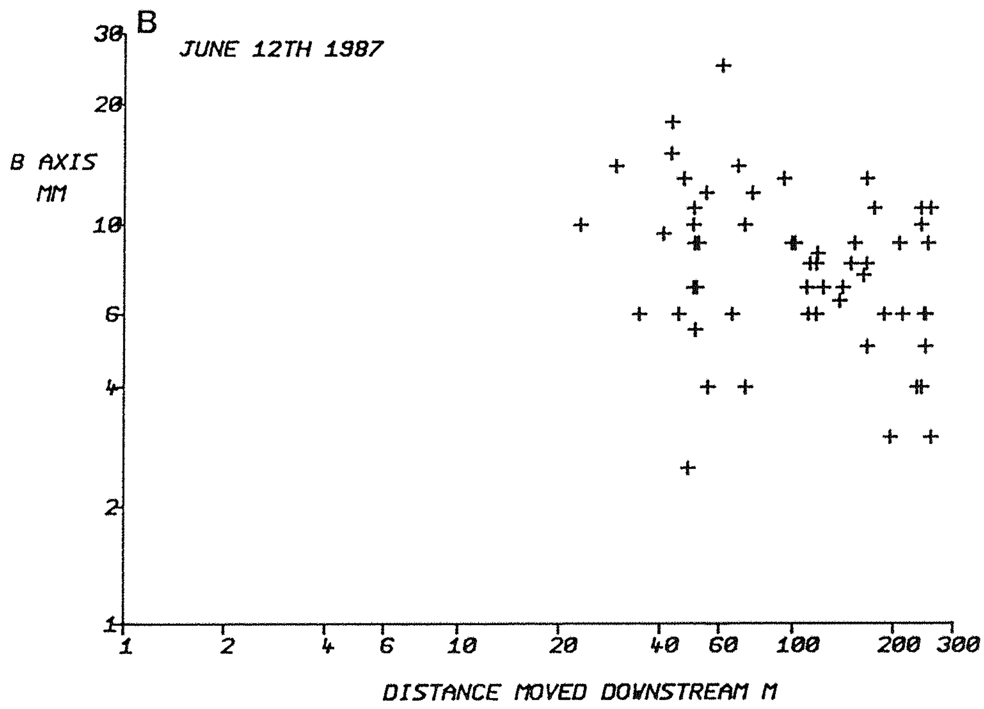


Figure 4.19 Downstream distribution of tracers Bas Arolla proglacial stream, June 10th (A) and June 12th (B) 1987.

effective in streams with coarse gravel beds and low suspended sediment concentrations, but in proglacial streams recovery rates are very low. The most obvious way to improve the number of clasts recovered is simply to introduce a larger number of stones into the stream. This will increase the absolute numbers of stones recovered but not the rate. Stone colour seemed important. Red painted stones stood out from the background better than the yellow stones, which were more noticeable than peach coloured clasts. Clasts need to be distinguishable from the background colour of the bed and through the semi-opaque water of the meltwater stream. The problems of locating clasts in deep-water laden with suspended solids can only really be overcome by using particles with enhanced magnetism detected using a search loop (Arkell et al., 1983) This method is also capable of finding stones buried in the sediment. If the method of manual recovery is still used, search efforts should be at least daily with several sweeps at different stream stages. Searches should be carried out at the lowest possible flows. Tracers should be introduced to the bed during low flow conditions. However to get a true picture of movements daily surveys should be carried out after each peak flow.

Because the recovery rates in this experiment were so low, trends shown in the results cannot be given too much weight. However overall the experiment was important in demonstrating the high sediment transport rates in proglacial streams, and the role of pools as temporary storage sites. For more meaningful results the methodology needs to be improved.

4.5 Summary

Sediment sampling of fluvial gravels in the Bas Arolla proglacial zone (Section 4.2) indicates that an abundance of fine sediment is available at the surface, and along the channel, early in the season. Comparison between surface and sub-surface samples shows a crudely developed armour.

Suspended sediment transport increases by approximately 10% (+/- errors in the estimation of suspended sediment concentration) over the proglacial reach. Cross correlation analysis shows close correspondence between proximal and distal suspended sediment series. This suggests minimal inputs from tributary sediment sources. Tributaries only contribute as a major sediment source during rare intense rainstorms but their overall effect on the mainstream sediment series is probably as 'noise' (Section 4.3).

Bedload transport cannot be satisfactorily predicted using the Schoklitsch transport equation (Section 4.4.1). The main problem is defining the appropriate threshold of motion from representative particle size data. However, defining the threshold critical discharge from field measurements (Figure 4.8a) and applying the equation produces results where the Schoklitsch equation describes, in a very general way, the initial transport phase. In mountain streams it is suggested that the two-phase models of bedload transport (e.g. Jackson and Beschta, 1982) be extended to three phases in order to account for the hierarchy of bed structures associated with the great diversity of sediment sizes in this type of stream.

Field measurements of bedload transport variability (Section 4.4.2) confirm other studies examining transport variations, in that marked temporal and spatial variations exist. The movement of bedload as 'threads' in the main channel is very marked with transport rates varying by a factor of 10 in the space of a metre. Down channel transport is largely controlled by the presence of step pool structures in the bed.

The Bas Arolla proglacial channel acts as an efficient conduit for transport of suspended load from the glacier. Suspended load over the period May to July increased by 580 tonnes giving a sediment delivery of 110%. Inputs and outputs of bedload could not be measured directly but bedload transport observations suggest that channel sources also contribute bedload especially during flood. These conclusions generally support the finding of Duijsings (1985) that sediment delivery in small basins can be 100% but vary markedly over shorter timescales due to fluxes in channel storage. Analysis of channel cross section profile adjustment (Chapter 6) provide indirect evidence of the effectiveness of bedload transport through the Bas Arolla proglacial stream channel.

Chapter 5.

TRIBUTARY SEDIMENT SOURCES AND INPUTS

5.1 Introduction

Fluvial processes, dominate sediment transfer in the valley train and as such are of primary interest in a sediment budget study However, the fluvial domain is not solely restricted to the valley train, it extends in the form of tributaries into the slope domain. The slope domain can be particularly important in providing sediment for tributary transport, or in deliverying sediment directly to the valley train or main channel. For example, Fenn and Gurnell (1987) quote an example of a snow and debris avalanche in the proglacial zone of the Bas Arolla Glacier which caused partial blockage of the valley train leading to channel diversion and regrading prior to and following removal of the sediment plug. Nevertheless, in alpine environments slope processes have little influence on the fluvial system (Caine, 1974). These kinds of events are episodic and unless an event occurs within the period of observation their importance is difficult to evaluate. This is one of the major limitations of short-term measurement schemes in evaluating long-term glacio fluvial sediment dynamics Clark (1987b).

Measuring sediment delivery from tributary sources is a difficult task because the stream network is ephemeral. Small channels are highly mobile and runoff can be channelised, unchannelised, overland and sub-surface. The lack of a definable drainage network suggests that the valley side slopes should be considered in the same way as 'zero-order basins' (Dietrich et al., 1987) since they are relatively steep and consist of a series of topographic depressions (Thorne et al., 1987b). However, if these slopes are defined as 'zero-order contributing areas' then this raises methodological problems in applying a fixed station measurement programme to a dynamic entity. The lower side slopes adjacent to the Bas Arolla proglacial zone (Figure 5.1) are, therefore, best considered as intermittent (fluvial) sediment contributing areas which represent a 'switching zone' between first and zero-order conditions. Recognition of the importance of these basin elements as runoff and sediment production areas is not new (e.g. Newson and Harrison, 1978), but only recently has it been formalised within the zero-order concept. Dietrich et al., (1987) identify three key elements in the study of these landscape units: drainage density, sediment transport and hillslope morphology. This chapter discusses fluvial sediment transport by ephemeral streams in side slope areas of the Bas Arolla proglacial zone and, therefore, addresses the second of these elements (Figure 5.1). The aim is to assess the contribution of tributary sources to the overall proglacial fluvial sediment budget.

The measurement framework for studying tributary inputs was concerned only with material that was contributed either directly to the proglacial zone or into tributary channels. Because sediment eroded from hillslopes is carried downslope to tributaries, where it may be stored in tributary bed and banks, attempts were also made to estimate soil loss which may be important in the sparsely vegetated environment of the majority of the Bas Arolla side slopes Estimates of soil loss were obtained using trough traps. Stream load components were investigated using bedload traps, bedload tracers, channel surveys and suspended sediment sampling. Contributions of sediment by processes supplying material directly to the proglacial zone, without transport in tributaries, were determined from morphological evidence and surveyed on a event basis. Figure 5.2 shows the relationship between the main

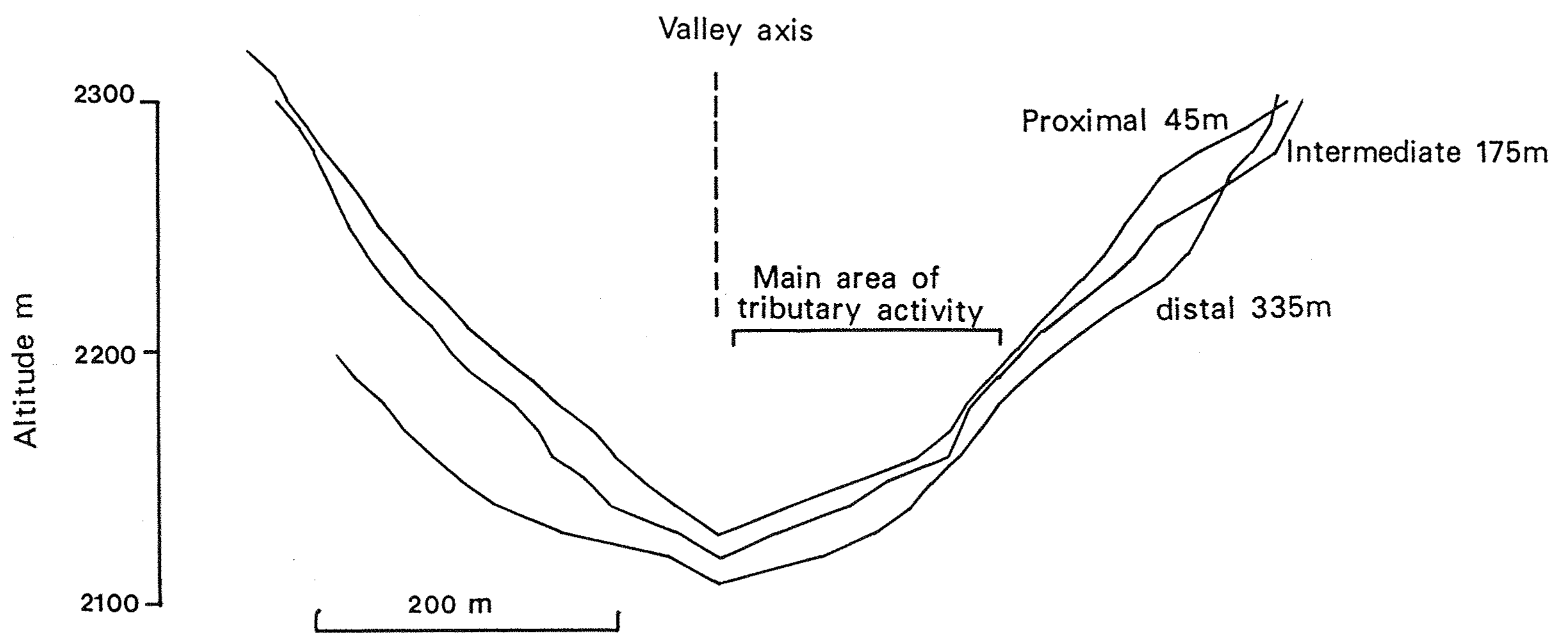


Figure 5.1 Valley cross-profiles, Bas Arolla proglacial zone (distance of profile from glacier snout labelled).

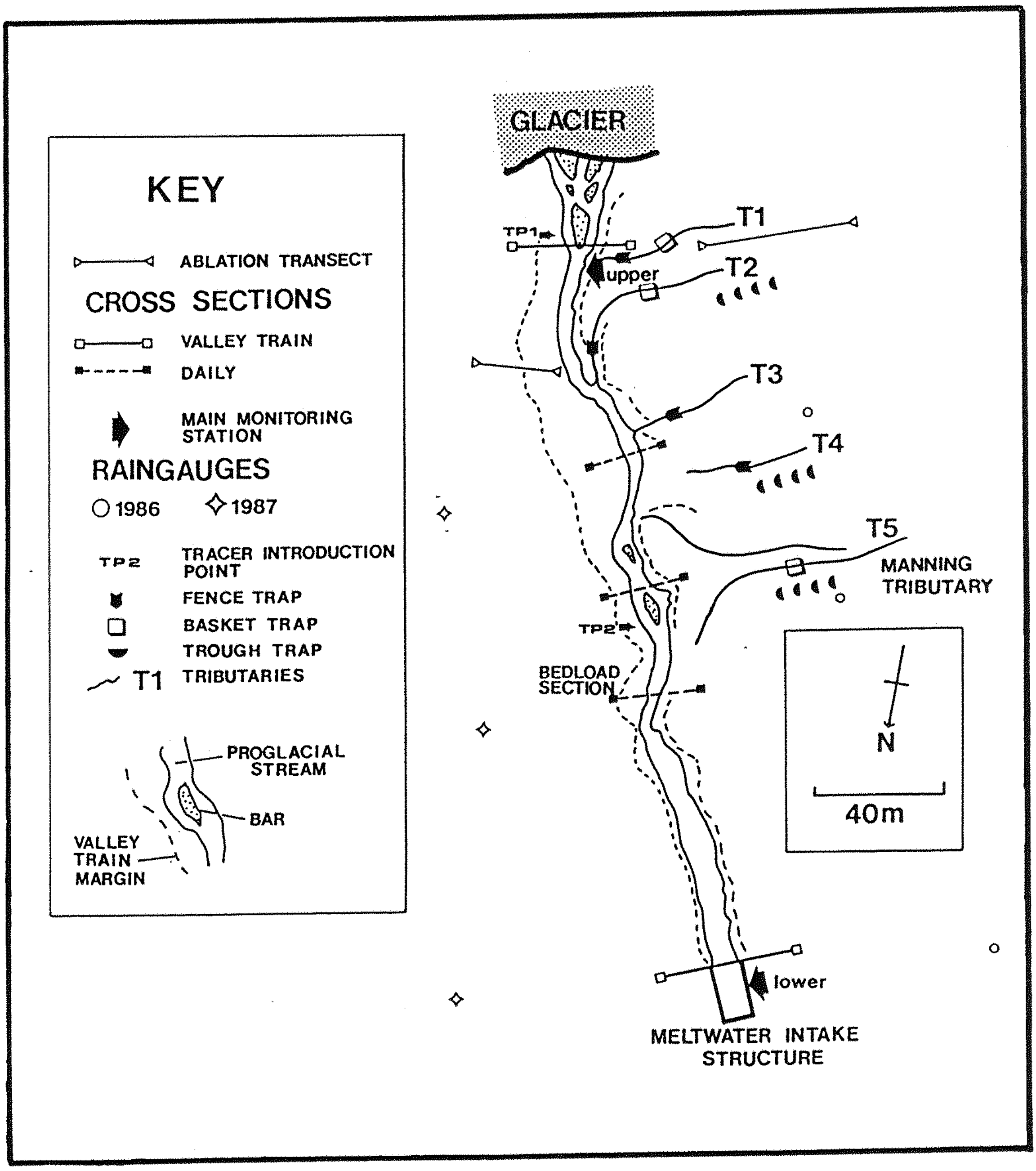


Figure 5.2 Main monitoring and measurement sites in the Bas Arolla proglacial zone 1986 to 1987. The main 5 tributaries are labelled.

channel, valley train and the main tributaries. All tributaries drain from the western slope, where there are 5 principal flow paths (Figure 5.2). All 5 tributaries had some form of bedload trap installed in their beds (Figure 5.2) were sampled for suspended sediment contributions. The position of the raingauges, ablation transects and valley train cross sections monitoring sites are shown (Figure 5.2). The location of transect lines of small trough traps (4 traps per transect) used to collect slope sediments are also shown (Figure 5.2) All were located on the western side slope.

This chapter is divided into two parts. Section 5.2 discusses within channel tributary sediment sources and Section 5.3 considers sediment supply from adjacent slopes.

5.2 Tributary channel sources

This section considers within tributary sediment sources, beginning with a short discussion of runoff from tributary areas (Section 5.2.1), followed by a consideration of the fluvial sediment transport components: suspended sediment (Section 5.2.2) and bedload (Section 5.2.3). Changes in the tributary channel morphology are discussed in the final Section (5.2.4) since these largely reflect changes in sediment storage.

5.2.1 Streamflow and runoff

Runoff in the proglacial zone can be generated from three main sources: rainfall, direct glacier melt and snowpack ablation. In terms of the tributaries, snowpack ablation and rainfall are the main sources of streamflow. Figure 5.3 shows the rainfall records for 1986 and 1987 and the ablation measurements for 1987. In 1987 monitoring of rainfall and ablation was only carried out until the end of July. Rainfall (Figure 5.3a) shows marked contrasts for the two years, 1987 was much wetter than 1986 in the early ablation season, particularly up to the middle of July. The highest rainfall total in both observations periods was recorded on the 7th July, 1986. In 1987 the highest recorded rainfall was on the 8th of June. August 1986 was the wettest month in the observation periods (no data are available for August 1987).

A second contribution to runoff is from ablating snow. Snow accumulates to its greatest depth along the margins of the valley train and at the foot of the main valley side slopes. Valley train snowpacks contribute directly to the main streamflow, and playing little role in tributary hydrology. Ablation measurements on two valley side snowpacks in 1987 are shown in Figure 5.3b.

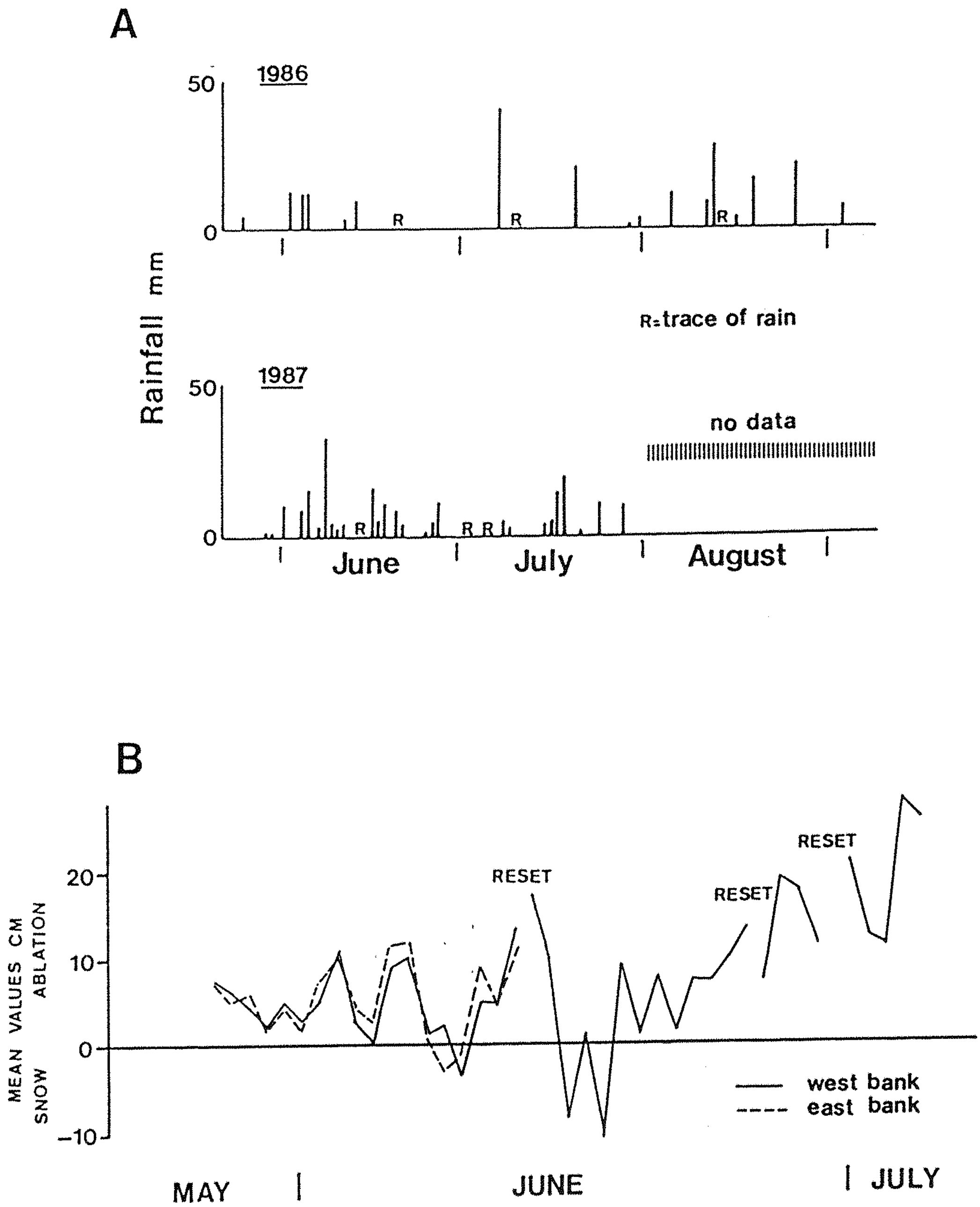


Figure 5.3 Rainfall and ablation measurements Bas Arolla proglacial zone. A) Rainfall - event based observations based on a 3-gauge-mean 1986 and 1987. B) Ablation - measured in east bank snowpack (n = 6-7) and west bank snowpack (n = 5-10) (Figure 5.2).

The pattern was one of peaks and troughs of high and low ablation in early May and June followed by a period of erratic melt and accumulation days from mid to late June. From late June ablation continued at an increasing rate as temperatures increased. Virtually all snow was melted by the 5th of July. The erratic nature of the ablation rate and the persistence of the snowpack can be partly explained by the overcast rainy conditions of the early 1987 ablation season.

The nature of the streamflow varied according to the source of runoff. Runoff from snowpacks was minimal (0.05 $m^3 s^{-1}$, measured by salt dilution gauging) and transported very little sediment, with typical suspended sediment concentrations of 25 to 50 mg l^{-1}. Ephemeral streams draining the lateral moraines after heavy rainfall, had much greater discharges (up to 0.225 $m^3 s^{-1}$, salt dilution gauging) and higher but very erratic suspended sediment loads (250 to 8470 mg l^{-1}).

Attempts to gauge tributaries using the relative salt-dilution method (Østrem, 1964; and Newson and Harrison, 1978) were of limited success and only a few accurate discharge measurements were obtained. Problems arose due to experimental limitations and the nature of tributary streamflow. Experimental problems included incomplete mixing (leading to complex multipeak records, non-reproducibility in gauging runs and conductivity calibration problems) and very shallow and dividing flow. Tributary flows were sporadic and often lasted only a few hours or days in many of the minor tributaries. Furthermore the path of flow changed rapidly - new streams emerged and old ones disappeared regularly. These ephemeral characteristics of the streamflow prevented the establishment of permanent gauging sites and, therefore, continuous estimation of discharge. One stream, referred to as the 'Manning Tributary' (Figure 5.2) was the only tributary with a reasonable duration of flow but the experimental problems described above prevented a stable discharge relationship being constructed.

An indication of the importance of tributary streams, on the 27th and 29th of June 1987, when accurate determinations of tributary discharge were possible, suggests that when they are flowing tributaries normally contribute less than 0.8% of the water yield but this can rise as high as 16% early in the season. Because the major emphasis of this work is related to sediment discharge, gauging of tributaries only becomes important when high discharges occur in association with high sediment transport. Spatial suspended sediment concentration samples obtained during daily site visits and as part of the regular spatial sampling programme largely accounted for these occurrences and showed the 'Manning Tributary' was the most important source of suspended sediment.

5.2.2. Tributary suspended sediment dynamics and loads

The ephemeral, flashy nature of proglacial tributary streams coupled with sporadic variations in sediment concentration make a quantitative evaluation of sediment load difficult. The Manning Tributary was studied in order to determine suspended sediment dynamics and concentration variations. This was the main tributary draining the west valley slope (Figure 5.1) and provides the greatest tributary contribution to both suspended sediment and discharge in the main proglacial stream. The record of suspended sediment concentration for this tributary is shown in Figure 5.4a. Plotted are daily determinations of suspended sediment concentration and the range of daily values as estimated using a pumping sampler. Detailed observations of suspended sediment concentration are shown in Figure 5.4b. Flow occurred in this tributary during the periods 27th May to 20th June, 24th June to 5th July and 12th July to 18th July after which flow to the channel was cutoff upslope. A

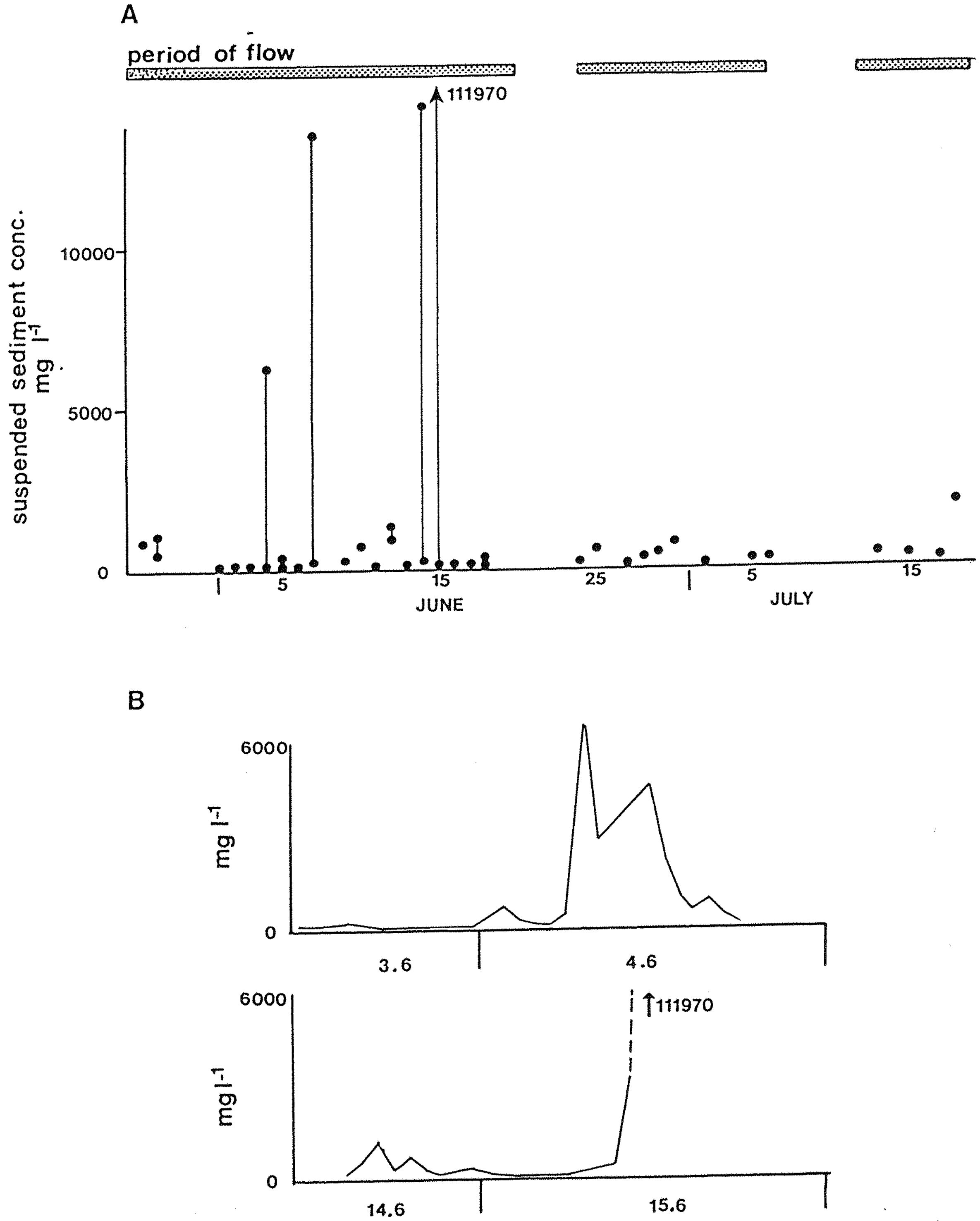


Figure 5.4 Tributary suspended sediment dynamics Manning Tributary (T5) 1987. a) Seasonal flow history and suspended sediment concentrations. b) Detail of record.

complete hourly series of suspended sediment concentration as a result of these temporary breaks in flow and the silting up of the pumping sampler intake.

The suspended sediment series in Figure 5.4 suggests generally low suspended sediment concentrations (200-300 mg l^{-1}) with occasional observations exceeding 1000 mg l^{-1}. However, during a few exceptional events very high concentrations existed (e.g. on the 4th, 7th, 14th and 15th of June). In each of these cases high concentrations were associated with high rainfall. Figure 5.4b plots the observations for two of these days in more detail. It is evident that there were extreme temporal variations in suspended sediment concentration. Extreme variations of this type were probably associated with debris flow activity and small-scale slope failure in the upper reaches of the tributaries (see section 5.3). The exceptional concentrations observed on the 15th of June were certainly associated with debris flows entering and travelling down the tributary channel. Figure 5.5 illustrates the heavily sediment laden flows on this day. The sediment contribution from this tributary on the 15th of June supplied 52% of the estimated total suspended load from the tributary for the study period.

Because flow in the Manning Tributary was sporadic and problems with salt dilution gauging prevented detailed discharge estimates, a continuous record of flow in the tributary could not be established. Nevertheless because the period of flow duration is known, this can be multiplied by minimum and maximum discharge estimates (0.073 to 0.225 m^3 s^{-1}) to give estimates of the range of daily water yields. These estimates can be combined with mean suspended sediment concentration estimates to arrive at an estimate for tributary suspended sediment output of between 97.2 to 299.6 tonnes. The total suspended sediment yield from all the tributaries for the same time period (estimated by

Figure 5.5 Manning Tributary (T5) June 15th 1987.
A) Upper photograph - shows the lack of mixing between the tributary and mainstream flows at their confluence.
B) Lower photograph - Shows the very high tributary sediment discharge and highly turbulent flow on this day.

multiplying mean observed flow with mean suspended sediment concentration in each case) was 98.2 to 302.6 tonnes implying that, the Manning Tributary contributed roughly 99% of the total suspended sediment load from the tributaries. Estimated for the entire observation period from 25th May to 30th July the total load in all tributaries was 103.6 - 319.2 tonnes. These estimates should be treated with caution because of their crude method of derivation, but they are probably at least of the right order of magnitude.

Variations in tributary suspended sediment discharge may also manifest themselves in the spatial pattern of suspended sediment concentrations (Gurnell, 1982). Tributaries will have a 'dilution' or 'concentration' effect on mainstream suspended sediment concentration, the magnitude of which depends on the relationship between the concentration of suspended sediment and volumes of water associated with the two water sources. Figure 5.5a, or example, clearly shows the effect of highly turbid tributary water being discharged into an essentially 'clear' mainstream flow, but the dilution effect is such that no marked increase in sediment concentration is recorded 75 metres downstream.

Suspended sediment is carried by the streamflow which is generated from several sources. The principal sources of streamflow in the Bas Arolla proglacial zone are glacial (supraglacial, englacial and subglacial), snowmelt, rainfall-runoff, and groundwater. The sources which are of principal importance in transporting suspended sediment are subglacial, snowmelt and rainfall-runoff contributions. In describing the water sources referred to in this section a simple nomenclature is observed. Sources are designated Glacial (G) or Non-Glacial (NG); glacial sources are indicated by supraglacial ((sup)), englacial ((eng)) subglacial ((sub)) and ice marginal ((im)). Non glacial sources are snowmelt ((s)), rainfall ((r)), or groundwater ((g)). Flows are also described as clean (c) or dirty (d) depending on the relative suspended sediment concentration. Therefore an ephemeral non-glacial tributary flowing after heavy rain and transporting high sediment loads would be NG(r)d. Distinctions of this kind are important since the source of water influences the type of material transported and the mechanism of transport (e.g. very large calibre load can be transported in relatively small closed englacial passageways, which have smooth ice-walls and where flows are under hydrostatic pressure. Figure 5.6 shows the results of spatial surveys of suspended sediment concentration on the Bas Arolla proglacial stream network in 1986 (May 28th to September 3rd) and 1987 (June 8th to July 30th).

In 1986 (Figure 5.6) the trend in suspended sediment concentration variations in late May and early June was one of great variability with most of the discharge contributed from G(sub)d, G(eng)c and G(sup)c sources. There were minor contributions from NG(r)d runoff in late May, but this was followed by a 'lull' in suspended sediment discharge from all sources in mid June. From late June until early July suspended sediment concentrations increased greatly. The glacier appeared to be flushing out sediment from G(sub)d and G(sup)d sources. Tributaries were also contributing sediment during this period, manly NG(r)d sources, with great variability in the spatial pattern. By July 4th most of the tributaries were discharging from NG(s)c and NG(g)c sources with high sediment concentrations restricted to the main stream channel. This pattern was punctuated by a major flood on the 6th July following which the channel pattern was greatly altered and suspended sediment concentrations were reduced. From this time on the drainage pattern was progressively rationalised and suspended sediment concentrations were reduced reflecting the change from G(sub)d to G(sub)c waters.

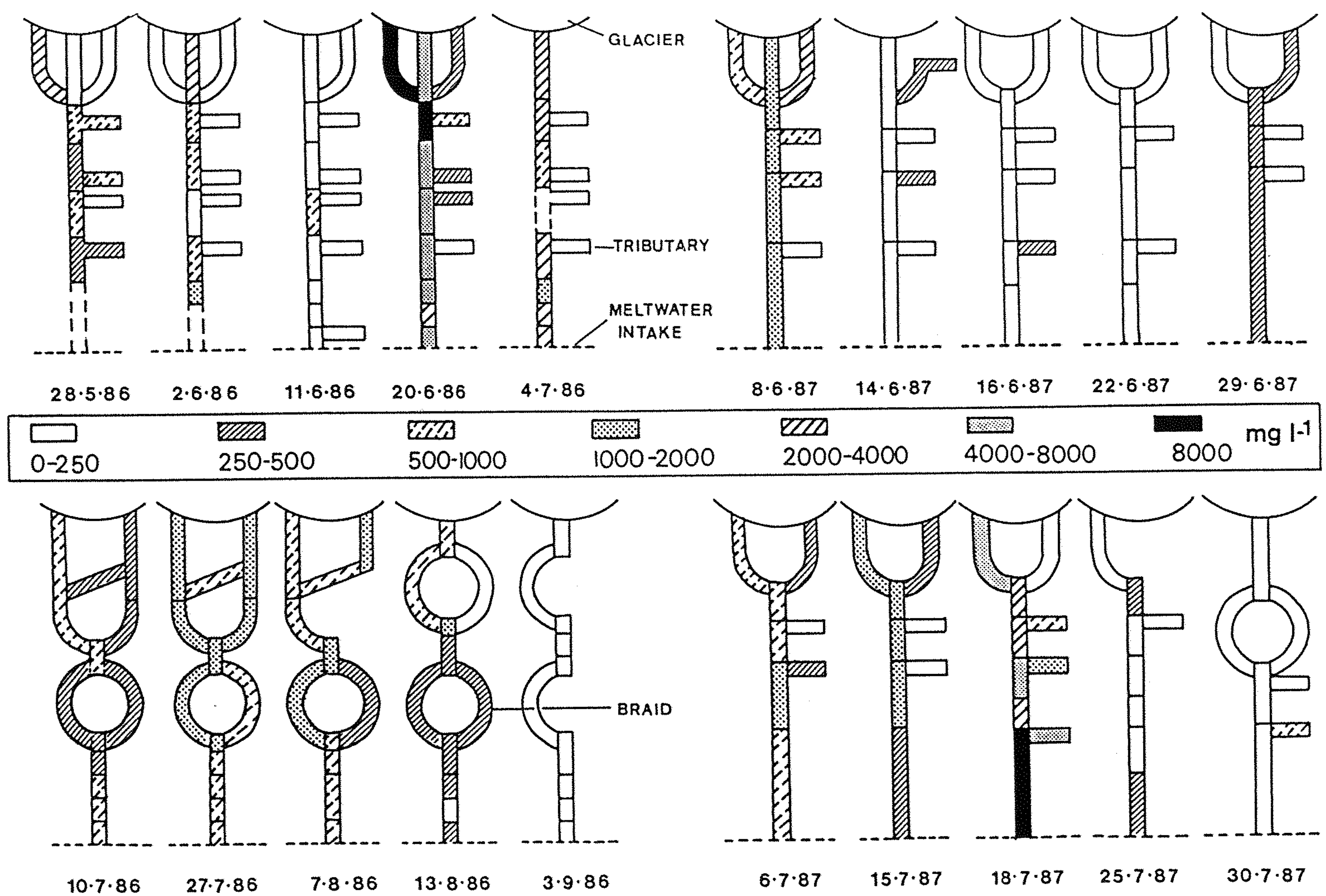


Figure 5.6 Spatial surveys of suspended sediment concentration on the Bas Arolla proglacial stream network 1986 and 1987.

The shorter observation period in 1987 (Figure 5.6) showed similar trends. Early June was characterised by variable patterns of suspended sediment concentration, with sediment being contributed from G(sub)d G(im)d glacial sources and from NG(s)d NG(r)d non-glacial runoff. From mid to late June discharge remained low and little suspended sediment was contributed except from NG(r)d sources on the 14th June and G(sub)d waters on 29th June. In early and mid July sediment concentrations increased with G(sub)d source becoming particularly important. A large flood event between the 15th and 18th July produced extremely high suspended sediment concentrations with virtually all runoff sources contributing sediment except for G(sup)c which was probably runoff generated directly from rainfall. Following this event flows were substantially reduced, the channel pattern simplified and suspended sediment loads generally declined.

Downstream trends in suspended sediment concentration are not readily apparent from Figure 5.6. This is probably because such patterns are disrupted by tributary inputs of additional water. These inputs are often minor and contribute little to the total load but may locally disrupt suspended sediment concentration trends (e.g. on July 18th 1987 the lower tributary had high discharge and high suspended sediment concentration which may account for the high concentrations encountered in the lower reaches of the mainstream channel). Tributary sediment discharge on June 15th is discussed in Chapter 4, showing that under these extreme conditions tributary contributions may be important. It is also noticeable that channel segments with high suspended sediment concentrations in the snout zone are restricted to the immediate vicinity of the snout suggesting rapid deposition of much of the suspended sediment load. Furthermore, show marked suspended sediment concentration contrasts in tributary sources have little effect in altering main-channel suspended sediment concentration patterns because their relatively small discharges contribute little suspended sediment load in absolute terms (i.e. a 'drop-in-the-ocean' phenomenon). The coarse spatial resolution of the data collection and thus the plots in Figure 5.6 may also conceal downstream trends.

Stream braiding (illustrated by the circular 'roundabout' channel segments in Figure 5.6) in some cases appeared to influence sediment concentration. Surveys on the 27th July and 7th August 1986 both showed contrasts between right and left sides of the braid. These differences are thought to be related to the hydrodynamics of sediment transport in adjacent channels. Certainly some of the variation in pattern is likely attributable to morphology influencing the hydraulics of flow and sediment transport. This is confirmed by Polyanchuk (1982) who found that variations in suspended sediment concentration in both cross-section and long profile could be explained by the proportion of suspended material greater than 606 microns (i.e. the coarse fraction of the suspended load is sensitive to the hydraulics of flow).

In summary, the ablation season pattern in suspended sediment concentration appears to sub-divide into four phases:

1) *Early season* - Characterised by low to intermediate sediment concentrations with variations caused by inputs from G(sub)d, G(im)d, NG(r)d and NG(s)d sources.

2) *Quiescent phase (lull)* - generally low concentrations and low discharges with minor contributions from both glacial and non-glacial sources.

3) *Flood / flushing phase* - a period of a few days where discharge and sediment concentrations are very high and all sources seem to contribute to the sediment load. Accompanying this phase is usually a change in channel pattern.

4) *Post-flood to late season phase* - This is a period where concentrations are substantially reduced, there are few tributaries and the channel pattern is rationalised, often becoming single-thread.

The variability in the coefficient of variation for each group of spatial suspended sediment survey data over time reflects a similar pattern to the four phase summary described above (Figure 5.7). Data for 1986 (Figure 5.7a) are plotted for the entire ablation season, and show large variations early in the season (Phase 1.) followed by a decline (Phase 2.) and a slight rise again in August (Phase 4.). This can be explained as follows. Early in the ablation season multiple sources contribute to the proglacial stream producing large variations in suspended sediment concentration. Progressive rationalisation of the drainage net both glacially and proglacially resulted in contributions mainly from G(sub)d or G(sub)c. The slight rise in variation in August was related to the increased frequency of rainstorms during this period which triggered tributary sediment discharge i.e. NG(r)d or even G(im)d. The 1987 plot (Figure 5.7b) is of shorter duration but shows greater detail and identifies the influence of the large flood event during Phase 3. Again variation was high early in the season, due to multiple source contributions (Phase 1.). This was followed by a period of low variation (Phase 2.) and then a second peak in variation (Phase 3.) associated with the flood event when water was generated from a variety of sources. The absence of a flood peak in the 1986 plot (Figure 5.7a) is more a function of sampling since the flood event which occurred was short lived and so was not coincident with a suspended sediment survey.

Although this simple scheme matches the observations reasonably well it must be acknowledged that these are observations of suspended sediment concentration and these values should ideally be combined with discharge

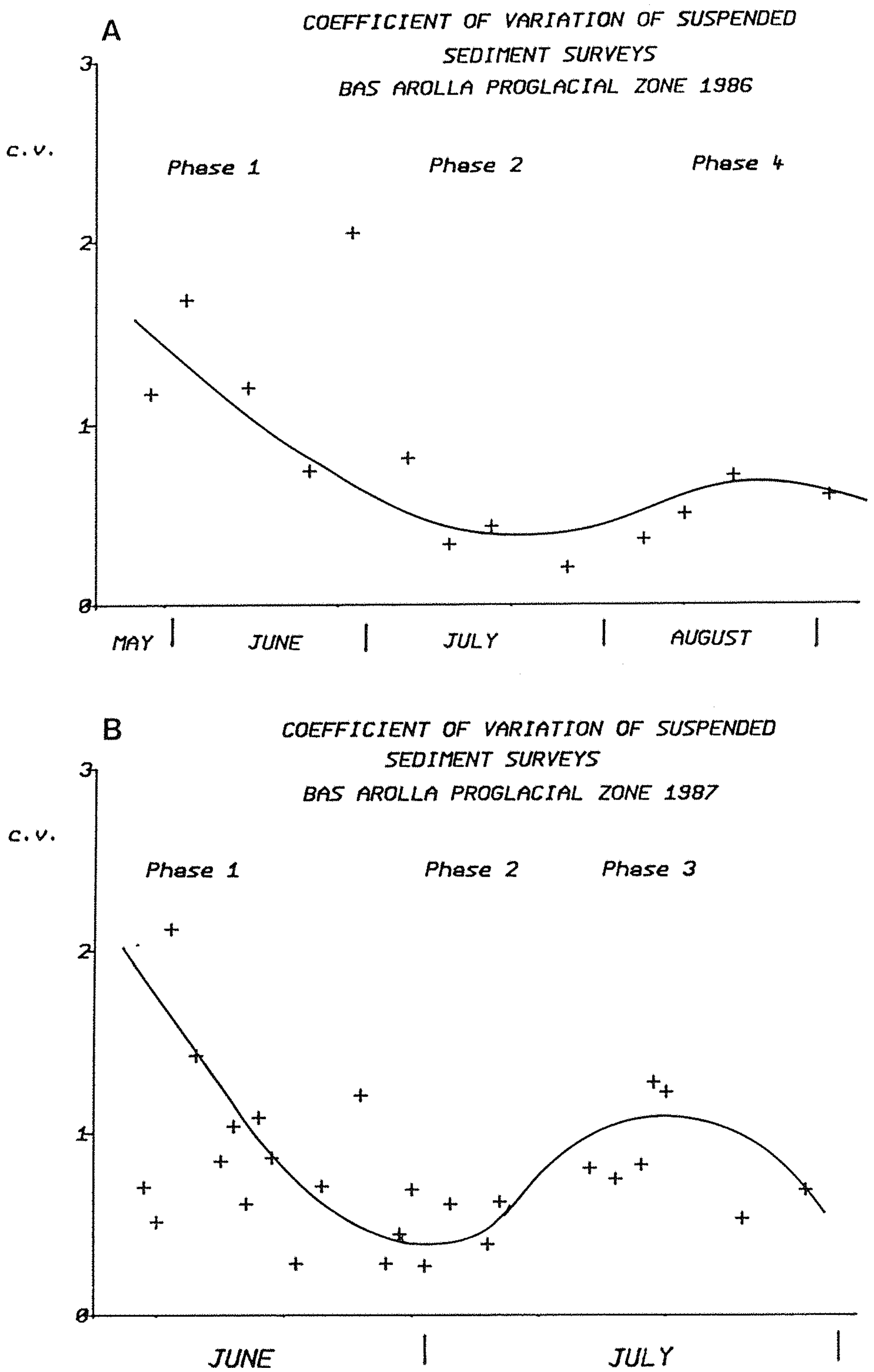


Figure 5.7 Patterns in the coefficient of variation for the suspended sediment surveys on the Bas Arolla proglacial stream network 1986 (A) and 1987 (B).

measurements if patterns in suspended sediment load are to be considered. There were also some problems related to sampling. Firstly, the ephemeral nature of tributary streams meant their effect between surveys cannot easily be accounted for. Secondly, sources of sediment and water were not always apparent - it is often impossible to accurately separate G(sup) and G(eng) water sources or, for that matter, G(sub)c and G(eng)c. Thirdly, local within-channel variations exist in suspended sediment concentration and a single sample may not adequately account for this (e.g. sampling the mainstream in the centre of flow compared to sampling of steep, very shallow tributaries may lead to an over-estimation of tributary concentrations since the bottom load may be sampled). Finally, one of the dangers of spatial sampling is sampling different parcels of water as they travel downstream (Metcalf, 1986). To some extent this has been accounted for in this study because suspended sediment surveys were completed in 4-5 minutes in comparison with the 6-8 minutes typically taken by a salt pulse in travelling from the upper to the lower site (on one occasion a sediment pulse was observed travelling down the at the time of survey and it was possible to stay in advance of the pulse and complete the survey). Because the sampling procedure progressed at rate comparable to the rate at which suspended sediment moved downstream this method approximates a Lagrangian sampling scheme of the type specified by Metcalf (1986). In addition, continuous turbidity records suggest that for most of the time suspended sediment concentration is relatively stable over such a small timescale. Nevertheless, care is still needed in interpreting the data since variations can be caused by short timescale sediment pulses.

Daily spatio-temporal patterns of suspended sediment concentration and electrical conductivity were investigated on June 25th (at 7 sites) and July 6th 1987

(8 sites). Figure 5.8 shows the results for the 25th of June, for which marked differences existed in suspended sediment concentration between tributary (sites 5 and 7) and mainstream sites. All sites showed initially low concentrations until approximately 1200h after which site 5 started to contribute suspended sediment, increasing to a peak concentration of 750 mg l^{-1} at 1500h. At the same time tributary 7 showed a rapid rise in suspended sediment concentration to a peak of 1300 mg l^{-1} at 1600h. This then declined to background levels by 1900 - 2000h. At all other sites (the mainstream ones) suspended sediment concentrations were low (100 mg l^{-1}). A general rise in suspended sediment concentration at all sites from 1100h is evident but this is slight. The lack of response in mainstream concentrations to tributary inputs suggest low discharge in the tributaries which emphasises their minor role overall contribution to the sediment load.

Electrical conductivity measurements taken simultaneously with the suspended sediment determinations at the 7 sites allows discrimination between contributing water sources. All mainstream locations, except site 2, showed relatively high electrical conductivity values. Tributary 7 also showed high values. Sites 2 and 5 however, had much lower conductivities. At this stage in the season site 2 was thought to be fed from englacial and supraglacial sources (G(sup)c and G(eng)c) with site 5 fed largely by snowmelt. Therefore, both water sources had little contact with rock or minerals and their electrical conductivity was low. All sites exhibited a strong diurnal cycle in electrical conductivity with lowest values at around 1700h, which corresponds with the peak of discharge and reflects the dilution effect of the extra water.

Figure 5.9 for July 6th represents a reversal of the situation presented in Figure 5.8, since mainstream

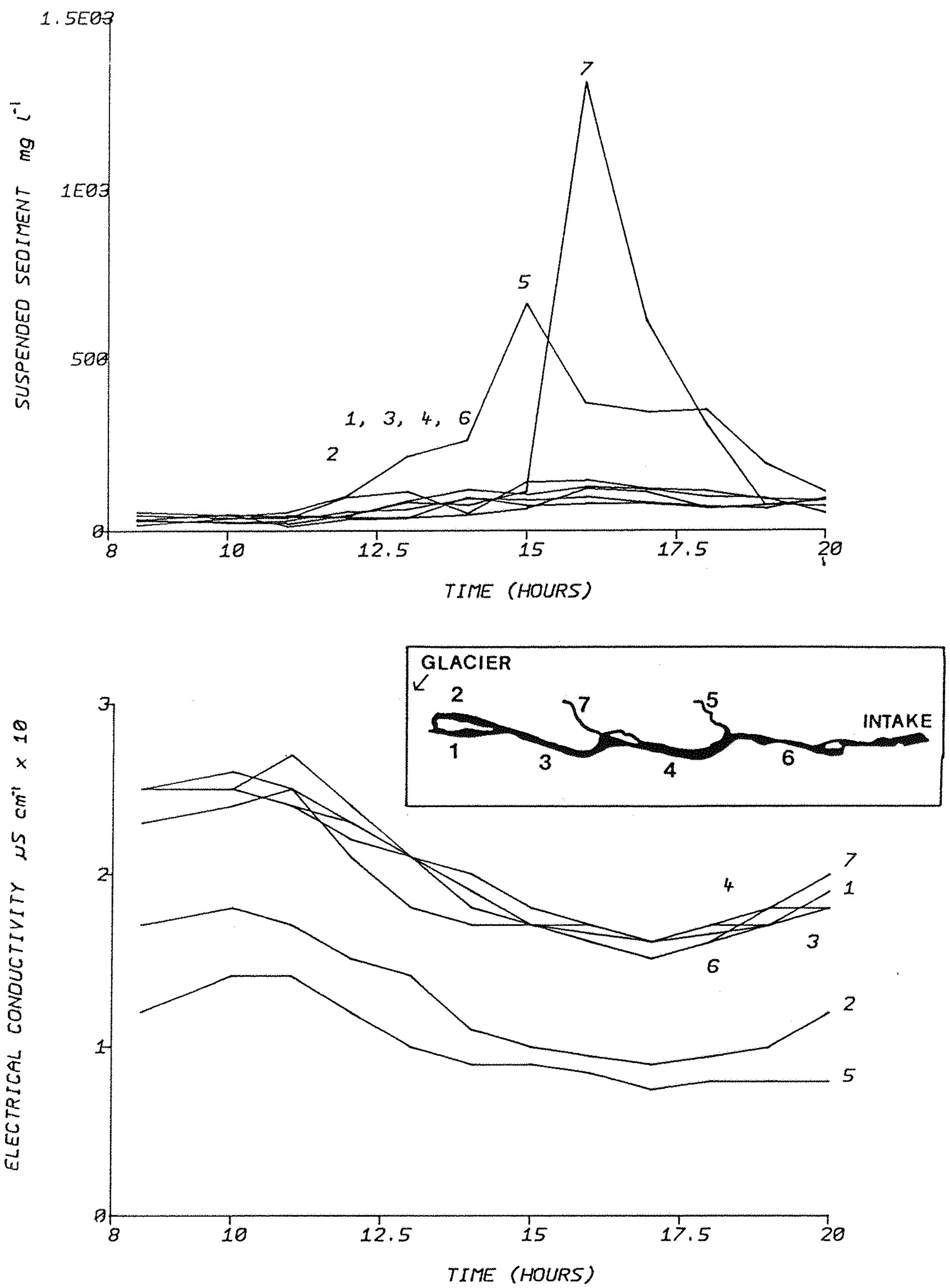


Figure 5.8 Daily spatio-temporal patterns of suspended sediment concentration and electrical conductivity June 25th 1987. The 7 sampling sites shown in channel pattern inset.

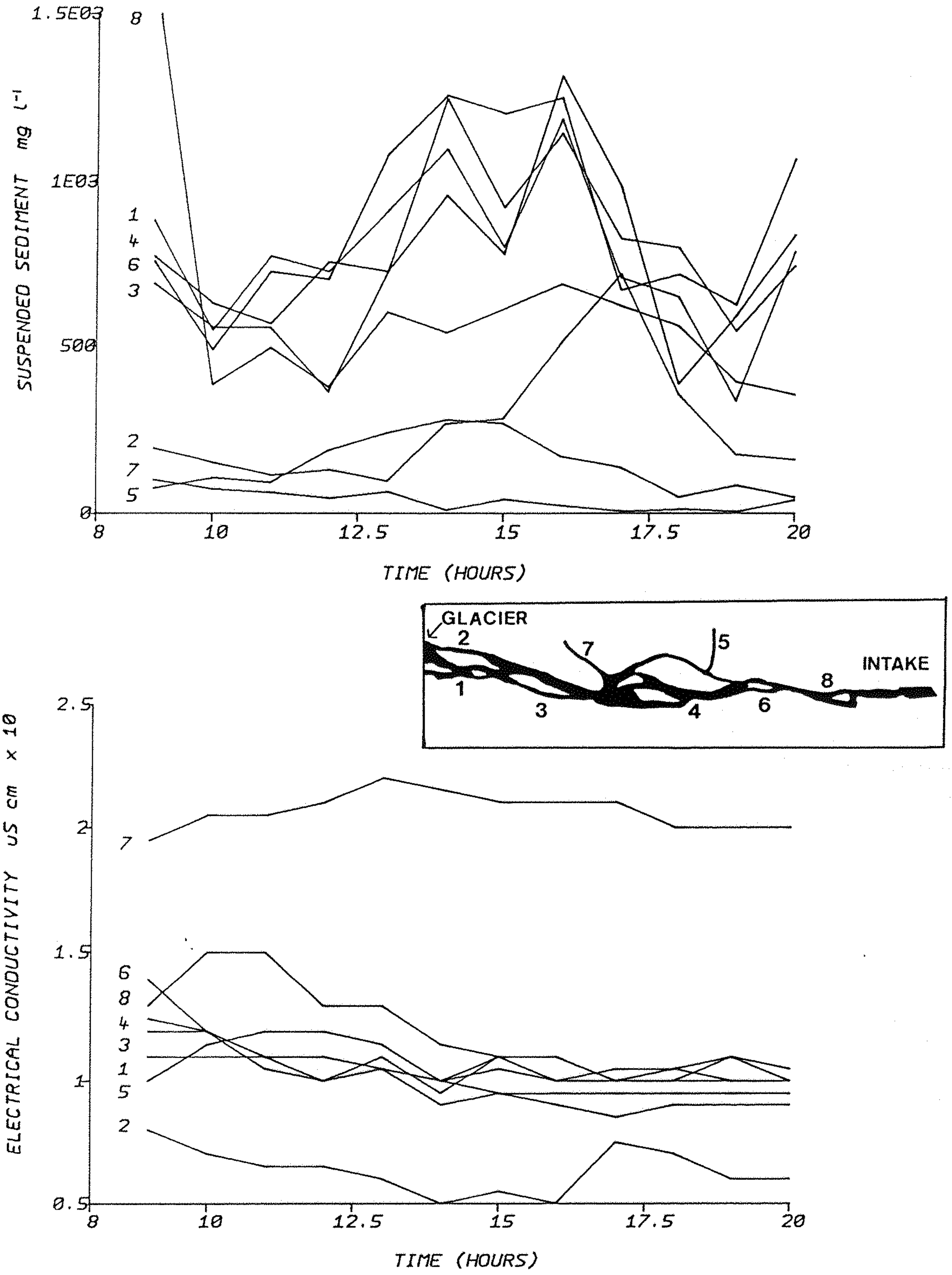


Figure 5.9 Daily spatio-temporal patterns of suspended sediment concentration and electrical conductivity July 6th 1987. The 8 sampling sites shown in channel pattern inset.

suspended sediment concentrations were higher than those of the tributaries (sites 5 and 7). Although there was considerable variations in at-a-site mainstream concentrations all sites apart from site 2 showed parallel shifts in concentration with a common peak value occurring around 1500 - 1600h. The record at Site 2 was an anomaly since it produced a quite separate and distinct suspended sediment concentration peak at 1700h. Field observations showed that the peak in sediment concentration corresponded to peak discharge because some of the sediment laden water passing site 1 was re-routed past site 2. This suggestion was confirmed in the conductivity measurements which also show a peak reflecting the presence of site 1 water at site 2.

Electrical conductivity measurements on 6th July were generally lower than on June 25th and showed no marked diurnal rhythm (Figure 5.8). A slight decline in conductivity was evident at most sites throughout the day, with the only marked deviation being in the record at site 2 when (as mentioned above) some of the site 1 water was diverted past this station. Tributary (T3) site 7 has the highest conductivity values and may be largely fed by groundwater sources. At this time in the season water from melting snow is unlikely to be of significance in the generation of tributary flows.

Data from the two survey days, June 25th and July 6th, demonstrate the nature of suspended sediment contributions from tributary and mainstream (glacial) sources. However, high suspended sediment concentrations in tributaries have little observable effect on mainstream suspended sediment concentration trends. Part of the reason for this disparity is that suspended sediment concentration sampled in shallow, steep tributaries incorporates a proportion of the 'bottom load' (coarse fraction). This is not sampled in the mainstream since the greater flow depth allows the components to be sampled more accurately in the greater depth of water. In addition, on entering the main channel tributaries deposit small fans of sand which supports the idea that coarse sediment although reaching the main channel in suspension may then continue transport as bedload or even go into temporary storage. Electrical conductivity measurements are also useful since they help to explain the origin of various waters and therefore aid in the interpretation of the sediment data.

5.2.3 Tributary bedload dynamics and loads

Bedload movement in tributaries assessed using three measurement techniques: fence traps, pit traps (Stott, 1987) and tracers (Thorne and Lewin, 1976) (see Chapter 2. for further details). The aims of these measurements were two-fold:

1) To estimate bedload yields from tributaries.
2) To demonstrate whether the coarse sediment transport system of the tributaries is connected with bedload movement in the main channel.

Table 5.1 summarises the results of bedload trapping on the principal tributaries of the Bas Arolla proglacial zone (Figure 5.2). Of the two types of trap used, pit traps were the more effective. Fence traps, although easily erected, became clogged with organic debris at high flows leading to upstream ponding and eventual failure, and the loss of trapped sediment. Figure 5.10a shows a fence trap working effectively (Tributary T4, June 22nd 1987) however Figure 5.10b shows a fence trap which failed due to clogging by organic debris (Tributary T1, July 17th 1987). Basket traps produced more consistent yields but provided only minimum estimates because the capacity of the trap was exceeded (buried) in some instances (e.g. Tributary T5, June 12th 1986). Furthermore, because the history of sediment accumulation is unknown, further uncertainty surrounds trap estimates. The data in Table 5.1 reflect the fact that bedload movement in tributaries only usually occurs during 'storm' events and bedload transport is temporally and spatially variable. In 1986 the total bedload yield for the period July 20th to September 13th in tributaries T1, T2 and T5 totalled 0.234 tonnes (0.5 tonnes for the ablation season). Tributaries T3 and T4 showed no appreciable sediment movement. Assuming tributary stream flow was generated from rainstorms, doubling of the half season tributary bedload yield to

Table 5.1 Sediment trap measurements from Bas Arolla Tributaries

Tributary	Trap Type	Dates [1]	Yield Kg	Packing Density	Dominant Sediment	D50 mm	Comments
T1	Basket	20/7 - 13/9/86	1.2		sandy gravel	1.8	
		13/9 - 6/6/87	7.0		sandy gravel		
T2	Basket	20/7 - 13/9/86	1.15		sandy gravel	0.5	
		13/9 - 22/6/87	1.5		sand	0.4	
		22/6 - 27/7/87	1.0		silt	0.05	Disturbed ?
T5	Basket	20/7 - 29/7/86	0.5		silt		
		29/7 - 12/8/86	152.5	1.57	sandy gravel	1.0	Minimum estimate Coarsening upwards Base D50 = 0.1 mm
		12/8 - 18/8/86	77.5	1.72	sandy gravel	0.9	No structure
		19/8 - 13/9/86	0.85		sandy silt	0.9	Some gravel
T1/T2/T3	Fence	12/6 - 15/7/87	0				No sediment
		15/7 - 17/7/87					All traps destroyed Some clasts trapped b axes = 40-220 mm
T4	Fence	12/6 - 22/6/87	28.5		sandy gravel	1.5	Minimum estimate
		22/6 - 27/7/87	0				No sediment

NOTE: 1. Observation dates only refer to periods when there was sediment movement.

Figure 5.10 Fence traps in the Bas Arolla proglacial zone tributaries. A) Effective fence trap - Tributary T4 June 22nd 1987. B) Failed fence trap - Tributary T1 July 17th 1987.

obtain an estimate for the (whole season would appear to be reasonable since 1986 rainstorm activity (Figure 5.3a) was approximately evenly distributed throughout the summer.

Over the winter tributary flow is negligible and yields are very low. Tributaries T1 and T2 yielded small amounts of sediment during the 8 months from September 1986 to June 1987. In 1987 T4 yielded the greatest amount of measured bedload which is surprising since in 1986 no sediment was contributed by this channel. Tributary T5, the largest contributor in 1986, could not be effectively monitored in 1987 due to theft and vandalism of traps. Because of these problems an overall estimate for 1987 tributary bedload yield is difficult.

The type of bedload moved in the tributaries depends on the flow conditions. Figure 5.11 demonstrates the wide range in grain-size distributions of trapped sediments. Much of the trapped material is of a size suitable for transport in suspension, which implies that a significant proportion of fine sediment is stored in tributaries during low flows. Individual tributaries show a wide range in transported grain-sizes (e.g. T5) as do comparisons between tributaries. Notably, the coarsest sediment transported from a tributary source area was in the slushflow event of June 8th 1987 (section 5.3.1), reflecting the glacial moraine source area from where it originated and the difference in the transport mechanisms of slush and water flows. Tributary trapped sediments have characteristic D_{50} grain-sizes which vary between 0.05 to 1.0 mm with the majority of sediment of a sandy gravel grade. This suggests a large store of relatively fine material in the tributaries, although some of this fine material may enter the trap by ingress through the gravel matrix. Very little of the trapped sediment showed any marked structure which supports the hypothesis that the

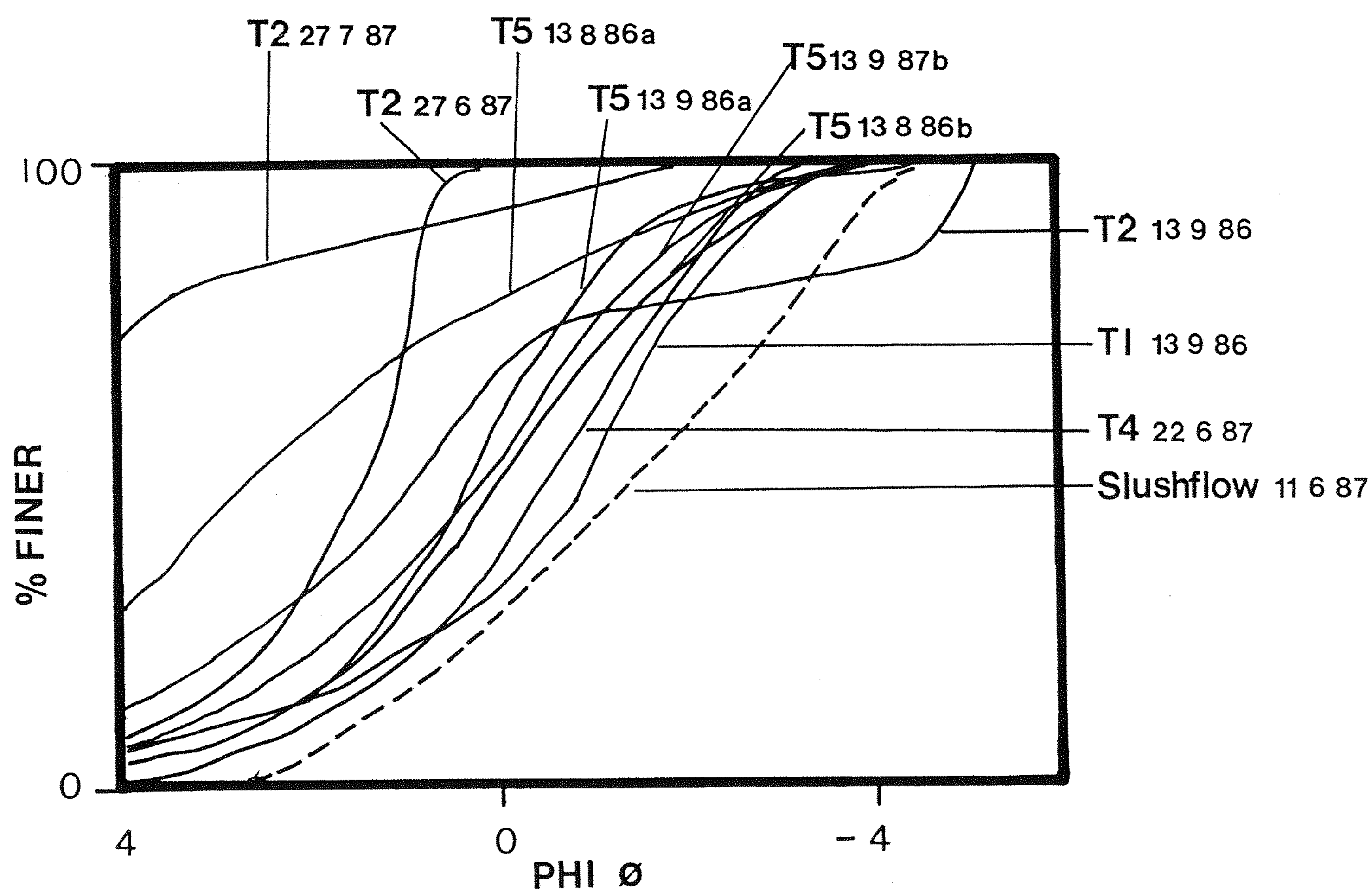


Figure 5.11 Range in grain size of trapped tributary sediments. Captions show location of sample and date of sampling. a = sample from the base of the trap, b= sample from centre of trap.

majority of tributary sediment transport occurs 'en-masse' during storm events. This is supported by survey evidence from tributary T5 which showed a 'hump' in the tributary long profile corresponding with a slug of sediment. Movements of slugs of sediment have been identified in a number of fluvial systems (e.g. Sawada et al., 1983).

Tracer studies in tributaries T1, T4 and T5 in 1987 attempted to demonstrate the link between tributary and mainstream coarse sediment systems. Tracers where introduced ('seeded') in a 1 m wide strip on the stream beds. Results in Table 5.2 demonstrate movement of tracers from the tributaries to the mainstream. Movements in tributaries up until the middle of July showed only minor downstream movement. It was only in the period 15th - 19th July that clasts really started to move. After 19th July tracers were found in the mainstream, and by 7th September one tracer (axis 50 mm) was found at least of 920 m from the introduction point (the exact distance was not known since the clast could have come from tributaries T1 , T4 or T5). A total of 35 tracers were found beyond the tributaries; 22 were in the main channel with a further 13 found below the water intake structure. The tracers showed crude downstream sorting with the majority of clasts deposited in pools. Most of the recovered clast had axes in the range 30 to 80 mm. All three tributaries contributed clasts to the main channel, a result which is somewhat surprising since sediment trap data suggest that most of the bed sediments are sandy gravels with relatively fine D_{50}. However, because of the restricted grain-size range of the tracer particles (coarser than 15mm b-axis), they do not mimic the grains of the bed. The combination of relatively low sediment yields and active transport of coarse clasts into the main channel may at first seem contradictory. For example in tributary T1 a large clast (a axis 210 mm) was transcported 31 m in a single event (15th July

Table 5.2 Bas Arolla Tributary tracer experiments

Tributary	Dates	Number of tracers	Colour	Percent recovered	Stationary	Lost	Moved	Maximum distance moved m	Size of clast moved farthest a axis mm
T1	2/7 - 17/7/87	66	green	100	94	0	6	31	230
T4	2/7 - 17/7/87	138	green	100	138	0	0	0	0
	17/7 - 19/7/87	138	green	9	0	125	13	24.8	55
T5	6/9 - 13/9/86	184	peach	100	184			0	
	13/9 - 2/7/87	184	peach					6	52
	2/7 - 19/7/87	184	peach	11	0	163	21	119.8	35
	19/7 - 6/9/87	184	peach					204	35
	2/7 - 17/7/87	104	green	100	82	0	22	7.3	55
	17/7 - 19/7/87	104	green	31	0	72	22	57	30
T1/T4/T5	2/7 - 17/7/87		green					>150 *	85
T1/T4/T5	17/7 - 7/9/87		green					920 *	50

Notes: * = Refers to observation of a single tracer, found in the main channel, which could have originated from either of tributaries T1, T4 or T5

1987) and deposited 10m short of the main channel where it stayed immobile for the rest of the study period. This example demonstrates the high magnitude, low frequency mobilisation of bedload in the tributaries (i.e. active sediment transport occurs but it is shortlived and sediment yields as a consequence are generally low).

The results of tributary bedload sampling studies suggest: preliminary yields of less than 0.5 tonnes; sporadic 'event-based' movements of bedload; and a connection between the coarse sediment transport systems of the tributaries and the main channel. These conclusions may be further tested by examining changes in the morphology of tributary cross sections and long profiles (i.e. channel scour and fill).

5.2.4 Channel scour and fill

Table 5.3 represents changes in cross-sections in three tributaries: T1, T2 and T3. 3 cross-sections on each tributary were resurveyed on 4 dates between June 23rd and July 27th 1987. Survey of tributaries T4 and T5 was also initiated but, could not be completed due to loss (theft) of survey pegs. Results presented without brackets are estimates of erosion or sediment loss (m^2). Results presented in brackets represent deposition or storage (m^2). Cross-section changes were generally small, the greatest change occurring between July 17th and 27th. Tributaries T1 and T2 showed the greatest change, T2 has the greatest net erosion, T1 shows large variations in erosion rates and T3 shows only minor change with the upper section showing net deposition. These data are useful since they give an impression of sediment flux within the tributaries but, because the cross-sections are closely spaced and all are located on the lower section of the tributary stream, they may give the wrong impression of sediment flux over the whole profile. In order to get a

Table 5.3 Cross-profiles of tributary streams, Bas Arolla proglacial zone 1987

	PROFILE			
	Lower	Middle	Upper	Mean
Tributary 1.				
23/6 - 13/7	0.01 (0.00)	0.04 (0.02)	0.02 (0.00)	0.02 (0.01)
13/7 - 17/7	0.30 (0.05)	0.10 (0.01)	0.17 (0.03)	0.19 (0.03)
17/7 - 27/7	0.08 (0.26)	0.14 (0.22)	0.10 (0.06)	0.11 (0.18)
All	0.39 (0.31)	0.28 (0.25)	0.29 (0.09)	
Balance	0.08	0.20	0.03	0.10
Tributary 2.				
23/6 - 13/7	0.10 (0.00)	0.00 (0.03)	0.02 (0.04)	0.04 (0.02)
13/7 - 17/7	0.06 (0.03)	0.03 (0.03)	0.15 (0.00)	0.08 (0.02)
17/7 - 27/7	0.05 (0.05)	0.26 (0.07)	0.15 (0.14)	0.15 (0.09)
All	0.21 (0.08)	0.29 (0.13)	0.32 (0.18)	
Balance	0.13	0.14	0.16	0.14
Tributary 3.				
23/6 - 13/7	0.00 (0.02)	0.02 (0.00)	0.05 (0.00)	0.02 (0.01)
13/7 - 17/7	0.02 (0.02)	0.00 (0.03)	0.00 (0.08)	0.01 (0.04)
17/7 - 27/7	0.08 (0.00)	0.03 (0.12)	0.04 (0.00)	0.05 (0.04)
All	0.10 (0.04)	0.05 (0.15)	0.09 (0.08)	
Balance	0.06	0.01	(0.10)	(0.03)

NOTES: Results without brackets represent scour
Results with brackets represent fill

better representation of sediment flux, the lower sections of the tributary long profiles were surveyed on two dates (July 13th and 17th, Table 5.4). Study reaches varied from between 6.2 to 18.2% of the total tributary lengths. Comparison of the two profiles allowed determination of the proportion of the channel which was either eroding, depositing or stable. The proportionate lengths of channel eroding, depositing or stable were approximately equal in the three tributaries (Table 5.4). In tributary T1, 55% of the channel is erosional (net erosion 0.027 m^2 m^{-1}). In tributary T2 the length undergoing deposition is greater than the length undergoing erosion but due to deep scouring in certain sections of the profile result in a net balance which is erosional (0.045 m^2 m^{-1}). T3 has a net depositional balance with 54% of the channel undergoing deposition and a large proportion of the reach (15%) which is stable. This is largely in agreement with the general patterns observed in the cross-sections.

Estimation of preliminary loads can be calculated by two methods by multiplying either mean sediment loss (Table 5.3) or net erosion (Table 5.4) per unit length of channel, by the cumulative length of all tributary channels. Yields estimated for the cross-section method equal 50.4 tonnes with a yield of 17.1 tonnes determined from the long-profile measurements. The later estimate is only based on sediment loss between July 13th to 17th and does not apply for the whole study period. However this appeared to be the period when the majority of sediment was transported (a period of rainstorm-generated high flows) and there appeared to be little change in profiles before and after these dates. In addition, since both types of calculation relate to total scour and fill of the stream bed a proportion of the sediment will have been lost in suspension rather than as bedload. The proportion of bed material lost in suspension could be as high as 50 % by weight if taking 1mm is taken as a suitable suspended/bedload boundary (Figure

Table 5.4 Channel change in tributary streams T1, T2 and T3 (13th to 17th July)

Tributary	Length of reach m	Length of reach as a % of total channel length	Average width m	Percent channel length Erosion	Deposition	Stable	Volume eroded m3	Volume deposited m3	Net change m3 m-1
T1	20.5	9.3	1.90	55	45	5	3.02	2.46	-0.027
T2	25.5	18.2	1.75	40	48	12	4.47	3.31	-0.045
T3	11.5	6.2	1.80	33	54	13	1.61	2.42	0.070

5.11). Re-adjusting the yields to take account of this results in new values of 25.2 tonnes for the cross-section method and 8.55 tonnes for the long profile technique. Furthermore, because survey of long profiles and cross sections were confined to the lower reaches of the tributary, multiplication by overall stream length is misleading, since it may be only the lower part of the tributary channel which is contributing sediment. In addition a large proportion of the flow in tributaries T1 and T2 was not generated by rainfall-runoff but was supplied direct from glacial sources. Glacial water was released from a portal in the ice, high on the western margin of the glacier. Flow drained down the western slope of the proglacial zone along streambeds T1 and T2 (T3, being lower down the valley was unaffected). Because this major part of flow entered the tributary channels T1 and T2 only in their lower reaches, it was only these areas that were significantly involved in contributing sediment. Flow of this kind occurred between July 15th to 17th and, under these conditions, the active channel lengths for T1 and T2 were 50 and 65 m respectively. Re-calculation of bedload yields on this basis and allowing for suspended sediment loss, results in estimates of 9.87 tonnes (cross-section method) and 3.32 tonnes (long profile technique) for bedload yield from the tributaries. The cross-profile method is thought to be the least reliable of the two since 3 cross-sections cannot adequately characterise down channel variations in scour and fill in these steep tributary channels. In this kind of step-pool channel, the long profile better accounts for storage/erosion, sites since local channel characteristics, such as step-pool morphology (Whittaker, 1987) and local gradient, influence sediment transport (Laronne and Carson, 1976). If the estimate of sediment yield from tributaries is calculated by the long profile method to be of the order of 3.32 to 8.55 tonnes (depending on the 'active' stream length) this is an order of magnitude greater than the estimate from sediment traps. A proportion of this discrepancy will be explained by differences in the measurement techniques and measurement error. In addition, errors also arise due to: 1) Sediment loss overbank and as storage in levees. 2) Exceedence of sediment trap capacity. 3) Sediment loss from traps because the trap is sited in an erosional reach. 4) Not al the sediment transported by the stream is of tributary origin (e.g. flows in T1 an T2 between July 15-17th carried glacial sediment). 5) Not all sediment stores may have been recognised.

This last point may well account for a major part of the discrepancy since sediment may be stored in the intervening channel between survey reaches and trap locations (e.g. Table 5.4 tributary T3). However because of incomplete observations in all 5 tributaries, these yields are to be regarded as minimum estimates especially since bedload transport in T5, the largest tributary, was not fully characterised. A very rough estimate of bed material yield from the tributaries can be obtained by taking: the maximum estimated bed sediment loss, based on long profile survey, from tributaries T1, T2 and T3 for 1987 (8.55 tonnes); the fence trap yield from tributary T4 (Table 5.1, = 0.029 tonnes); and an estimate of the yield in Tributary T5 from the 1986 data (Table 5.1, 153 kg in 56 days) = 1.17 tonnes. This gives a total yield of bed sediment from all 5 tributaries of 8.749 tonnes.

The implications of these findings are that sediment traps may underestimate yields if they are badly designed and sited. It appears that in tributaries T1 and T2 sediment was actively transported. In order to evaluate this last assumption a crude residence time can be calculated by dividing the mass of sediment in storage by the flux (Dietrich et al., 1982).

$$= S\ min\ \ or\ S\ max\ /\ S\ flux$$

where S min and S max = minimum and maximum estimates of stored sediment in tonnes and S flux = the flux of sediment in tonnes yr. The mass of sediment in storage is calculated by multiplying the length of the tributary where deposition is occurring by the mean depth of the sediment stores (determined from excavations and probing of storage sites). This is multiplied by an appropriate packing density (Table 5.1) to obtain an estimate of the stored mass. Residence times are calculated for tributaries T1 and T2:

T1	*S min*	*16.46 / 6.3*	*= 2.6*
	S max	*65.86 / 6.3*	*= 10.5*
T2	*S min*	*26.33 / 5.94*	*= 4.4*
	S max	*105.3 / 5.94*	*= 17.7*

These results suggest very active reworking of the sediment stored in tributaries. Between 6 and 38% of the tributary sediment store is reworked annually. To some extent this is supported by tracer evidence which suggests active mobilisation of tributary bed materials and supply to the main channel over short timescales. These estimates are still subject to the same limitations as those imposed on estimating bedload yields, since there is still the discrepancy between trapped and surveyed losses. Calculating residence times using trap yields (0.5 tonnes) suggest values in the order of 30 - 200 years. Comparison of these estimates with those of Stott (1987) for similar sized channels in Balquidder, Scotland show these values to be much lower (i.e. Balquidder residence times were estimated to be 1000 years). The main difference between the two sites is the presence of organic debris dams (Heede, 1972) at Balquidder, which suggests that they are very important for the storage of tributary sediment. However as this study has shown, there is no consistent pattern to the erosion of tributaries except that fine material tends to be removed first (Figure 5.11) leaving an armoured bed (Dietrich and Dunne, 1978; Lehre, 1983).

5.3. Supply from hillslopes

Sediment supply from hillslopes can be directed into tributary channels and transported by fluvial action or may be deposited directly on the floor of the valley train. Inputs include rockfalls, landslides and other rapid mass movements all of which contribute to the available sediment in the proglacial zone. The importance of these inputs will be controlled by several factors including: valley morphometry (which will determine run-out distances and the nature of the deposit); the frequency of these events; the seasonality of these events (i.e. over-snow events in comparison with events at times without snow). Direct supply to the valley train by-passes storage in the lower slope area and involves rapid mass movements (slope failure, avalanche, debris flow or rockfall) or transport of debris over a seasonal snowcover. The influence of snow on slope processes and the transport of debris is considered in Section 5.3.1. Rockfall, involving accumulation of moraine at the glacier snout is considered in Section 5.3.2 Rapid mass movements are discussed in Section 5.3.3. Estimates of direct sediment supply to tributaries by soil loss from inter-tributary areas are than considered (Section 5.3.4). This combination of slope and soil movements defines a set of geomorphological processes which are characteristic of zero-order basin sediment transfers (Dietrich et al., 1987).

5.3.1 The influence of snow

The presence of snow in the proglacial zone influences sediment transport in three principal ways:

1) Ablating snowpacks provide a water source for fluvial erosion and transport (Caine, 1980; Clark, 1988).

2) Snow in the form of avalanches and slushflows is

important in the erosion and transportation of debris (Fenn and Gurnell, 1987; Dowdeswell, 1982).

3) Snow acts as a surface over which sediment can be transported, temporarily stored and then redeposited following melt.

The role of snowpacks in accelerating erosion has been discussed at length by Thorn (1987). The conclusion from measurements of tributary streams draining snowpacks in the Bas Arolla proglacial zone is that they transport relatively low concentrations of suspended sediment with concentrations rarely exceeding 500-700 mg l^{-1}. In the early season concentrations from snowpack tributaries may be the highest in spatial surveys (e.g. Figure 5.6, 14th June 1987) but the low discharges mean that sediment yields are very low. In the Canadian Arctic, rills draining snowpacks are important in transporting fine sediment (Wilkinson and Bunting, 1975) with peak rill concentrations in this environment exceeding 120 g d^{-1}.

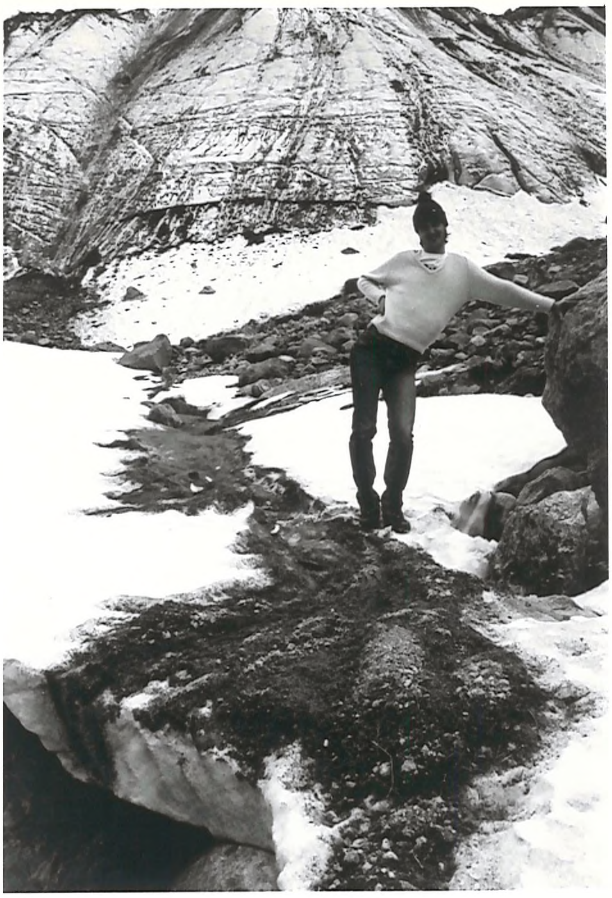
Snow movements in the form of avalanches, constitute sporadic events but they can yield large amounts of debris directly to the proglacial valley train. This debris not only serves as a source of sediment but can also cause modification of the channel structure. Fenn and Gurnell (1987) describe such an event on the valley train of the Bas Arolla proglacial zone which caused channel diversion and readjustment in the channel profile prior to and following incision of the valley train blockage. Since no events of this kind were observed in the study period their role cannot be assessed.

Snow also has an indirect effect on sediment transport indirectly since it temporarily stores sediment on its surface and, when it ablates, the debris is deposited in areas where the sediment would not normally be released (had the snow not been present). Over-snow or supra-nival sediment transport is important in delivering sediment directly to the stream channel or the immediate channel margins. Three types of supra-nival process were observed: a) Transport of material from bluff and bank faces (Koutaniemi, 1984). b) Transport of sediment in streamflow over snow (Woo, 1979). c) Transport of sediment in a debris/ice mix and deposition as an ice/sediment delta (similar to Dowdeswell, 1982) and in slushflows similar to those described by Rapp (1960), Caine (1969) and Washburn (1979).

Transport of bank and bluff material over a basal snowpack is shown in Figure 5.12b. This process will be considered in more detail in Chapter 6 in the discussion of bluff and bank erosion. It is mentioned here because similar processes, although on a much smaller scale, occur in tributary areas and usually involve the transport of small boulders.

Transport of sediment in streamflow over the snow surface only occurs when the transition from channelised flow over the ground to over-snow flow can be accomplished without significant loss of discharge. The example shown in Figure 5.12a demonstrates this process. In this case one of the west slope tributaries, following rainfall on the 9th June 1987, 'broke-out' of it s channel and flowed down to the margin of the valley train which was infilled with snow. Flow continued over the snow for a further 7-9 metres before being halted by discharge losses due to percolation into the snow; leaving a debris train on the snow surface. Once the snow melted this debris was supplied directly to the channel. This was the only flow of this type to reach the valley train during the 1987 field season. Several similar flows, were observed on the adjacent slopes but none contributed sediment directly to the valley train. Most flows terminated in the zone adjacent to the valley train margin. The

Figure 5.12 Transport of sediment over snow. A) Upper photograph - sediment transported in supra-nival streamflow. B) Lower photograph - transport of bluff and bank material over a basal snowpack.

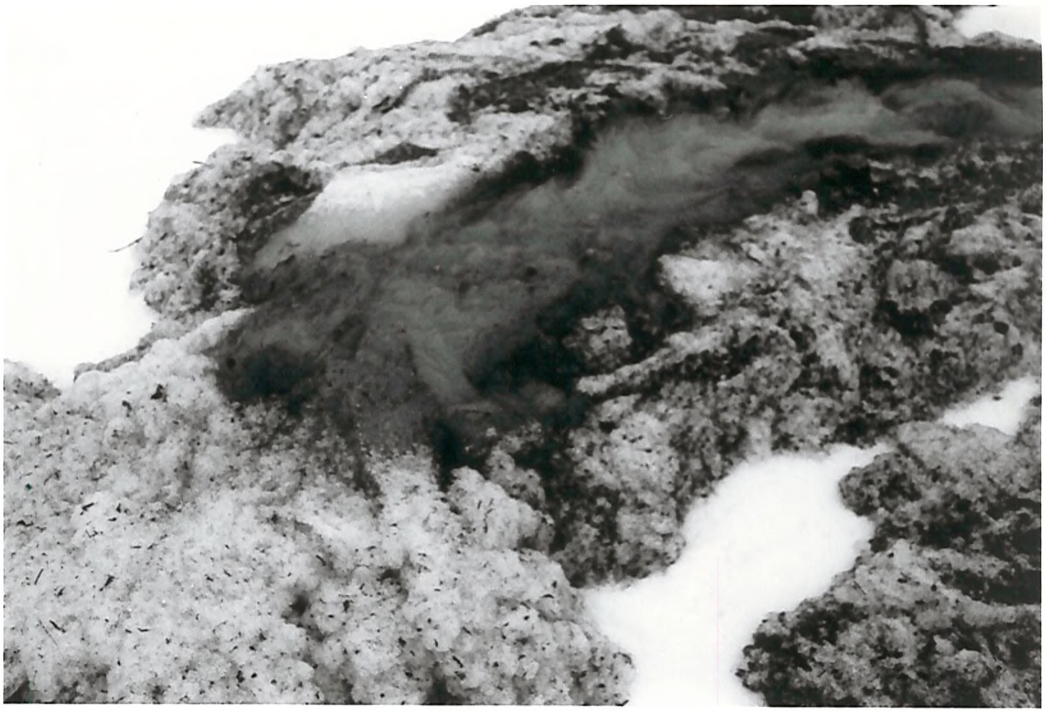


amount of sediment delivered directly to the valley train by the flow described above was 0.39 tonnes (calculated from survey of the dimensions of the deposit and estimates of the packing density of the material).

A much larger input of sediment was generated by a slushflow (Washburn, 1979) on the evening of the 8th of June (Figure 5.13). Streamflow down the eastern margin of the glacier entrained a mixture of surficial debris, fresh and old snow and a significant proportion of coarse organic debris. The deposited sediment was coarse (D_{50}, 2.88 mm) which probably reflects the fact that most of the sediment was derived from the immediate glacier snout zone and was, therefore, most likely to be of a coarse angular nature. Sediment was supplied to the valley train along two pathways. Initially the flow was supra-nival and a large (15m x 7m) fan developed (Figure 5.13a). This was subsequently head-cut, as the flow turned from slush to water, and flow was diverted under the snow at the margin of the valley train, along the line of the bluffs. Deposition of debris in the fan was in the form of small debris trains and miniature fan deltas (Figure 5.13b). The main thread of flow was continually separating, splitting and changing direction as small debris-ice dams were formed and then by-passed. Out-building of the fan occurred in a series of small deltas (Figure 5.13b). Flow over the snow continued until the water percolated into the snow, causing deposition of the debris. Refreezing of ice crystals after percolation, in turn created a base of reduced permeability on the snowpack over which subsequent flow could be maintained. New flows passed over this basement until they reached the fan front, where they would flow out over new snow until the flow was lost to percolation. Out-building of the fan was accomplished by repetition of these processes.

These processes are essentially the same as those described by Dowdeswell (1982) in his description of

Figure 5.13 Slushflow of June 8th 1987. A) Slushflow 'delta'. B) Detail of advancing slushflow front.



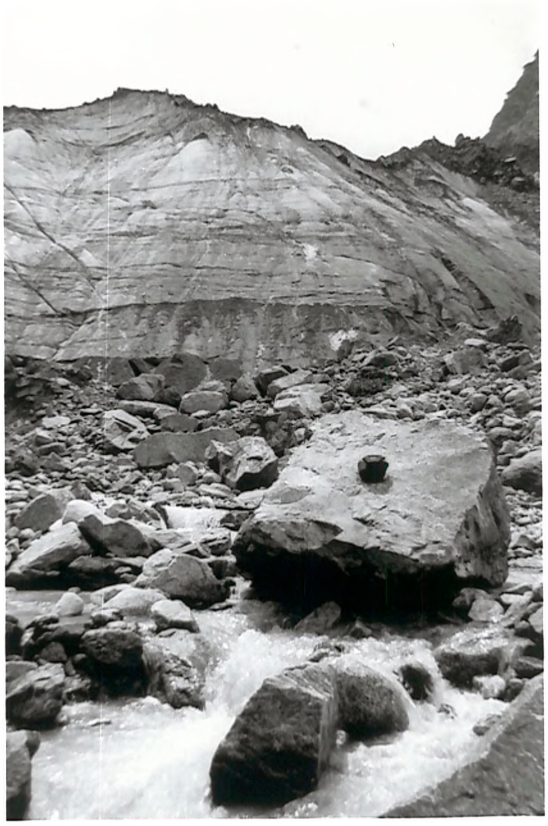
re-sedimented debris trains. Dowdeswell (1982) concluded that these flows were of little significance in terms of proglacial sedimentation. In the present example, which was the only event of its type observed in 1987, the amount of sediment contributed supra-nivally to the valley train was 4.10 tonnes (calculated from survey of the dimensions of the deposit and estimates of the packing density of the material). Undoubtedly more debris was transported by the event but a proportion of the load was routed under the snow directly to the main channel. At this later stage, flow was no longer of a slushflow type and only continued for approximately two hours after the bulk of the fan delta had been deposited.

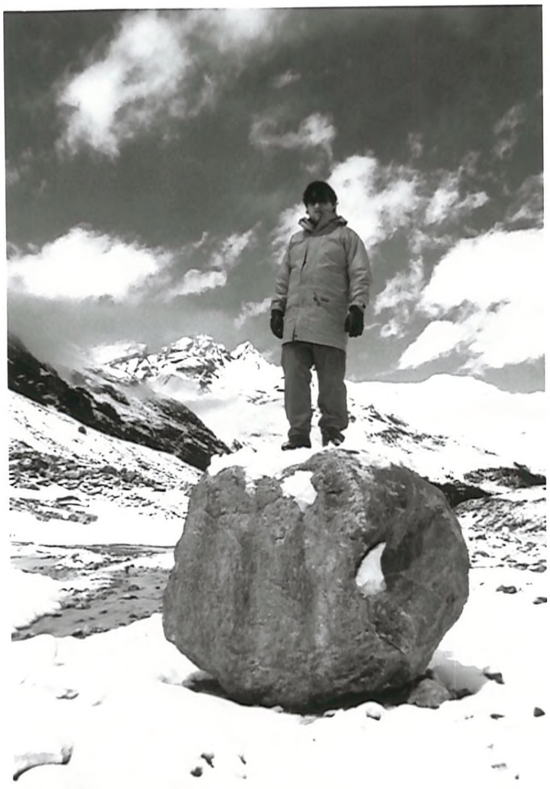
The importance of supra-nival transport processes is in 'by-passing' intervening sediment stores. Material is temporarily deposited on the snow surface, which then ablates and releases sediment to the valley train/proglacial stream. Sediment is released in areas which may not normally receive this type of sediment input. The events observed in 1987 where small in scale and no avalanche activity was monitored. However, if large quantities of debris were delivered via supra-nival pathways to the valley train, the effect on the sediment system would probably be considerable.

5.3.2 Rockfall

Although no mass movements supplied slope debris directly to the valley train, large blocks and boulders, as a result of rockfall, was deposited in the channel zone. Rockfall (the free, bounding and rolling of rock debris downslope under the influence of gravity (Luckman, 1976)) only supplied two large blocks to the valley train in 1987 (Figure 5.14). It was observed that rockfall was most prevalent in late May and early June with the diurnal rhythm of activity focussed around solar noon (Lewkowicz, 1987). Most rockfall was

Figure 5.14 Large blocks supplied to the Bas Arolla valley train in 1987. B) Block fell from west slope lateral moraine via tributary zone to the valley floor June 8th 1987. A) This boulder fell directly from the snout of the glacier into the valley train July 26th 1987.





released from the high lateral moraines on the western side of the Bas Arolla valley. Minor falls released from the snout of the glacier, most of which accumulated as a fringing moraine. The moraine was deposited immediately in front of the glacier. Detailed survey of moraine accumulation was difficult due to the danger from further rockfall. The amount of debris was estimated from the aerial extent of the debris apron deposited on snow (determined from photographs) multiplied by average thickness and an estimated packing conversion factor. The average thickness of the deposit was determined by probing the depth of the accumulated debris.

$$\begin{array}{lcrl} \textit{Area of moraine} & = & 189.0 & m^{-2} \\ \textit{Average depth} & = & 0.4 & m \\ \textit{Packing factor} & = & 1.6 & \textit{tonnes } m^{-3} \end{array}$$

$$\textit{Debris} = 120.9 \quad \textit{tonnes}$$

This estimated weight of debris related to the period 25th May to 24th June 1987. A rough estimate of the total accumulation until the 30th July can be obtained by multiplying the daily accumulation rate (determined from the above data) by the number of days where accumulation was not measured. When added to the above total this gives an estimate of 265.9 tonnes of deposition. These estimated rates of moraine accumulation compare reasonably well with estimates of annual marginal moraine deposition at the Glacier de Tsidjiore Nouve (Small, 1987).

Debris from lateral moraines was deposited at the slope foot outside the margins of the valley train (Figure 5.1) and so this is not considered in the present valley train study. Only two very large blocks reached the valley train (Figure 5.14). One block (Figure 5.14a) originated from the west lateral moraine and fell into the valley train on the 8th June. This block had dimensions of 1.6 x 1.4x 1.2 m. Calculating the volume of the block using these dimensions and multiplying by 2.57 (the specific gravity of the rock type) yields a weight of 6.91 tonnes. The second block (Figure 5.14b) which fell directly from the snout of the glacier, had dimensions of 1.5 x 1.4 x 1.2 m and weighed 6.48 tonnes (calculated by the same means as above). The lack of smaller material in the valley train is thought to be a function of lack of momentum of smaller blocks when falling, such that they come to rest on the intervening slopes before reaching the valley train.

It is suggested that the importance of these blocks in the valley train is not in their addition of significant amounts of debris (although in solid rock terms it is an order of magnitude greater than supra-nival processes) but in the effect they have on fluvial processes within the valley train (Chapter 7).

5.3.3 Rapid mass movements

Most hillslopes are erosional in nature and are therefore an important source of sediment. In alpine areas, high-levels of geomorphic activity should be expected due to the high gravitational stress on slopes and morphoclimatic influences (Caine, 1974). This is especially true of the Bas Arolla proglacial zone where steep (70^{o}) (Whalley, 1976) corrugated moraines tower above the lower proglacial valley slopes (Figure 5.1). However, slopes have long geomorphic response times and so it is very difficult to study slope processes. Supply of sediment from the hillslope to the proglacial fluvial system was therefore not quantified during the study period.

Rapid mass movement occurs when slopes fail, which is usually in response to a finite yield strength. The potential for failure can be evaluated by stability analysis but this requires detailed knowledge of soil strength parameters and slope geometry prior to the slide. Although this detail is not available, empirical studies based on relations between rainfall characteristics and observed failures provide limiting thresholds for this type of slope instability. It is presumed that rainfall has an indirect effect on slope stability through its influence on pore water conditions in the slope mantle.

Rainfall data were obtained from 24h rain catch totals measured in the immediate proglacial zone (Figure 5.2). Using the limiting thresholds proposed by Caine (1980) for shallow slides and debris flows and Innes (1983) for just debris flows, it is possible to calculate the minimum rainfall input in a 24 hour period for which slope failure is predicted to occur. Figure 5.15 plots these thresholds against the 5 largest rainfall totals observed at the Bas Arolla site for 1986 and 1987. None of the observations exceed the 24 hour threshold of Caine (1980) and only 3 exceed the lower limiting 24 hour threshold of Innes (1983). This may be misleading since the rainfall recorded during the 24 hour sampling period may all have fallen within the first hour. Alpine storms tend to be relatively short-lived (see data in Caine, 1976; Barry, 1981), therefore the 6 hour thresholds are also calculated for comparison (Figure 5.15). In this case observations are much closer to the *Caine threshold* but none exceed it, yet all storms exceed the *Innes threshold*, implying that debris flow and shallow slip activity is likely during periods of heavy rainfall. This kind of activity was observed in tributary heads during three storms (July 7th 1986, June 8th 1987 and June 15th 1987). On all three occasions flows and slides were small. More precise definition of the threshold exceedence events requires the use of a recording rain gauge.

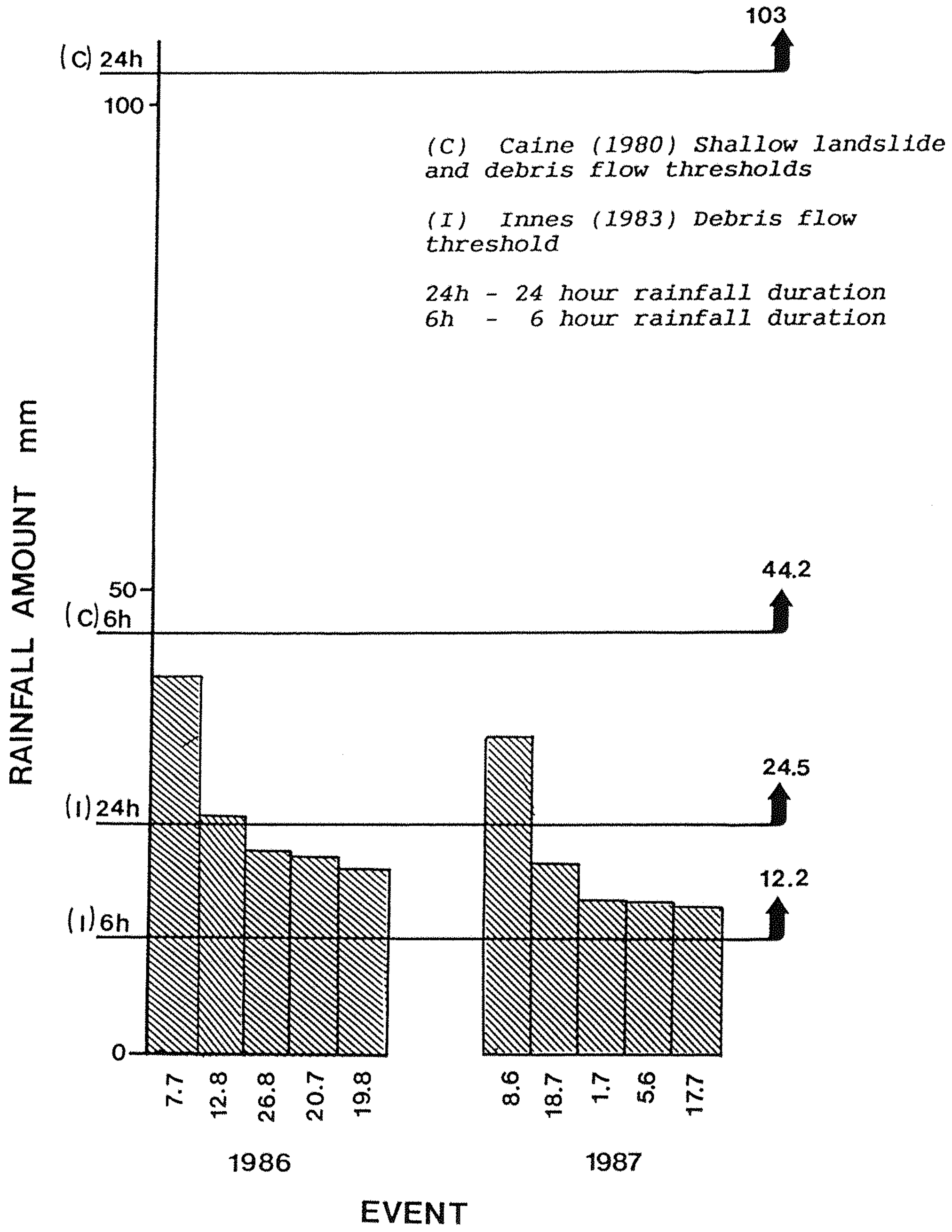


Figure 5.15 'Largest 5' 24 hour rainfall events in 1986 and 1987 plotted in relation to shallow landslide/debris flow thresholds.

5.3.4 Supply of sediment from inter-stream areas

Sediment can be mobilised by running water on slopes where there is no apparent channelised flow. This can be an important source of sediment in the immediate vicinity of channels and was therefore measured using sediment traps of the 'Gerlach type' (Morgan, 1979). Because the traps were not fitted with lids, they not only sample sediment loss from runoff but also from rain-splash. Therefore, a proportion of the trapped sediment may have been derived from below the trap (due to retrograde movement) giving a slight over-estimate of the downslope sediment flux, although this was probably partly offset by sediment loss from the trap caused by direct raindrop impact.

Table 5.5 shows the results of the sediment trap studies. Referring to the 1986 data, specific yields varied between 0.08 and 2.37 g m^{-1} d^{-1}. The sediment yield seems to reflect a combination of slope angle and vegetation cover (low slope angles and high vegetation covers produce low yields). The dominant grain-size intercepted in the traps was sand with smaller amounts of gravel and silt/clay. Organic contents were low, generally less than 5%. The 1987 measurements showed similar specific yields, but more erratic variations in grain-size and generally lower organic contents generally (less than 2.5%).

Contrasts also existed at the same trap for different periods (Table 5.5). Trap 2 in 1986 had a specific yield of 0.63 g m^{-1} d^{-1} for mid July to mid September. Over the winter yields were low (0.12 g m^{-1} d^{-1} (mid September to mid June 1987) From mid June to the end of July 1987 yields increased again to 0.48 g m^{-1} d^{-1}. Trap 5 showed a similar pattern of reduced yields over the winter period. Grain size also varied markedly during this period (e.g. in Trap 2, gravel increased from 15.3% to 26.4% to 62.9% for the

Table 5.5 Sediment yields estimated using trough traps, Bas Arolla proglacial zone

TRAP	SAMPLING PERIOD	SLOPE ANGLE o	VEG. COVER %	GRAIN SIZE % GRAVEL	SAND	SILT/CLAY	ORGANIC CONTENT %	TOTAL YIELD g	SPECIFIC YIELD g m-1 day-1
1986									
1	19/7-13/9	5	0	7	82	11	4.6	30.84	1.10
2	19/7-13/9	9	40	15.3	75.2	9.5	1.4	17.62	0.63
3	19/7-13/9	17	15	14.5	77.3	8.2	0.9	66.36	2.37
4	19/7-13/9	20	5	11.7	70.7	17.6	0.8	18.06	0.65
5	19/7-13/9	11	90	18.7	56	25.3	5.1	5.42	0.19
6	19/7-13/9	13	80	9.6	74.7	15.7	3.1	2.36	0.08
1987									
2	13/9-22/6	9	40	26.4	51.6	22	1.4	16.85	0.12
2	22/6-29/7	9	40	62.9	34	3.1	0.9	8.91	0.48
4	5/7-29/7	20	5	9.8	74.1	16.1	0.5	3.88	0.32
5	13/9- 5/7	11	90	2.4	44.4	53.2	2.2	14.12	0.10
5	5/7-29/7	11	90	25.4	54	20.6	2.4	3.35	0.28
6	5/7-13/9	13	80	13.3	28.6	58.1	2.5	2.06	0.17

three periods mentioned above).

Pooling of the data presented in Table 5.5 gives a mean specific yield of 0.54 g m^{-1} d^{-1} (standard error = 0.65). Assuming that this figure is a reasonable estimate of sediment delivery adjacent to the tributary stream courses where the traps were located, and accepting the inherent variations in vegetation cover, slope angle and sediment availability (Owens, 1969), the total amount of sediment supplied to the channel can be estimated. For the period 25th May to 30th July 1987 (67 days) the sediment supply from tributary margins is given by:

$$= \overline{SSY} \ x \ n \ x \ (TL \ x \ 2)$$

where SSY = mean specific sediment yield, n = number of days and TL = the summation of all tributary stream lengths. TL is multiplied by 2 to account for the contribution of sediment from both channel sides. Total tributary length is difficult to define given the ephemeral nature of the stream network. The total sediment yield calculated by this method is 0.076 tonnes.

5.5 Summary

The lower side slopes of the Bas Arolla proglacial zone are drained by ephemeral tributaries of the main proglacial stream and have characteristics of both zero-order and first-order basins. Runoff is contributed from snowmelt and rainfall, with rainfall-runoff events being responsible for the bulk of the sediment transport.

Sediment contributions from tributary sources are highly variable and are generally most important early in the ablation season. Suspended sediment yield between May 25th and July 30th 1987 was between 103.6 and 319.2 tonnes with approximately 99% contributed from one tributary (T5). Of this load, 52% was transported in one day on June 15th.

Bedload yield was difficult to estimate but (based on tributary surveys and trapping of transported bed material), varied between 3.32 to 9.87 tonnes or 17.1 to 50.4 tonnes depending upon the estimate taken. The two orders of magnitude difference between suspended sediment and bedload yield estimates was surprising. However, sediment transport in tributaries cannot be easily partitioned into load components since at high flows small pebbles will be transported temporarily in suspension (the largest particle sampled by hand in suspension was 6mm). A mean estimate of sediment output by tributary stream flow is 220.15 tonnes for the period 25th May to 30th July 1987. The error of this estimate is +/- 71%, but may, be considerably greater considering the assumptions and uncertainties surrounding the measurements.

Direct inputs into the valley train in between 25th May and 30th July 1987 came from: flow over snow (0.39 tonnes); slushflow (4.10 tonnes); rockfall (13.39

tonnes); and moraine deposition at the glacier snout (265.9 tonnes). Inter-rill areas supplied a small amount of debris (0.076 tonnes) directly to the channel.

Therefore within tributary channels contributed between 106.92 to 329.07 (120.7 to 369.6) tonnes of sediment to the valley train with another 283.78 tonnes contributed directly by slope processes.

Given the measurement and estimation techniques used in calculating these sediment contributions, to the valley train they must be treated as very preliminary estimates. Furthermore estimates only apply for the period of study in 1987 (25th May to 30th July) when no large scale slope failures were monitored. An appropriate conclusion to this chapter is that '....diversity in time and space is perhaps the single most significant characteristic of the alpine zone'(Caine, 1974).

Chapter 6.

PROGLACIAL CHANNEL CHANGE

6.1 Introduction

Mobile alluvial channels have been the focus of a vast geomorphological literature over the past two decades (e.g. since Leopold, Wolman and Miller, 1964). The recognition of various mobile channel types and the documentation of their rates of movement, and channel patterns, has considerably enhanced understanding of alluvial channel behaviour (Gregory, 1977). Understanding of alluvial channel behaviour is based on the premise that alluvial river channels are 'self-formed' (Richards, 1982). Alpine proglacial stream channels are not self-formed in the strict sense defined by Richards (1982) and cannot be considered in the same fashion as these types of channel. Rather, the morphology of these streams results from the interaction of entrainment, transport and deposition processes with mobile and stable bed elements. Spatial variations in stability and instability of the bed coupled with variations in sediment supply largely explain the episodic nature of sediment transport and channel adjustment in mountain streams, of which proglacial mountain streams are a good example.

In mountain streams, except during major flood flows, a proportion of the stream bed will be mobile and a proportion stable (Section 4.4). The juxta-position of stable and unstable bed elements produces discontinuity in sediment transport and large spatial variations in form. This notion is at odds with equilibrium and hydraulic geometry concepts, since the origin of channel form is only partly produced by flow conditions. This is not a new perspective. Furbish (1985) discusses this in relation to a mountain stream in the Colorado Front Range, USA, where the stream was essentially stable over the entire study reach with flows incompetent to mobilise bed material and as a consequence virtually zero bed material transport. For the Bas Arolla proglacial stream, studied here, the bed topography is similar to the Furbish example but, because of a much greater range in flows, greater sediment availability and the absence of vegetation, a large proportion of the bed is mobilized and, as a result, channel form begins to be influenced by the flows. However, mountain streams cannot be viewed as a simple continuum of gravel bed rivers since there is a marked imbalance between sediment characteristics and flow competence. For example, the transport of large boulders during flood has been thought to be indicative of the maximum power of the flood (Costa, 1983). However, given the large heterogeneity in grain-size in mountain streams large boulders can be transported at flows less than those that might be expected or predicted. This can be achieved by removal of surrounding sediment by undermining or scour on the lee side of boulders producing a 'hole-in-the-bed' into which the boulder topples. Alternatively, boulders may remain 'fixed' during the full range of flows, so having a profound effect on flow and sedimentation. Failure to recognise differences in bed material transport relationships will lead to misunderstanding of channel formation. The degree of 'self-formation' in mountain streams increases with flow but there is still a strong degree of erraticism in channel behaviour. Generalisations about one particular channel type are fraught with difficulty and cannot be easily extrapolated (Schumm, 1984).

Many authors have stressed the highly active nature of proglacial streams and channel forms which are continually adjusting (Krigström, 1962; Fahnestock, 1963; Maizels, 1979; and Fenn and Gurnell, 1987), but the nature of these adjustments is not fully discussed.

Models of sediment storage in mountain streams are rare. Trimble (1988) suggests that sediment storage in mountain streams is characterised by highly variable rates of sediment gain and loss. These changes are episodic and the sediment budget for such streams will largely reflect the influence of the last big flood event. Flood events are often associated with the crossing of an extrinsic geomorphic threshold (Schumm, 1977). Under these conditions channel morphology is dependent on flow and large scale channel adjustments can occur, but even so channel change is still influenced by the spatial distribution of sites which are amenable to readjustment (Anderson and Calver, 1980). The imprint of floods on channel form and subsequent stability will depend on several critical conditions: supply of coarse detritus, the recurrence interval of the large floods, the ratio of flood discharge to bankfull discharge and the inability of subsequent flows to modify flood forms (Gupta, 1983).

This chapter discusses proglacial channel change and its influence on sediment supply and storage in the valley train. Ultimately channel processes should be studied in three dimensions but accurate investigations of this type are very difficult and so two dimensional substitutes are often used (Thorn, 1988b).Emphasis is placed on changes in channel planform, profile and cross section and their relationship to erosion and deposition in the proglacial zone during flood. These properties are important in glacier-fed mountain rivers, where high sediment loads and channel instability create engineering and management problems for hydro-electric schemes (Bezinge, 1987).

Three major flood events occurred in the Bas Arolla proglacial zone during the 1986 and 1987 ablation seasons (Table 6.1). Two of the events were 'natural' in the sense that they were generated from a combination of water stored in the the glacier and heavy rainstorms

Table 6.1 Major Flood Events Bas Glacier d'Arolla Proglacial Zone 1986 and 1987

DATE	PEAK DISCHARGE $m^3\ s^{-1}$	RAINFALL mm	FLOW PATH	ORIGIN
<u>1986</u>				
July 6th	7 to 9	40.3 (14h)	Snout east margin	Outburst
<u>1987</u>				
July 15th-18th	10	15.6 (17th) 20.1 (18th)	Snout east margin overflow	Outburst
August 24th	20	Prolonged intense rainstorms	Rear of glacier east margin	High discharge Release of HEP water

(Table 6.1), whereas the flood event on August 24th 1987 was partly fuelled by natural waters but these were inflated by the artificial release of water from hydro-electric storage galleries in the Haut Arolla catchment. Flow from the two types of event was routed by two principal pathways: subglacial and ice-marginal - supraglacial. The August 1987 event is not considered further in this chapter since it occurred outside the period of field study.

This chapter begins with a description of the Bas Arolla proglacial stream (Section 6.2). This is followed by (Section 6.3) by a consideration of daily changes of cross section sediment storage from three channel sections on the main stream channel (Figure 6.1). Section 6.4 examines channel changes related to floods by considering changes in channel size and geometry at 28 sections before and after the July 1986 flood (Section 6.5.1) and contrasts in channel response between the July 1986 and July 1987 flood events (Section 6.5.2). Based on 28 surveyed cross sections, the amount of sediment contributed to the stream from the valley train, by channel processes, is computed (Section 6.6). Section 6.7 provides a summary of the major findings of this chapter and an evaluation of the Trimble (1988) model for sediment storage in mountain streams.

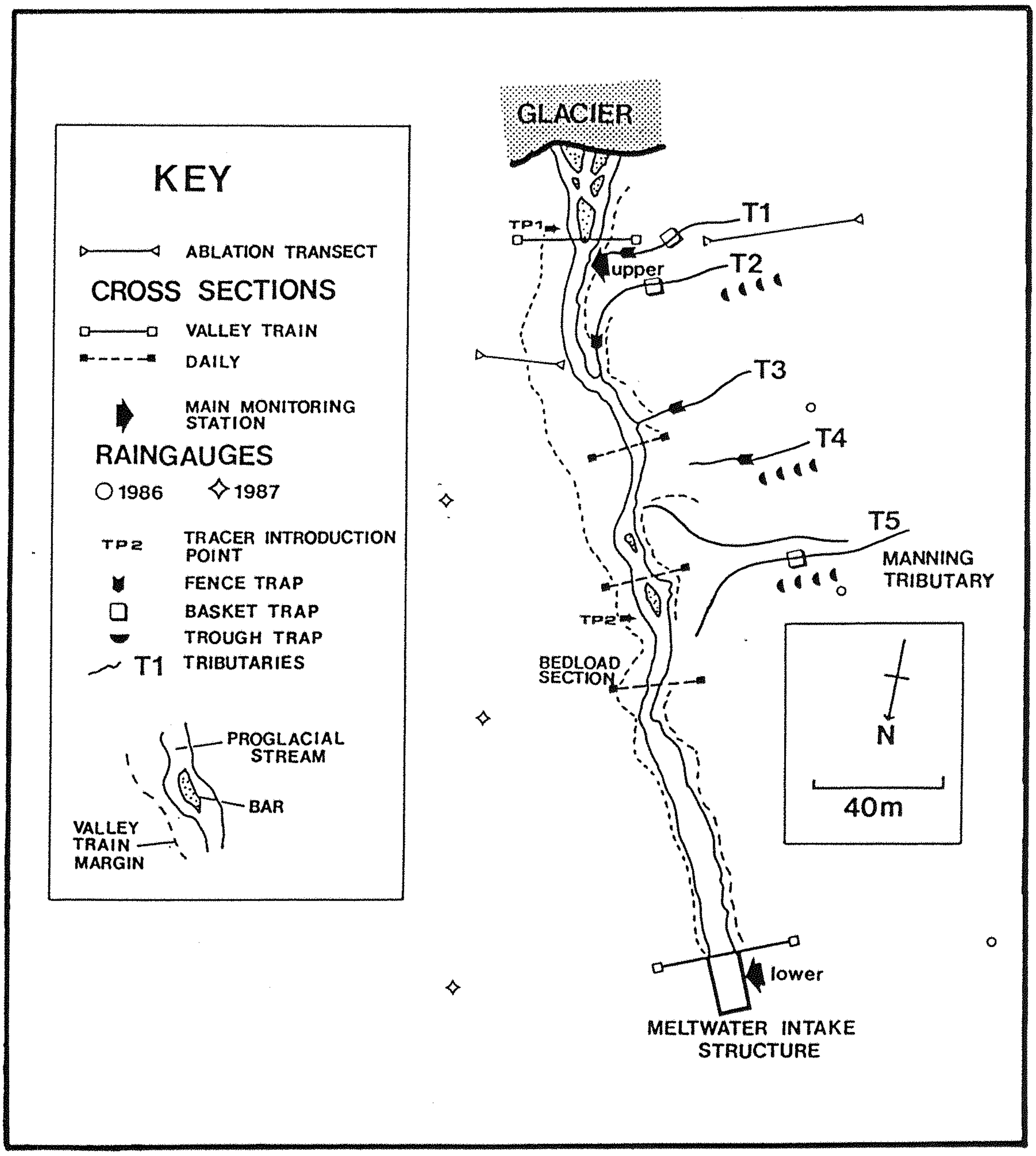
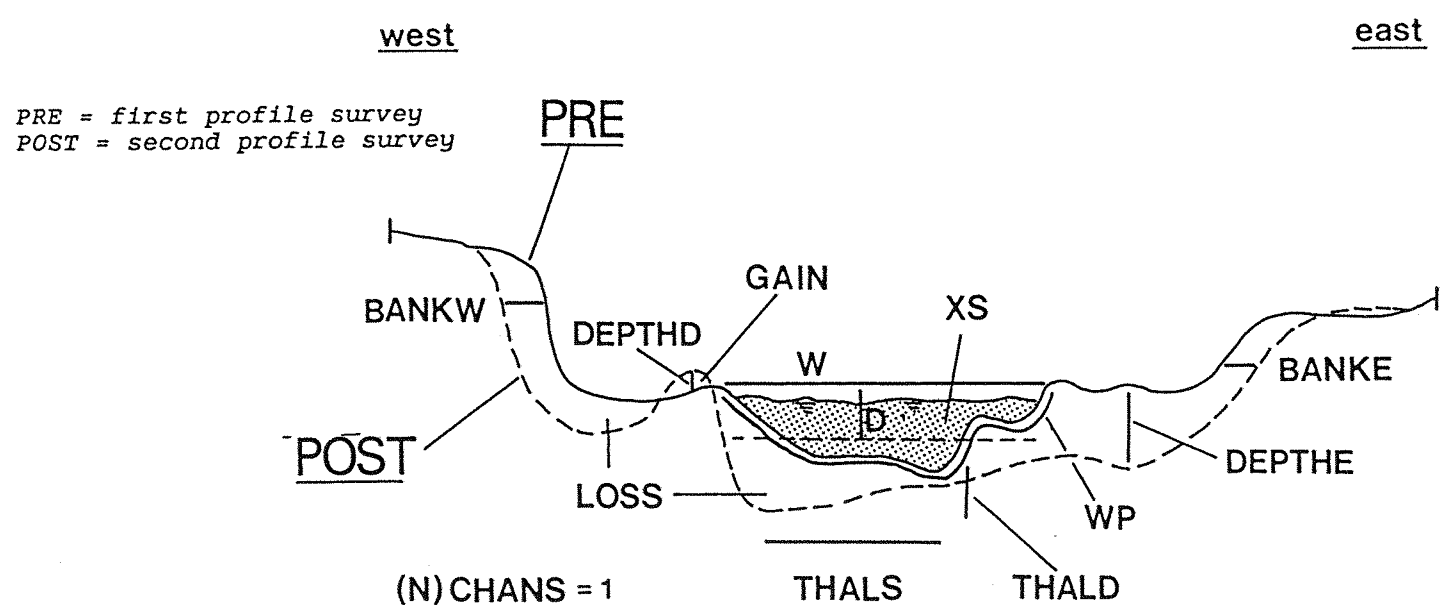


Figure 6.1 Main monitoring and measurement sites Bas Arolla proglacial zone 1986-1987. Locations of the upper and lower valley train cross-sections shown and the three detailed channel cross-sections.

6.2 Description of channel form

Changes in stream channel pattern and form, were monitored in the Bas Arolla proglacial valley train (Figure 6.1) in 1986 and 1987. Figure 6.2 defines the variables used to describe changes in channel form. Form variables follow the same definitions as Richards (1982, Table 1.2) but, channel change variables are more specific to this study. Width (W), wetted perimeter (WP) and cross section area (XS) refer to the morphology of the channel defined by the bank tops not the flow area. The number of channels (N) is the number of channels present in the valley train cross-section. Average channel depth (D) is estimated directly from the surveyed cross sections. Width/depth ratio (W/D) and hydraulic radius (HR) are derived from measurements of W, D, XS and WP. Channel change variables are estimated from 28 monumented cross sections (Werrity and Ferguson, 1980) (Figure, 6.1). All downstream distances shown in the diagrams (e.g. Figure 6.13) refer to the distance measured downstream from the upper cross section which is located 26 m down from the glacier snout. Monumented cross section surveys were carried out between June 3rd and 6th and July 8th and 9th 1986 and July 3rd and 4th and July 21st and 23rd 1987 in order to monitor the main channel change caused by flood flows. Bank erosion was measured as the distance of recession of the valley margin bluffs on east banks (BANKE) and west (BANKW) banks. Bank erosion is defined as erosion to the valley train margins. Streambank erosion within the valley train is considered in relation to the movement of the thalweg. Maximum depth of erosion (DEPTHE) and deposition (DEPTHD) refer to the maximum negative and positive deviation in the valley train surface observed at a section between surveys. Thalweg refers to the lowest point in the channel cross-section which corresponds to the maximum depth of flow. Movement in thalweg is defined by lateral shift (THALS) and vertical



CHANNEL FORM

W = width
D = average depth
XS = cross-sectional area
WP = wetted perimeter
(N) CHANS = number of channels

CHANNEL CHANGE

BANKW = west bank erosion
BANKE = east bank erosion
DEPTHE = maximum depth of erosion
DEPTHD = maximum depth of deposition
THALD = lowering of thalweg
THALS = shift in thalweg
LOSS = gross sediment loss from section
Gain = gross sediment deposition in section

Figure 6.2 Definition diagram of channel form and channel change variables.

changes (THALD). Sediment loss refers to the area (m^2) of sediment removed from a cross-section during flood (LOSS); and sediment gain is the area of deposition (GAIN). The balance between gain and loss defines the net sediment balance for the cross section.

Errors associated with estimating these variables depend on both operator error (the ability to survey all relevant morphological features) and measurement error. Measurement error is estimated from re-survey experiments to be +/- 1.0 to 2.0 cm vertical determination and between 1 and 5 cm for horizontal distances (depending on the surface roughness of the surveyed reach and the length of the cross section). These errors appear acceptable given the relatively large magnitude of the changes measured.

Several distinct morphodynamic zones exist within the valley train of the Bas Arolla proglacial zone and are characterised by distinct channel segments. The valley train, shown in Figure 6.3, extends approximately 320 m from the snout of the Bas Glacier d'Arolla to the meltwater intake structure. Valley train width is approximately 30-40 m, with the widest zone in the central channel section (width up to 50 m) and with the narrowest section just above the water intake, where channellisation has reduced the width to less than 20 m, (Figure 6.3). The average slope of the valley train is 0.07 $m\ m^{-1}$ with the steepest section immediately upstream of the lowest channellised reach. The lowest gradient section is in the central reach, which corresponds to the widest section of the valley train.

Channel pattern is low sinuousity, single thread with three "nodes" of braiding: immediately below the snout, in the central wide section and just above the lower channellised reach (Figure 6.3). The presence of braiding in the central section of the stream is interesting since this is a recurrent feature which is

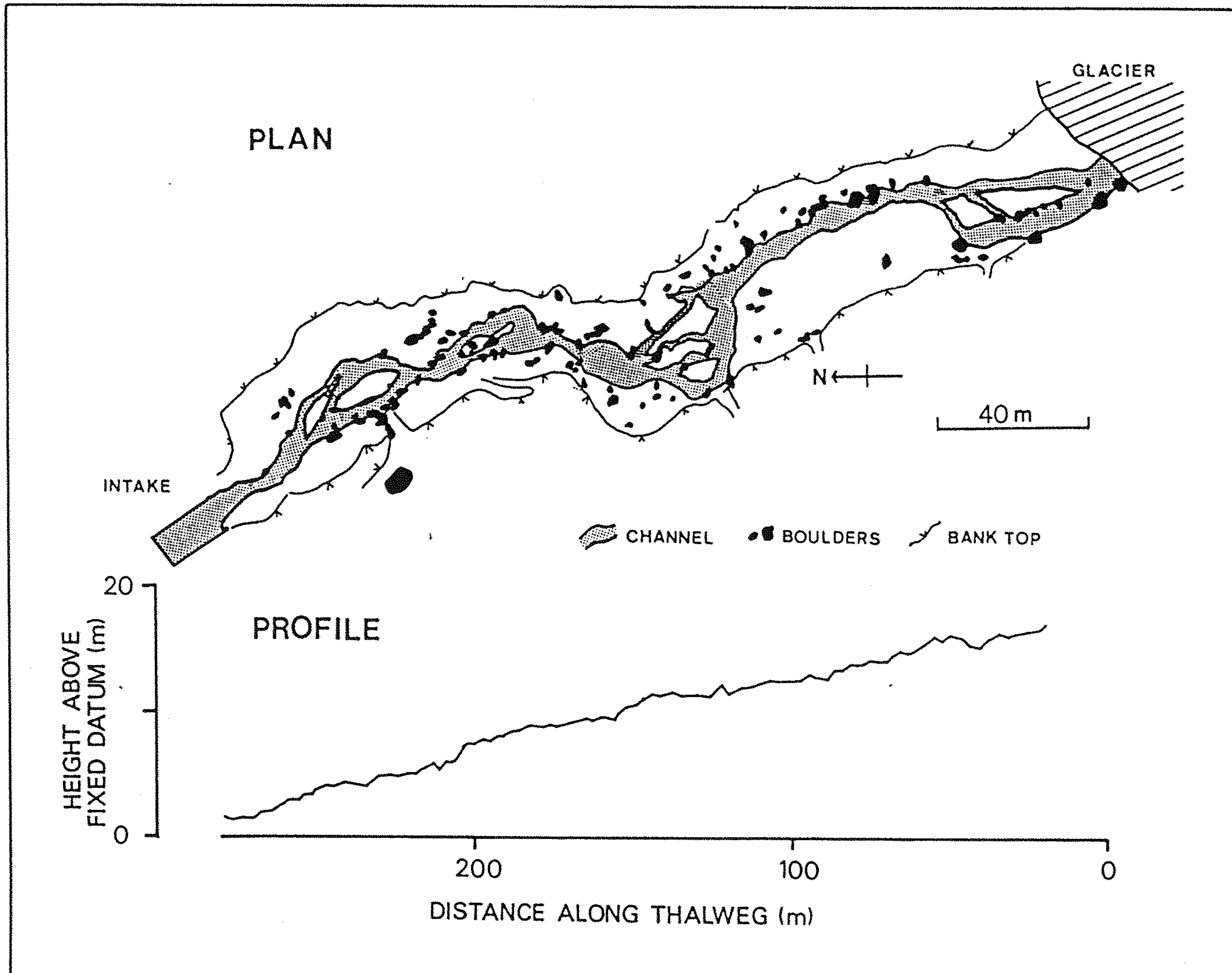


Figure 6.3 The Bas Arolla proglacial stream channel, valley train and long profile, September 1987.

very similar to the central braid bars discussed in flume studies by Leopold and Wolman (1957) and Ackers and Charlton (1970). As such, this may be an artifact of the entrance conditions imposed as the proglacial channel emerges on to the valley train. The average width of the channel varies with discharge, but is generally between 4 and 7 m except in the central channel reach where width is up to 12 m (Figure 6.3). Survey of the stream long profile reveals that in the steep single thread sections a step-pool morphology is developed (Figure 6.4). This type of structure is a fundamental control on sediment transport processes. (Sawada et al., 1983; Whittaker, 1987). In general the frequency of step-pools increases within the steeper sections of the channel (e.g. 200-220 m, Figure 6.4). In shallow gradient reaches (e.g. 80-120m, Figure 6.4) pools are few and the channel is free of large structural features. There is a large range in channel sediment size from very fine material (as shown in Chapter 4) to large boulders, some of which are of sufficient size to be mapped (Figure 6.3). In the absence of appreciable vegetation, the full range of sediment is potentially available. Some will be transported and some of which will be incorporated into the channel as large-scale roughness elements, which will in turn influence the channel structure and nature of the flow (Bathurst, 1978).

6.3 Short-term changes in channel cross sections.

Daily monitoring of channel cross-section change was undertaken at three sites (Figure 6.1) from 27th May to 30th July, 1987. Gaps in the record exist where, high flows prevented re-survey. Cumulative change in sediment storage (Figure 6.5) gives the impression that for the majority of time the channel is relative stable. This stability was dramatically punctuated by the flood of 15th to 18th July (Figure 6.5), after which channel

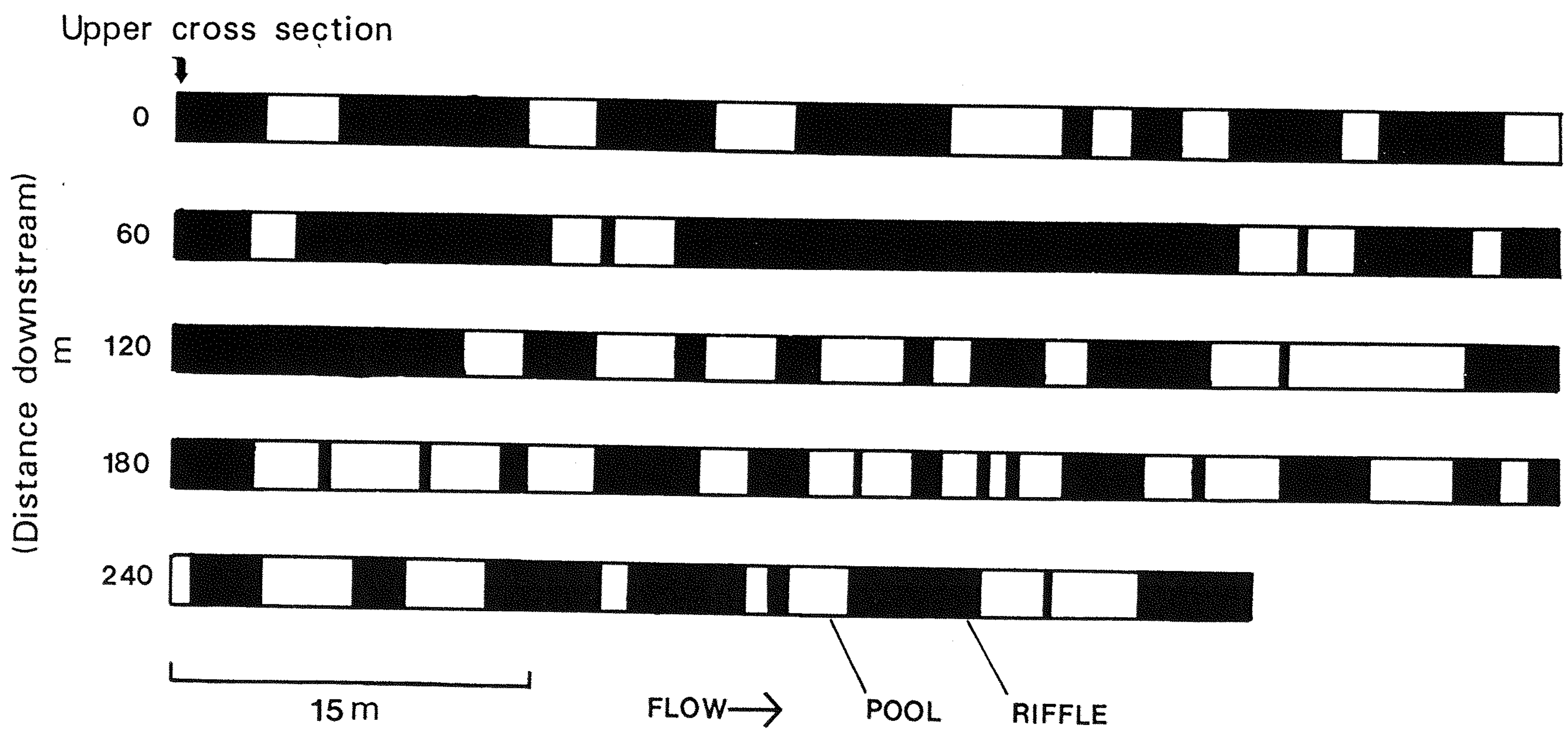


Figure 6.4 Diagram of pool and riffle step sequence in the Bas Arolla proglacial stream. All distances measured from the upper valley train cross-section, (Figure 6.1).

behaviour was highly erratic, with an order of magnitude change in channel storage. Individual cross-sections responded very differently to the flood both in the magnitude and the direction of their response (e.g. the upper section experienced net erosion before the flood but under went massive aggradation during the flood, whilst the lower section was slightly aggrading before the event but was very heavily eroded during the flood). Post-flood adjustment was rapid and tended to be in the opposite direction to the change imposed by the flood event (Figure 6.5). Observations ceased before the full pattern of re-adjustment was established.

The major implication of these observations is that channel adjustment does not continue at a steady rate. Small scale adjustments frequently occur, but it is the episodic flood events which are of major importance. This point is emphasised in Figure 6.6 which shows cross-sections at the lower daily survey site (Figure 2.1) which were re-surveyed at time intervals of 10 minutes (7 sections); 1 hour (12 sections); 5-6 hours (3 sections) and 24 hours (10 sections) over different time periods. Overall changes at all intervals were small with only minor adjustments in the bed. Indeed changes within an hour, day and over 10 days were virtually identical although a range of bedload transport rates (0 to 1.745 $kg\ m^{1}\ s^{-1}$) and water discharges (0.446 - 2.812 $m^{3}\ s^{-1}$) are represented. The only exception to this general pattern occurred on July 8th (Figure 6.6d) where, for this one day only, bed elevation was greatly increased. This corresponds to a day of very active bedload movement (almost an order of magnitude greater than on the other days) and of relatively high discharge. It seems that active bedload movement is accompanied by fluctuation in bed elevation but the profile soon re-adjusts to previous conditions.

Re-adjustment in bed storage can also be seen on a

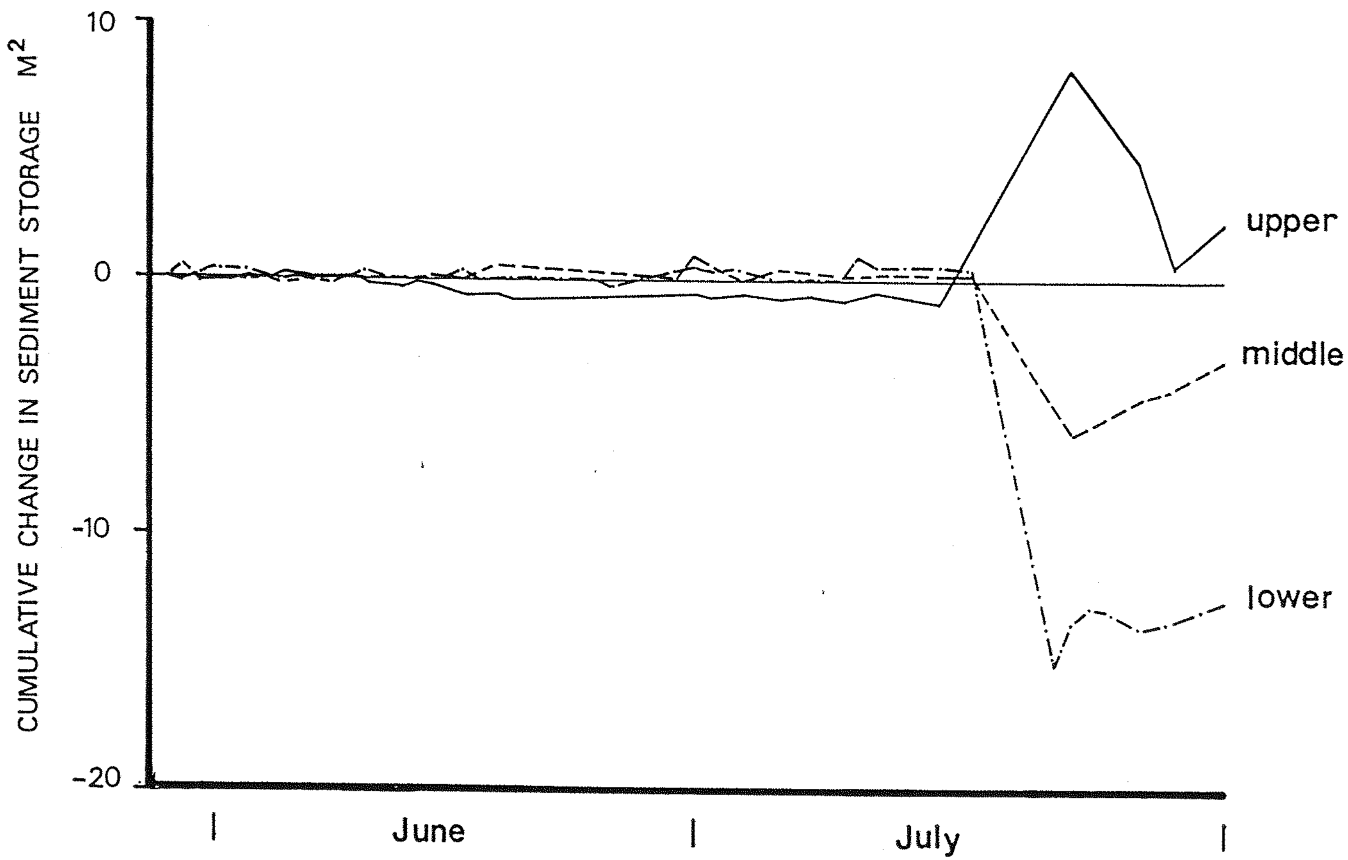


Figure 6.5 Detailed channel cross-sections of the Bas Arolla proglacial stream, (May 27th to July 30th, 1987) showing the cumulative change in sediment storage. Location of upper, middle and lower sections shown in Figure 6.1.

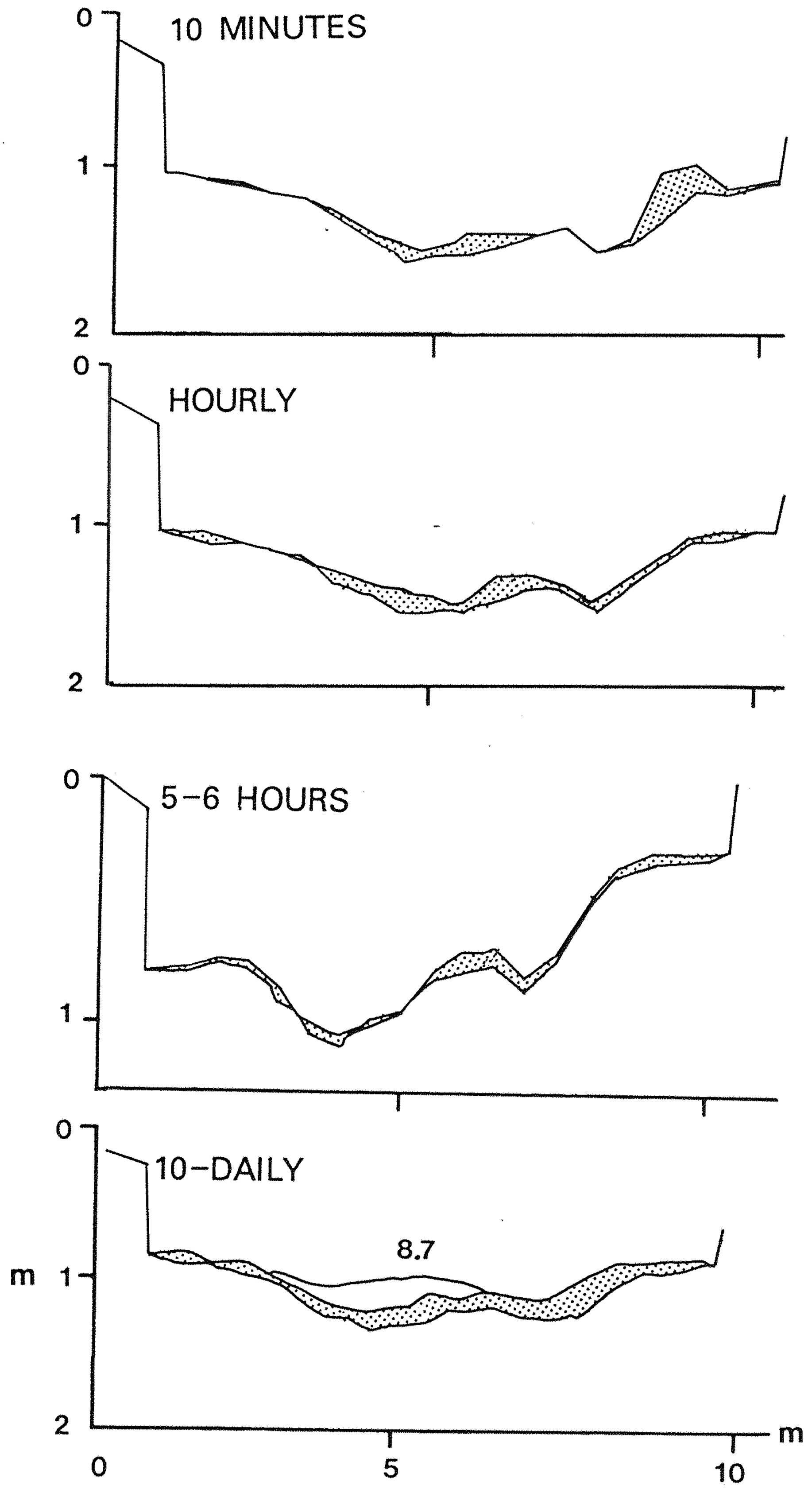


Figure 6.6 Changes in cross-sections at the lower site surveyed at 10 minute, hourly, 5-6 hourly and daily intervals.

longer time scale. The relationship between weekly cross-section sediment loss and weekly bedload yield (Figure 6.7) suggests a positive relationship between the two variables. This is expected, given that the channel is actively contributing sediment to the stream load. A line joining the data points in their temporal sequence gives the impression of a crude anti-clockwise hysteresis loop. This loop may be explained by the fact that early in the season bedload transport was inhibited due to the stable bed conditions. Following destruction of the bed armour and channel form by flood, both bank and bed sediment availability were greatly increased and so, sediment supply was increased and bedload transport rates were higher (Figure 6.7).

These changes seem to indicate episodic behaviour in channel change and channel bed elevation (Trimble, 1988). During flood these changes were abrupt, as if the channel had crossed a threshold (Schumm, 1968). Such a threshold is probably related to the threshold of full bed mobility, in response to extrinsic forcing by a critical discharge. The critical threshold discharge for the mobilization of boulder clusters, the most stable bed element, was calculated at between 8.5 - 12 $m^3 s^{-1}$ (Section 4.4), which approximates the discharge of the 15th to 18th July flood.

Changes at the three sections discussed above may not be representative of changes within the proglacial stream as a whole since the three cross sections were located on channel segments which were persistently single thread.

6.4 Channel pattern changes

The previous section demonstrated that at-a-station changes in cross-section were greatly influenced by episodic flood events and that large differences existed in the response of single stations to the same event.

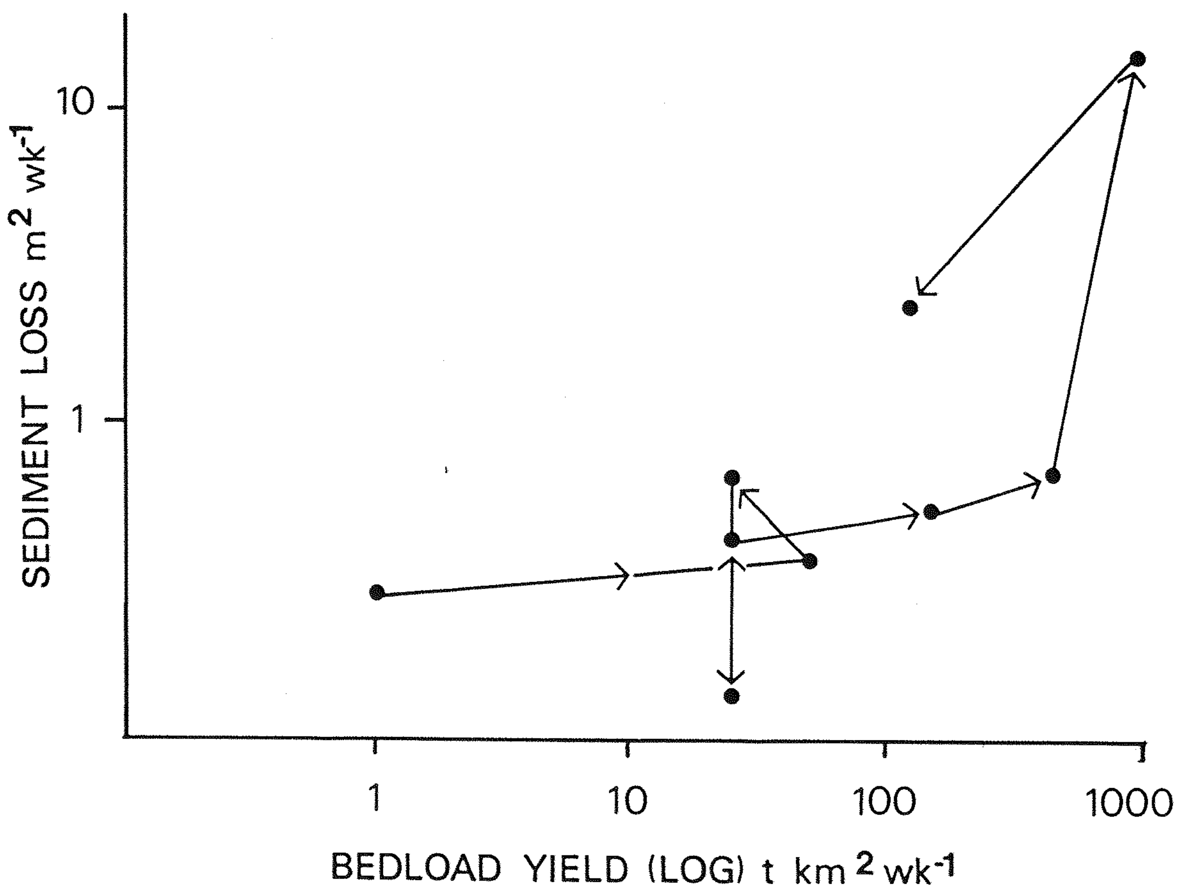


Figure 6.7 Bedload (gravel trap) yield in relation to sediment loss from the Bas Arolla proglacial stream lower cross-section.

One way in which spatial variations manifest themselves is in channel pattern. Channel patterns determined from planimetric maps (Figure 6.9) and oblique, high-angle, aerial photos (e.g. Figure 6.10) were used to assess the major changes in pattern over the 1986 and 1987 ablation seasons. Planimetric maps were based on field survey of morphological changes in the proglacial zone, whereas air-photo channel patterns were based on photographs where the edge of the channel is defined by the water edge. However, a pre-requisite for using the photo-maps was that the observed channel patterns were stable for the complete day of the photography ie. field observations show that the same pattern was maintained throughout the full diurnal discharge cycle. The distinction between stream-edge and channel-edge is important since this may lead to discrepancy in evaluating the channel pattern. Figure 6.8 shows a plot of channel bank widths (morphological width defined by the bank tops) and the water surface stream widths at the lower main cross section for a range of discharges. The regression line is oblique to the line of perfect agreement and the intersection of the two lines defines the channel bankful width (6.1m). To the right of the diagram data points plot below the 1:1 line, indicating that water surface width is greater than bankful width; to the left water surface width is less than bank width. The scatter of points is significant, since the close clustering of points to the left of the graph indicates that water surface width is within bank and is constrained to a minor degree of variation by bank sides, to the right the water surface is 'overbank' and greater variation occurs e.g. there is a "pan-handle effect" (Costa 1983). These conditions only apply to this one site different relationships will exist at other sites. In some proglacial - braided streams the channel is 'perched' above the adjacent floodplain and under these circumstances the overbank effect will be very much greater. This discussion illustrates that care must be taken in interpreting channel patterns from

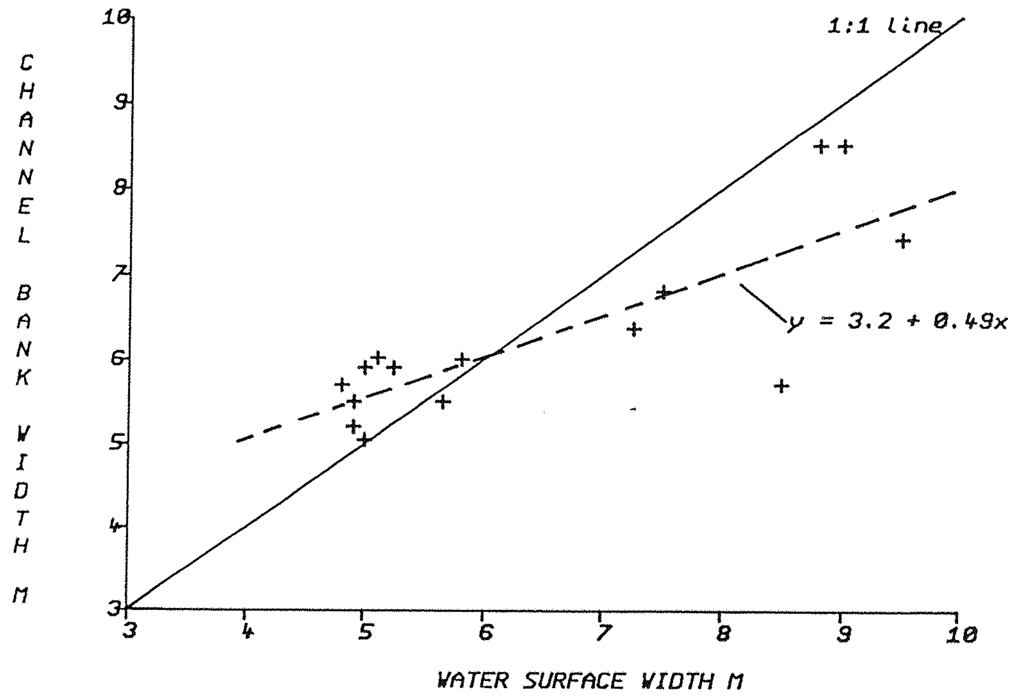


Figure 6.8 Relationship between channel bank width and water surface width at the lower cross-section, Bas Arolla proglacial stream.

different data sources.

Planform maps were constructed from plane table mapping for four dates (Figure 6.9: June 5th 1986, September 8th 1986, June 12th 1987 and September 7th 1987). These maps show the morphological channel margins and the extent of the valley train. In June 1986 the proglacial channel of the Bas Glacier d'Arolla was a sinuous, single-thread stream with minor braiding in the zone immediately in front of the glacier snout. In September the pattern was greatly modified. The extent of braiding was greatly increased in the area downstream of the snout, at the head of the wide central reach and lower steep section where the channel was divided by a long sinuous bar. The channel width was also generally much greater than in the June survey.

In June 1987 (Figure 6.9) there was a return to the simple, low sinuousity, single-thread channel similar to the situation in June 1986. The channel was displaced towards the east in the middle reach and braiding was limited to a few small mid-channel bars. Snow masked the the front of the glacier so that the channel pattern could not be seen. The channel was very different from September 1987. Width was greatly reduced and many of the bars had disappeared. The pattern in September 1987 showed a great increase in channel braiding and channel width; mid-channel bars occur widely, especially in the central, wide section of the reach. Sinuousity had also increased.

The general channel pattern in these four surveys was relatively: a low-sinuousity, single-thread channel early in the season, with few bars, followed by a late season channel which has greater width and braiding. Braiding tends to be focussed in areas of 'bulbs' or 'nodes' along the channel. The greater extent of braiding in September 1987 was probably due to the recent big flood (for the September surveys, the most

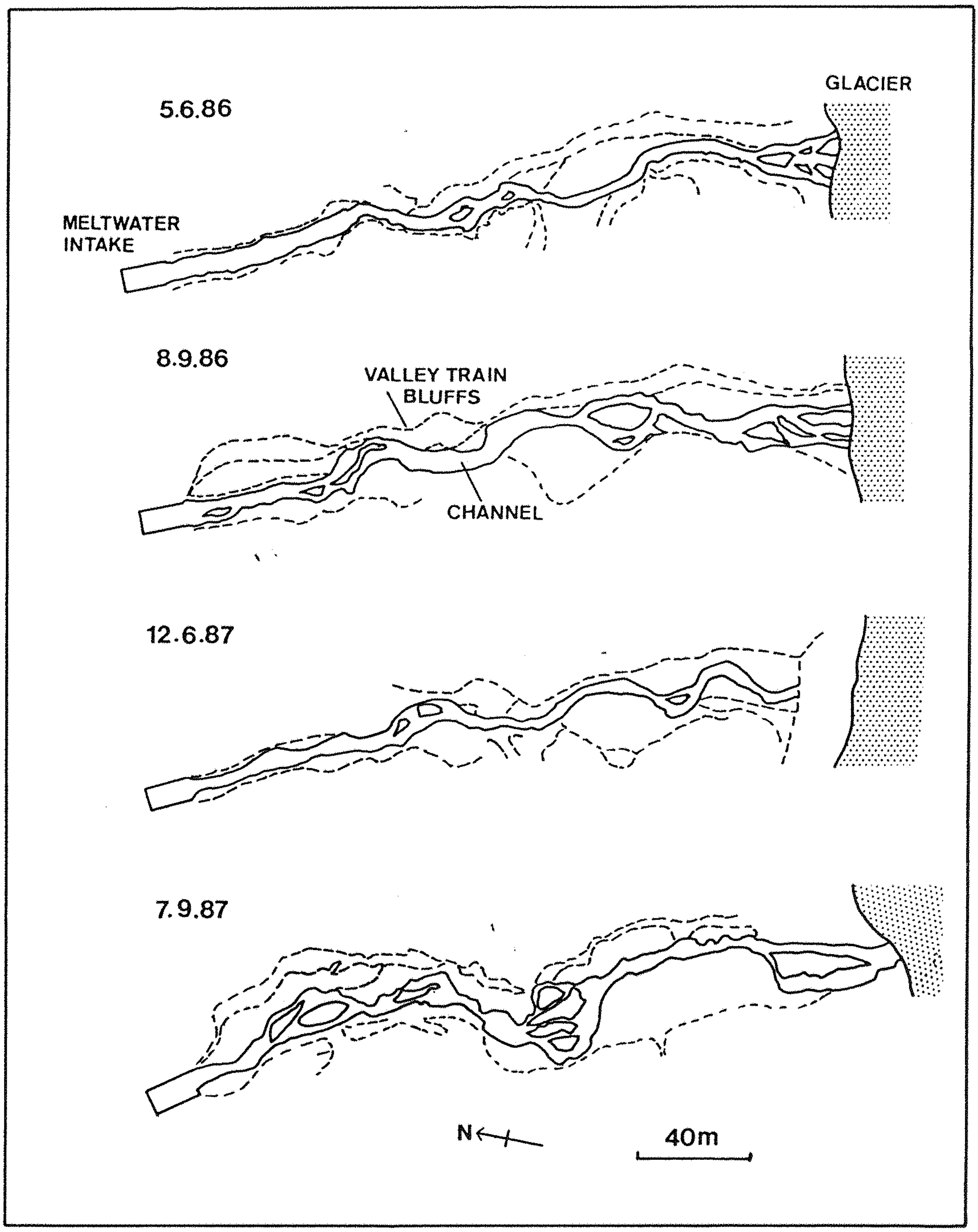


Figure 6.9 Planform map of the Bas Glacier d'Arolla proglacial stream channel and valley train.

recent flood in 1986 was on July 6th whilst in 1987 whereas floods occurred on July 15th and August 24th. Over the winter there appears to be a rationalisation of the channel pattern, a simplification of the channel form and a reduction in channel width. Fluvial processes after the time of the survey will be responsible for a part of this over winter change, but snow may also be an important agent. Snow fills the valley train to a depth of several metres which has three major effects: firstly the banks and valley margin bluffs are eroded by a variety of snow-related processes (Chapter 7); secondly, snow has a compacting effect on the underlying channel; and thirdly, the presence of snow banks early in the season influences the evolution of channel pattern by diverting flow. The second effect, although not investigated may be important since in the presence of a snow cover and stream flow 'alpine sub-nival boulder pavements' may develop (Hara and Thorn, 1982) The effect of this would be to produce a flattening of the valley train topography and disruption of the pre-existing channel pattern.

These maps in Figure 6.9, although useful, only provide a start and end of season view of the proglacial stream form. More detail is required to adequately characterise intra-season variations in channel pattern. One method of doing this is by means of sequential photography, which is a rapid, cheap technique which allows a frequency of observation not possible with conventional mapping. However there is a trade-off between accuracy and frequency of survey. In particular the two techniques may indicate slight differences in pattern in relation to the definition of channel width (Figure 6.8).

Channel patterns identified from photographs are shown for 1986 (Figure 6.10) and 1987 (Figure 6.11). The 1986 channel surveys (Figure 6.10) are not at equally spaced time-intervals, but were selected from a larger set of

CHANNEL PATTERN CHANGE 1986

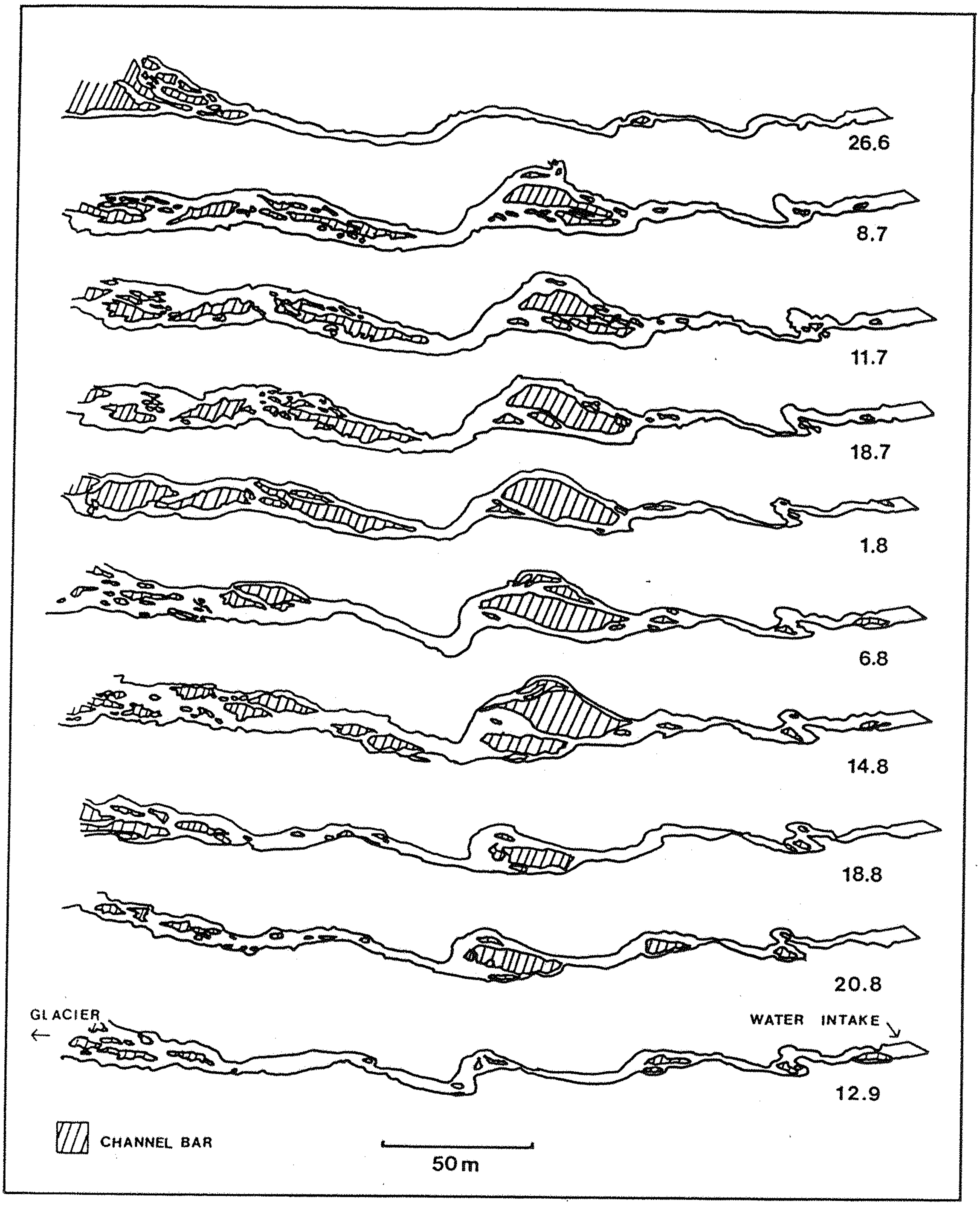


Figure 6.10 Channel pattern change 1986 (Mapped from oblique high-angle photographs).

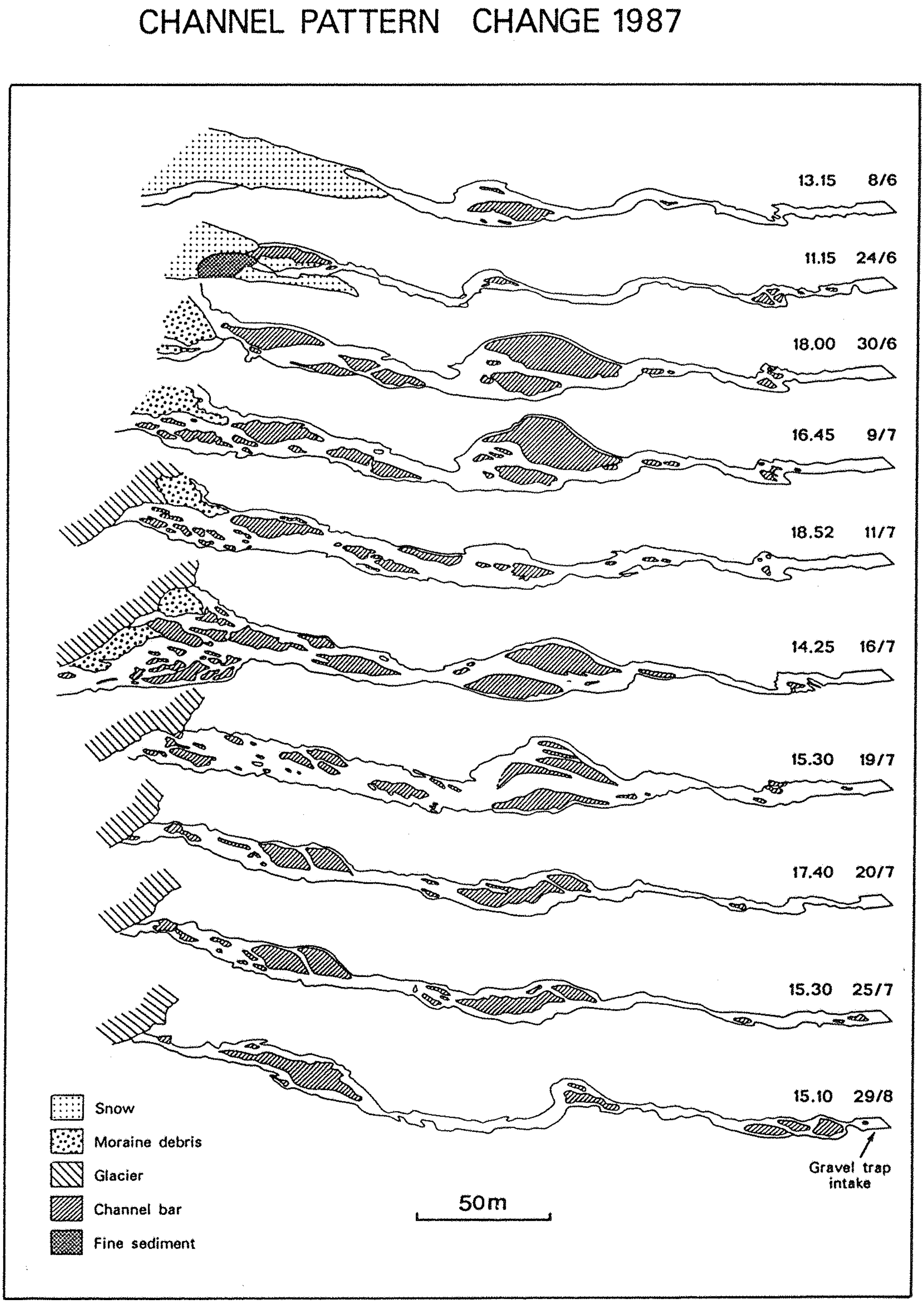
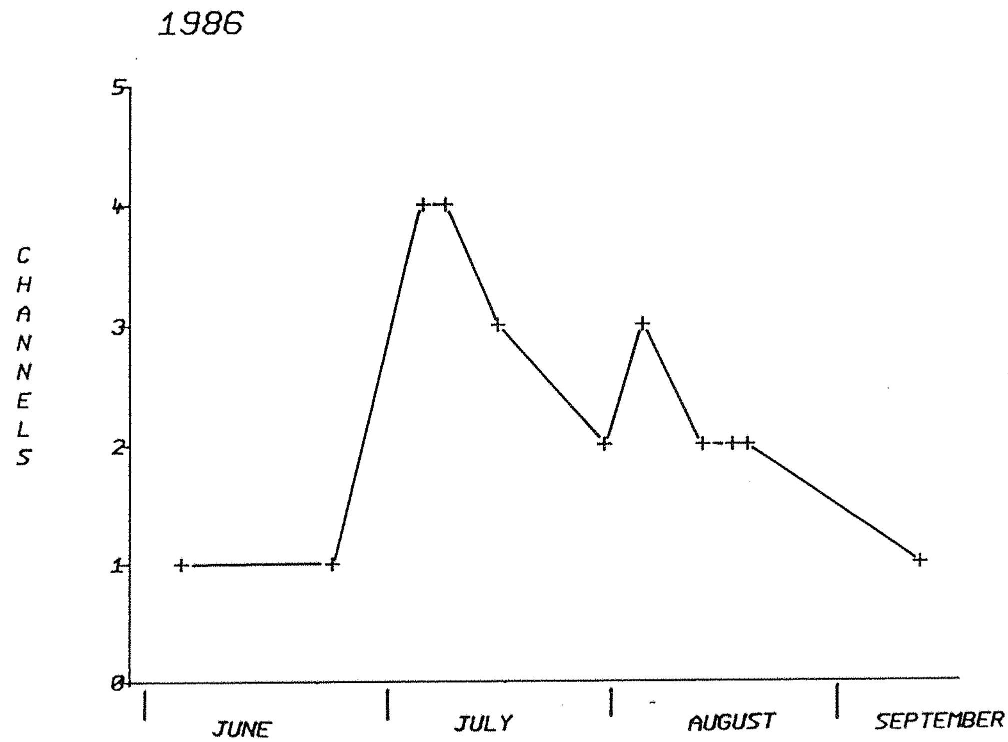


Figure 6.11 Channel pattern change 1987 (Mapped from oblique high-angle photographs).

photographs to give the best impression of change. The pattern was the same as in (Figure 6.9) from early to late June. Following a large flood event on July 6th, braiding increased dramatically in both the reach downstream of the snout and in the wide central section. This pattern was maintained with only minor modification until the middle of August (i.e. throughout the period of the highest meltwater discharge). From the middle of August onwards the channel pattern rationalised and many of the channels were abandoned. By August 16th the outer braid channel of the central bar had been abandoned and the braiding in front of the glacier snout was substantially reduced. Further reduction of braiding occurred by September 12th so that the channel was single-thread with only minor mid-channel bars. The overall channel pattern sequence in 1986 shows a low sinuousity, single thread channel switching to a braided habit as a result of a large flood. This pattern was maintained throughout the middle of the meltwater season, but the pattern had reverted to single thread by the start of September.

The situation in 1987 was slightly different (Figure 6.11). The initial channel state in June was braided in the central reach, but by the end of June this had changed to a single-thread pattern. Increasing discharge at the very end of June and in early July lead to anabranching (Brice, 1984) in the central reach. By July 11th the outer braid channel of the central section had been abandoned, but the upstream braid pattern had been maintained. Flooding between July 15th and 18th led to the braiding, in the zone immediately in front of the glacier and in the central reach. Channel change continued throughout the flood period and by the 19th July the pattern was very different from that prior to flooding. Upstream braiding was less pronounced, but the central braid system was highly complex. This pattern was very short-lived and by 29th July the central braid complex had been significantly rationalised and upstream braiding had been greatly reduced. Until July 25th this pattern was maintained. The observed channel form on 29th August, at the end of the meltwater season, showed little correspondence to the July pattern. This was because a flood event on the 24th August caused massive channel change eroding all previous channel structures. The effects of the flood appear to be a reduction in braiding in the central reach and an increase in sinuousity and in braiding above the meltwater intake.

Attempting to summarise the complex nature of river pattern change in a single index is difficult. However, the number of channels which occur in a transect across the central braided reach of the Bas Arolla proglacial stream may be a useful index since this is the section of channel which appears to be most dynamic in respect to changes in discharge and in sediment load. Using this index, Figure 6.12 shows that in 1986 the degree of braiding in the central reach followed a relatively simple pattern, moving from single-thread to multiple channels in response to the major flood and then reverting to a single-thread pattern again. The 1987 situation was more complex. An initially divided channel changed to a single thread condition followed by a period of fluctuating braiding intensity. No detailed channel history is available for August when the large flood event of 24th occurred but, by the end of August, the channel was again single thread. Cross-reference to the plane table map for September 1987 (Figure 6.9) still shows the presence of the morphology of a braid system in the valley train although but this was not always occupied by water. Thus, although the channel photo-maps are not truly independent of discharge and so do not indicate full details of channel morphology, they are useful in providing an indication of the rate and pattern of proglacial channel change.



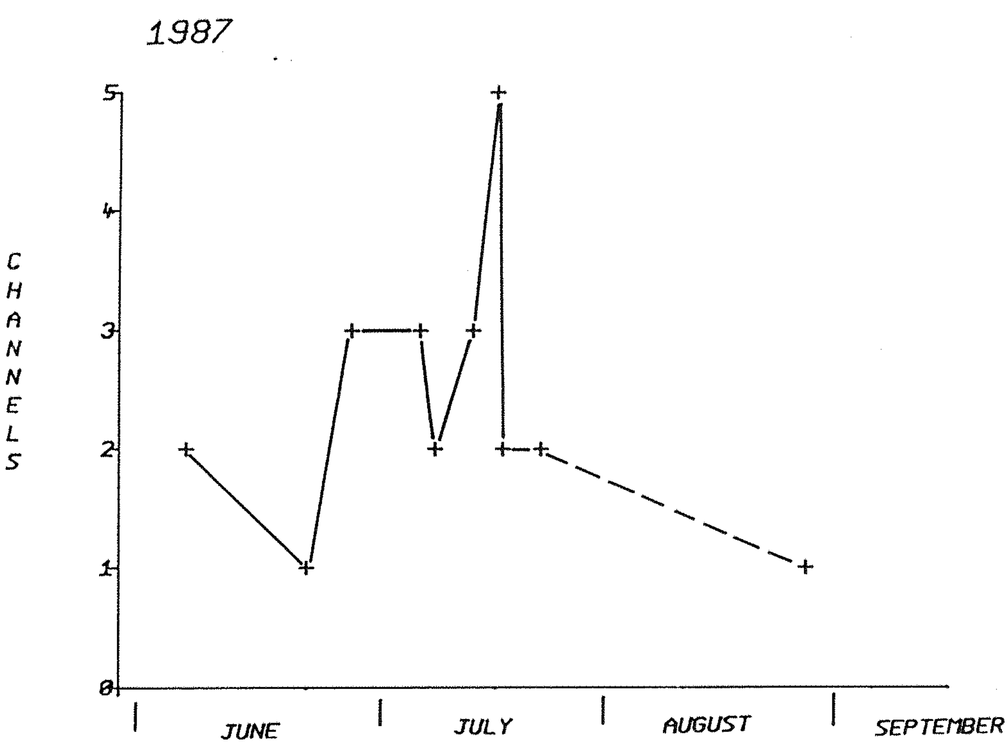


Figure 6.12 Variation in the number of braid channels in the central reach of the Bas Arolla proglacial stream 1986 and 1987.

6.5 Channel changes related to flood events

In the previous two sections, the importance of floods in transporting large amounts of sediment and in producing channel change was evident. This section discusses the direct role of floods and refers to changes brought about in channel geometry during the 1986 flood (Section 6.5.1) and the changes in the morphology of valley train cross sections in both the 1986 and the July 1987 floods. These changes will subsequently be related to the sediment balance of the proglacial zone (Section 6.6).

6.5.1 Adjustments to channel geometry in the 1986 flood

During flood flows, channels respond to the newly imposed water discharge and sediment load by adjustments in size and shape (Petts and Foster, 1987). Changes in channel geometry at 28 sections during the 1986 flood are shown in Table 6.2 and in Figure 6.13. Overall the magnitude of change is great (Table 6.2) with all post flood channel variables greatly increased by between 22 and 88%. These mean values (Table 6.2) show a greater increase in width than depth, which is clearly reflected in increases in the width/depth ratio, the wetted perimeter and the number of channels. Channel depth shows the lowest percentage change of all variables. The increase in hydraulic radius suggests that the new channel forms may be more efficient which may indicate a tendency towards an optimum shape. These statements are tentative since the values shown in Table 6.2 have large standard errors suggesting large variance between cross sections (i.e. downstream). Because of the heterogeniety of channel bed materials in the proximal proglacial zone, wide variations in hydraulic geometry are to be expected since these changes are explicable in terms of the hydraulics and geometry of each cross section (Ferguson, 1986b).

Table 6.2 Change in mean channel geometry variables at 27 valley train cross-sections before and after the July 6th flood 1986

Variable	Pre-flood	SE+/-	Post-flood	SE+/-	% change
W/D	14.93	6.1	19.00	11.9	+ 27
Width	5.40	1.7	8.44	4.1	+ 56
Depth	0.41	0.19	0.50	0.21	+ 22
Cross-section area	1.31	0.67	2.46	1.27	+ 88
Number of channels	1.15	0.36	1.52	0.70	+ 32
Hydraulic radius	0.17	0.04	0.21	0.06	+ 24
Wetted perimeter	7.65	2.68	11.78	6.87	+ 54

Channel cross sections (downstream distances measured from the top cross section which is 26.5 m from the glacier snout) show marked variations in geometry before and after the flood (Figure 6.13). Post-flood width is generally greater than pre-flood width except in the lowest reach. The magnitude of width variations is great. The channel section between 80 and 120 m (Figure 6.13) shows the greatest increase in width. The pattern in the rest of the channel tends to mirror the pattern from the pre flood series but with a slight increase in width. Depth shows much less variation, except for marked fluctuations in the zone from approximately 120 -190 m. Hydraulic radius, although generally higher in the post-flood series varies much less consistently between pre and post flood surveys than the other variables. In the lower reaches the hydraulic radius shows the most marked increase, probably related to greater channel depth changes and less variation in width. All 3 variables are generally greater for the post-flood series, reflecting a channel adjusted to increased water and sediment discharge.

Interpretation of these trends is aided by comparing the series with downstream changes in stream power/unit bed area (Figure 6.13) (Langbein, 1964). The distribution in stream power is relatively high in the upper reach but drops dramatically at 80 m. This drop corresponds to the start of the large width increases in the post-flood series. Beyond 100 m stream power shows a generally increasing trend accompanied by a reduction in the magnitude of width variations and an increase in channel depth. Small variations in stream power (i.e. peaks and troughs) do not directly correspond to variations in morphological variables. This may show that the resolution of the series is inadequate to indicate these, or it may result from leads and lags in the channel response (i.e. there is space-time serial dependence in channel response (Ferguson, 1976;

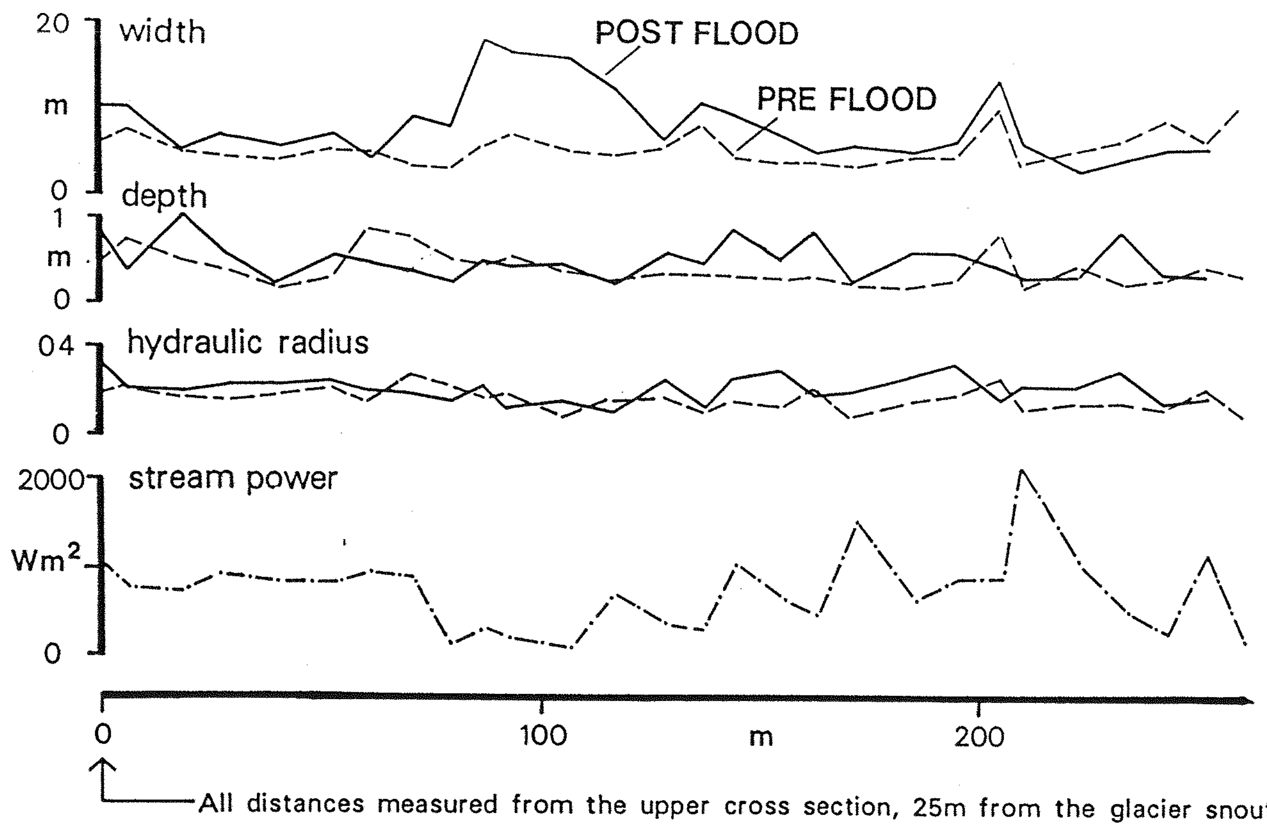


Figure 6.13 Pre and Post 1986 flood channel characteristics surveyed in the 27 valley train cross-sections in the Bas Arolla proglacial zone. The top cross-section is located 26.5m from the glacier snout (Figure 6.1). Stream Power is calculated based on survey of channel characteristics and an assumed discharge of 5 cumecs.

Richards, 1982). Variations in stream power (albeit dominated by channel slope) seem to be important in controlling channel behaviour in the Bas Arolla proglacial stream. The influence of slope is demonstrated in the relationship between the area of the least pronounced step-pool bed structure (lower slope) (Figure 6.4) and the large increase in channel width associated with a reduction in stream power. This implies that the bed is more alluvial in this section and less constrained by the presence of large roughness elements. The correspondence between a general reduction in slope and change in bed characteristics is also related to an increase in valley train width at the same point. This poses a difficult question as to which variable is controlling channel form. However, the interdependency in the variables ensures indeterminacy in causality (i.e. no one variable is the sole cause of the observed form - therefore the features are polygenetic in origin).

In order to evaluate the within and between channel changes in more detail each variable was correlated with itself (before and after the flood) and with other variables at the same observation time (e.g. an approach similar to Newson and Harrison (1978)) Pre-flood variables (A) and post flood variables (B) are shown (Table 6.3) and significant correlations ($P < 0.001$) are under-lined. Figure 6.14 attempts to summarise significant correlations within and between the two sets of channel observations. This correlation systems diagram (Chorley and Kennedy, 1971) is suited to morphological systems where the covariability of variables is of interest. The linkages imply nothing about causality but provide an a-posteriori method of looking for minimal structure in the system (Bennett and Chorley, 1978). If causal modelling is the goal, multiple partial correlation techniques are appropriate - but are difficult to manage with a large number of variables.

Table 6.3 Correlation of 1986 Pre and Post-Flood channel parameters

	W/D-A	W/D-B	W-A	W-B	D-A	D-B	XS-A	XS-B	CHAN-A	CHAN-B	SLOPE-A	HR-A	HR-B	WP-A	WP-B
W/D-B	-0.212														
W-A	0.363	0.188													
W-B	-0.242	0.832	0.274												
D-A	-0.688	0.303	0.331	0.331											
D-B	-0.037	-0.548	-0.039	-0.153	-0.050										
XS-A	-0.416	0.405	0.456	0.443	0.773	0.032									
XS-B	-0.278	0.286	0.029	0.663	0.191	0.423	0.288								
CHAN-A	0.134	-0.078	0.494	-0.015	0.342	0.130	0.265	-0.141							
CHAN-B	-0.359	0.599	0.238	0.702	0.453	-0.216	0.442	0.464	0.141						
SLOPE-A	-0.099	-0.313	0.215	-0.211	0.238	0.005	0.070	-0.100	0.243	0.009					
HR-A	-0.476	0.178	0.079	0.138	0.558	0.043	0.748	0.151	-0.019	0.168	0.139				
HR-B	0.056	-0.614	-0.310	-0.361	-0.270	0.508	-0.291	0.305	-0.128	-0.307	0.240	0.038			
WP-A	-0.123	0.420	0.642	0.504	0.620	0.065	0.825	0.304	0.365	0.414	-0.152	0.291	-0.456		
WP-B	-0.290	0.715	0.203	0.847	0.335	-0.039	0.441	0.696	-0.067	0.643	-0.330	0.111	-0.353	0.544	
WM-2-A	-0.136	-0.425	-0.330	-0.367	-0.083	-0.038	-0.301	-0.179	-0.068	-0.185	0.802	-0.104	0.331	-0.504	-0.449

Underlined values are significant (P < 0.001) d.f. = 27-2

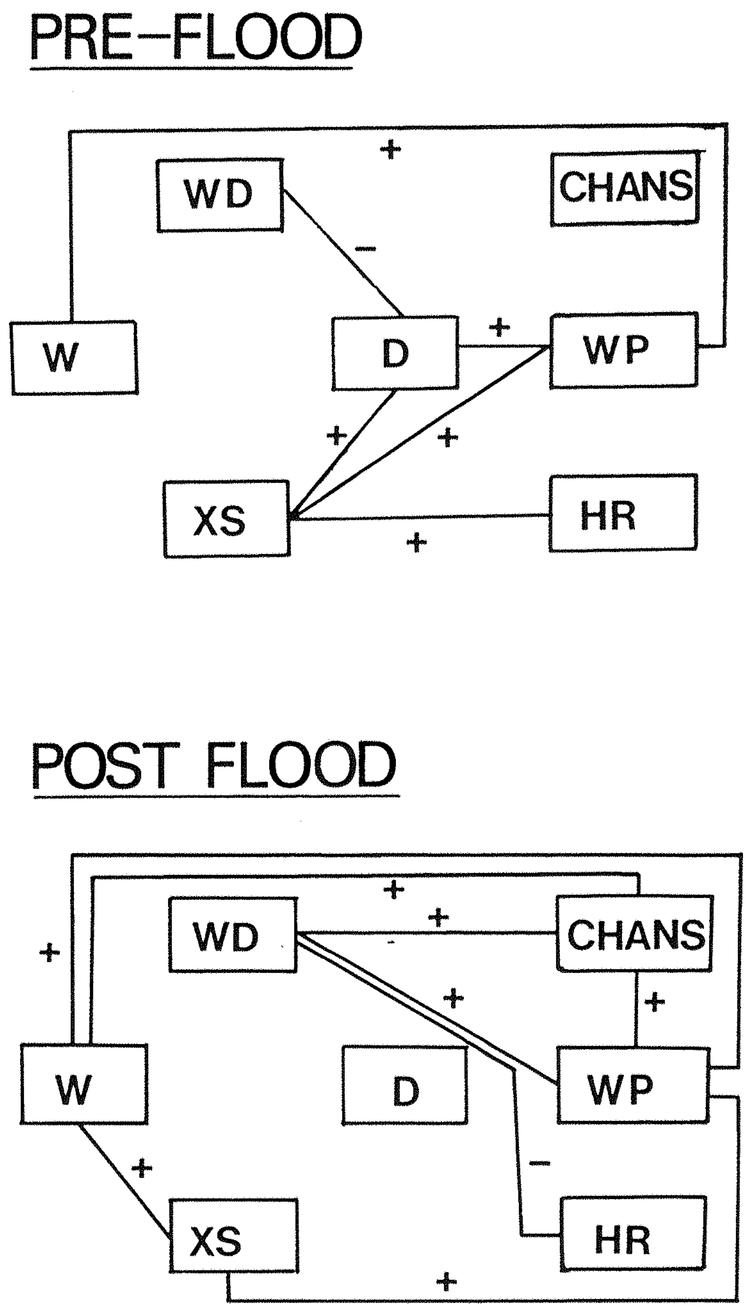


Figure 6.14 Pre and Post 1986 flood channel variable correlation systems, links relate to correlations shown in Table 6.3.

system. It could be argued that this represents a 'switching-on' of the proglacial channel system by the flood event. These generalisations are only true at the arbitrarily chosen significance level. By altering the significance level new patterns may emerge.

6.5.2 Contrasts in channel response to the 1986 and 1987 floods.

The two July flood events monitored in 1986 and 1987 not only showed differences in glacio-hydrological characteristics (Chapter 8) but also in channel change characteristics (Table 6.4). Mean values of channel change variables suggest that the effect of the 1987 event was greater than the 1986 event, in relation to all variables except west bank erosion and the maximum depth of deposition in the valley train (Table 6.4). The standard errors associated with the various attributes were proportionately much greater in 1986. The ratio of thalweg shift (lateral channel migration) and thalweg lowering gives a good indication of the type of flood event. The 1986 event produced a shift:lowering ratio of 51:1 whilst in the 1987 event yielded 2.6:1. It is tempting to cite this as the dominant control on bank erosion, since bank erosion occurred on both margins of the valley train and deposition was much greater in 1986, whereas in 1987 there was a marked asymmetry to bank erosion and there was little deposition in the valley train. The 1987 flood was characterised by lowering of the thalweg, much higher rates of sediment loss and much less deposition (Table 6.4).

The detail of the two channel response series (Figure 6.15 and Figure 6.16) show major contrasts in downstream trends in thalweg and channel sediment dynamics between 1986 and 1987. All distances on these graphs are measured downstream of the upper cross section which is 26.5 m from the glacier snout. Great variability between sections is characteristic of the 1986 data,

Table 6.4 Contrasts in mean values of channel change variables for the 1986 and 1987 flood events

VARIABLE	UNITS	1986	s.e. +/-	1987	s.e. +/-	CHANGE
Thalweg lowering (a)	m	0.01	0.7	0.65	0.62	+
Thalweg shift (b)	m	0.51	5.9	1.72	6.5	+
Thalweg ratio a:b		51.0		2.6		
East bank erosion	m	1.05	4.9	1.61	2.5	+
West bank erosion	m	0.85	5.9	0.33	1.8	-
Depth of erosion	m	1.07	1.1	1.52	0.7	+
Depth of deposition	m	0.84	0.6	0.39	0.4	-
Gross sediment loss	m^2	9.64	12.9	13.03	9.3	+
Gross sediment gain	m^2	5.07	7.0	2.86	4.1	+
Net sediment change	m^2	-4.57	16.1	-9.79	11.41	+

NOTES: Change relates to change in variable 1986 -1987
+ = increase and - = decrease

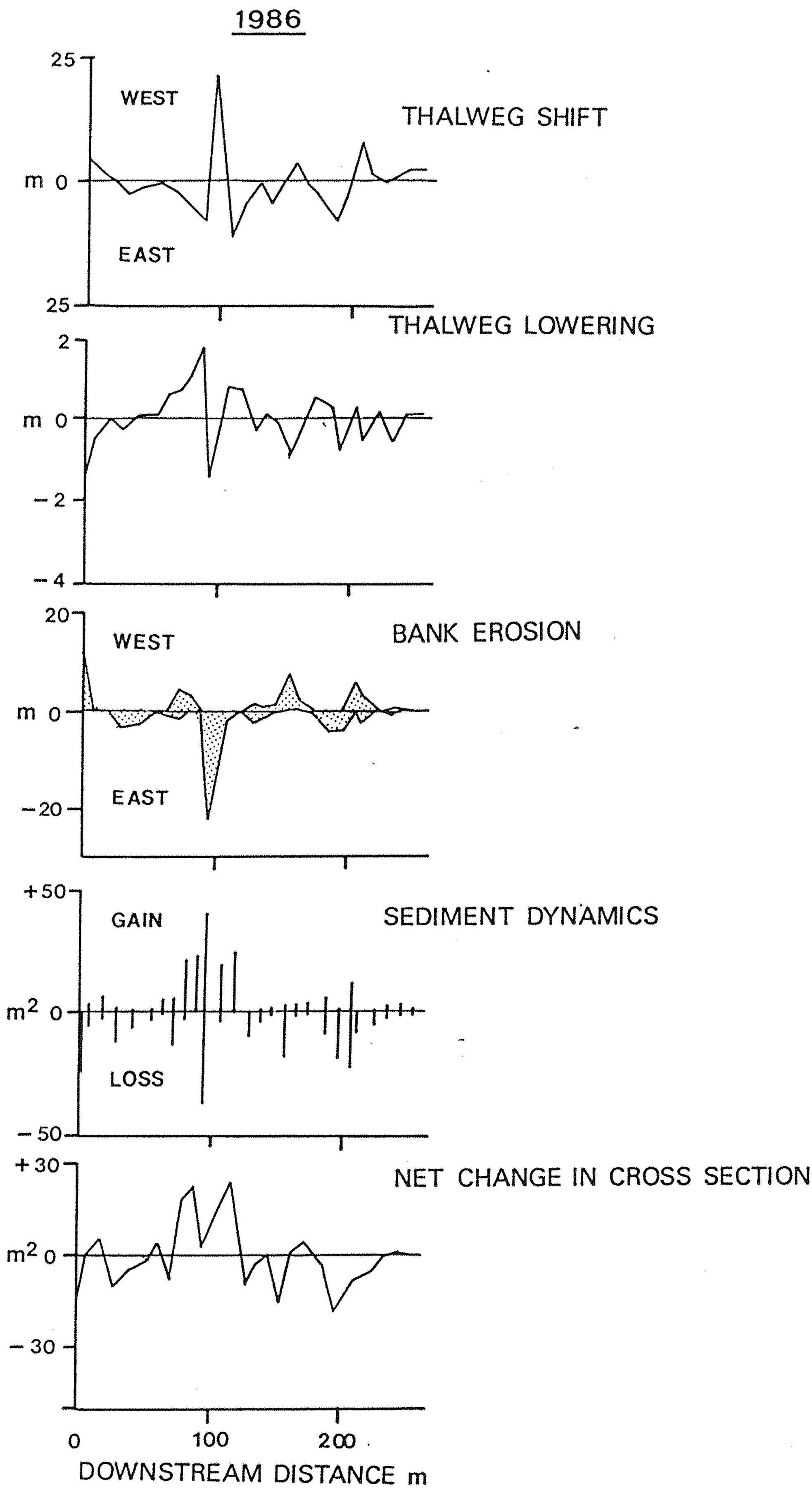


Figure 6.15 Downstream change in valley train cross-section characteristics as a result of the July 1986 flood. All downstream distances are measured from the upper cross-section which is 26.5m from the glacier snout. Sediment dynamics refers to loss or gain (deposition) of sediment in each survey cross-section).

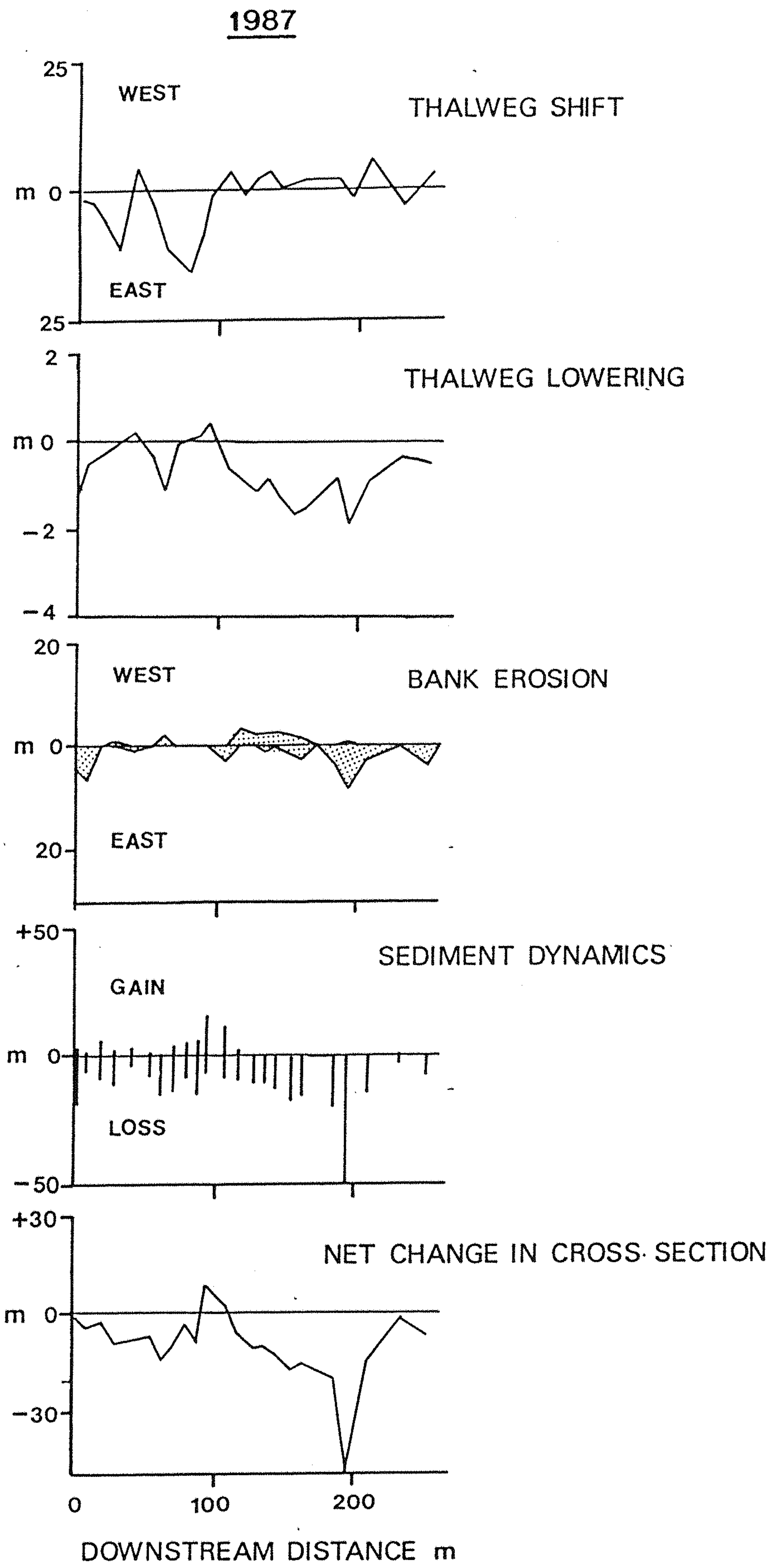


Figure 6.16 Downstream change in valley train cross-section characteristics as a result of the July 1987 flood. All downstream distances are measured from the upper cross-section which is 26.5m from the glacier snout. Sediment dynamics refers to loss or gain (deposition) of sediment in each survey cross-section)

Figure 6.14 shows marked differences between pre and post-flood channel morphometries. The 'tightness' of the inter-relationships is greater following the flood and may represent a better adjusted channel morphology (or fresher preservation of channel features in the landscape). Adjustment to fluvial activity would be manifest in a greater number of inter-relationships between variables, especially hydraulic radius. Before the flood, variables are largely unconnected. Depth, wetted perimeter and cross sectional area are the best related with width loosely related to wetted perimeter and cross sectional area linked to hydraulic radius. No significant correlation exists between the number of channels and any other variable (before the flood the channel pattern was predominantly single thread with very little width variation). The post flood system is markedly different since there are many more linkages between variables, particularly between width and the number of channels. Depth is unrelated to any other variable. This change reflects the increased braiding following the flood, which increases the importance of channel width and the number of channels. The difference between the two systems is striking (Figure 6.14). Only two variables are commonly linked in the two systems, width and wetted perimeter (this is not surprising). What is also remarkable is that no inter-relationships extend between the two systems. The frequency of linkages between the two systems would give an indication of a persistence structure to the fluvial morphology of the valley train. Since no linkages occur, the two systems are unrelated structurally (i.e. there is no systematic pattern to channel morphology before and after flood).

The correlation systems (Figure 6.14) of pre and post flood morphological variables suggest a transition between a depth dominated system with few morphological inter-relationships to a well-linked width dominated whilst the 1987 data show a a slight downstream trend (Figure 6.15 and Figure 6.16). The main feature of the 1986 series was the perturbation in the magnitude of variables at 80 to 90 m downstream from the upper section. This was related to a rapid reduction in stream power at this point (Figure 6.13). Figure 6.15 shows that the thalweg is elevated between 60 to 80 m downstream of the upper cross section and was accompanied by a rapid shift in thalweg position, probably producing braiding. Rapid lateral migration not only caused high rates of deposition but also increased erosion on the eastern side of the valley train. Sediment dynamics not only responded to changes in channel behaviour, as reflected in the thalweg changes, but also to the gross geomorphology of the valley train (e.g. the greatest changes in net sediment storage occurred in the widest part of the valley train).

In 1987 the situation was very different (Figure 6.16). Apart from early oscillation in the thalweg, the channel changed position very little but instead incised into the valley train. Maximum bank erosion rates in 1986 corresponded to the areas of maximum thalweg shift. In 1987, maximum bank erosion was in the areas where thalweg lowering occurred (e.g. 180 to 200 m in areas where the valley train was relatively confined). The likely cause of bank retreat was, probably related to bank under-cutting. In this confined section there was considerable sediment loss and very little deposition downstream of 140 m from the upper cross section. The same area of maximum deposition occurred in the central channel reach (90 to 120 m) as in 1986. Erosion immediately above the meltwater intake was rapidly curtailed as channel slope was greatly reduced. This could have been related to sedimentation in the gravel trap.

As demonstrated by the standard errors of the measured variables (Table 6.4) and the patterns shown in Figures 6.15 and 6.16, variations over the reach are great testifying to the dynamism of the channel during major floods.

Adopting the same morphological systems correlation approach used in the evaluation of channel geometry systems (Section 6.5.1) it is possible to compare the characteristics of channel adjustment during the 1986 and 1987 floods (Table 6.5, Figure 6.17). The main difference between the two floods was the degree of linkage of the thalweg variables (Figure 6.17). In 1986 both thalweg shift and lowering were tightly linked to channel change variables. In 1987 this was not the case and Thalweg shift bore no significant relationship with any other variable. Concentrating on the different linkages between the two years (Figure 6.17), improved links (from 1986 to 1987) existed between relationships emphasising vertical channel movements in 1986 than in 1987. This suggests greater incision, deposition, abandonment of valley train segments and bank erosion in 1986. West margin bank erosion was also incorporated in 1986 but with negative links suggesting poor correspondence between the depth of erosion and deposition on this side of the valley train. Links that existed in 1986 but not in 1987 (Figure 6.17) mainly related thalweg shift to depth of erosion and sediment loss. Care should be taken in interpreting these systems since correlations are not causal. Coding of values in the analysis is particularly important since this affects the direction of correlation (e.g. if thalweg lowering were to be expressed as negative then the correlation with depth of erosion would be negative).

These analyses suggest that the two events were markedly different in terms of channel adjustments. An *a posteriori* assessment of the morphologic inter relationships usefully identifies thalweg as being the main controlling variable. The dominance of thalweg

Table 6.5 Correlation matrices of channel change variables 1986 and 1987.

1986

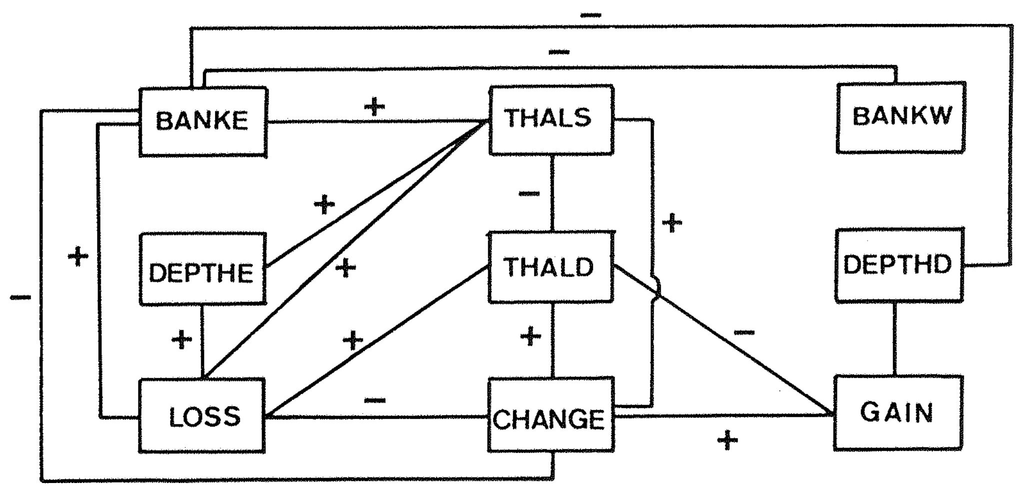
	BANKE	BANKW	THALD	THALS	LOSS	GAIN	CHANGE	DEPTHE
BANKW	-0.594							
THALD	-0.326	-0.144						
THALS	0.478	-0.133	-0.691					
LOSS	0.701	-0.298	0.556	0.677				
GAIN	-0.326	-0.065	-0.619	-0.344	-0.233			
CHANGE	-0.706	0.212	0.716	0.695	0.907	0.620		
DEPTHE	0.170	0.305	0.344	0.431	0.650	-0.070	-0.554	
DEPTHD	-0.408	0.168	-0.321	-0.325	-0.135	0.546	0.344	0.086

1987

	BANKE	BANKW	THALD	THALS	LOSS	GAIN	CHANGE	DEPTHE
BANKW	-0.237							
THALD	-0.521	-0.291						
THALS	0.250	-0.082	-0.231					
LOSS	0.588	0.120	0.619	0.052				
GAIN	-0.340	-0.510	-0.586	-0.078	-0.256			
CHANGE	-0.529	-0.287	0.722	0.074	0.931	0.556		
DEPTHE	0.415	0.250	0.612	0.073	0.914	-0.274	-0.867	
DEPTHD	-0.355	-0.430	-0.618	-0.128	-0.358	0.665	0.520	-0.283

NOTE: Underlined values significant at 0.05 (2)

1986



1987

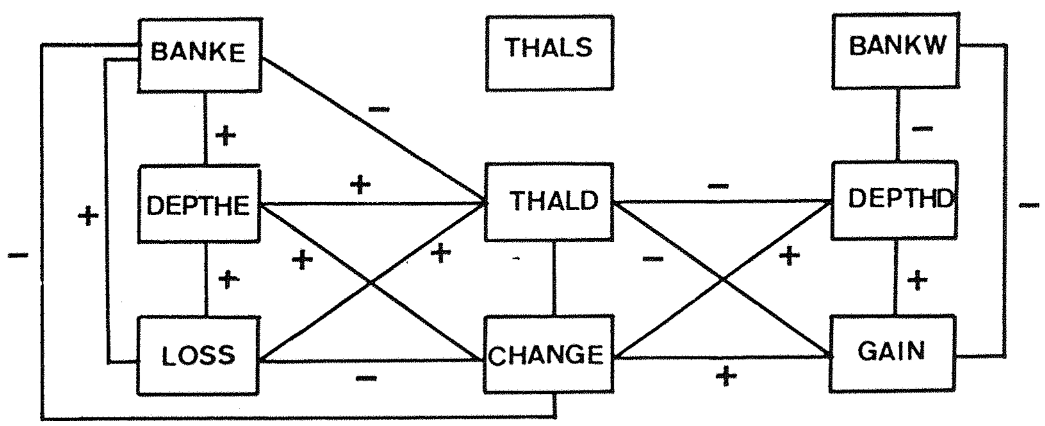


Figure 6.17 1986 and 1987 channel adjustments based on the correlation systems for the two years. Correlations are based on Table 6.5.

shift in 1986 compared to lowering in 1987 is responsible for the different patterns of erosion.

6.6 Sediment output from channel erosion

The implications of different modes of channel change for sediment output from the valley train can be assessed by calculating the amount of sediment evacuated from the proglacial zone during the July 1986 and 1987 floods (Table 6.6). Using data from the 28 monumented cross sections, the valley train can be divided into 28 'valley train cells' and the volume of sediment lost and gained can be calculated. Each cell estimate (m^3) can then be multiplied by a volume - weight conversion factor (1.6 tonnes per m^3 based on bulk density samples of floodplain gravels) to give an estimate of sediment yield. Computations for gross sediment loss, gross sediment gain and net yield can be calculated for individual cells and for the whole reach.

Clear contrasts exist between the two events (Table 6.6). The 1987 flood resulted in erosion of more than twice the weight of sediment of the 1986 event in both gross and net terms. Gross deposition was about the same for the two events, but the ratio of erosion to deposition for 1987 was twice the 1986 value emphasising the erosive nature of the event (i.e. thalweg lowering as opposed to thalweg shift). Calculations of sediment output are accurate in the area enclosed by the upper and lower cross sections but beyond these limits, as for example in the 26.5 metres from the top section to the ice margin, extrapolation is required. Using the sediment balance characteristics measured in the top cross section and extrapolating these values over the reach to the glacier snout provides an adjusted estimate of sediment output. These estimates (Table 6.6) should be added to the reach totals to get a full picture of channel erosion and deposition over the full length of

Table 6.6 Sediment Loss due to Channel Change Bas Arolla valley train 1986 and 1987 [D]

YEAR	NOTE	GROSS EROSION tonnes	GROSS DEPOSITION tonnes	NET BALANCE tonnes	EROSION: DEPOSITION ratio
1986	A	2456	1326	1130	1.9
	B	956	101	845	
	C	3402	1427	1975	
1987	A	5294	1176	4188	4.5
	B	831	157	674	
	C	6125	1333	4794	

NOTES:
A = Calculations for area within monumented sections.
B = Estimated sediment conribution form reach above top section to base of glacier
C = A + B
D Estimates only apply for the duration of the 1986 and 1987 field seasons.

the proglacial zone. Adjusted net outputs were 1975 tonnes for the 1986 event and 4794 tonnes for the 1987 event.

Estimating storage or residence times for sediment in the proglacial zone cannot be easily accomplished because no realistic limits can be determined for the valley train sediment store. However, assuming that the depth of the sediment store approximates the greatest observed vertical depth of erosion by the thalweg in 1986 and 1987 (2.45 - 3.8 m), then a crude store volume can be calculated given that the planimetric area of the proglacial zone (7545 m^2):

<u>Volume in storage in the valley train</u>

(A) maximum depth of lowering = 2.45 to 3.8 m (1986, 1987 values)

(B) area of proglacial zone (1987) = 7545 m^2

(C) weight : volume = 1.6 tonnes / m^3

(determined from valley train bulk density measurements)

(D) A x B x C = 18485 to 45874 tonnes

<u>Reworking time for stored material</u>

(E) range in estimated flood event yields (from Table 6.6)

= 1975 to 4794 tonnes

D / E = 3.86 to 23.23 years

This assumes floods occur on an annual basis, which may not strictly be the case (for example there were two floods in 1987). These calculations also assume a fixed storage volume but in the proglacial context this is misleading since the store is in a constant state of flux in all dimensions (from erosion of valley train bluffs, deposition of sediments from valley side slopes, and fluctuations of the ice margin) and the vertical extent of the store is undetermined. Even so this crude calculation suggests very active degradation of the proglacial zone which is in contrast to the belief that proglacial zones are aggradational (Smith, 1974; Maizels, 1983). However the Bas Arolla channel is an extreme proximal proglacial reach and is more characteristic of a fan head trench (Schumm et al., 1987). With greater distance away from the ice margin, the system would probably become aggradational.

Because no long-term observations of long profile changes in the Bas Arolla proglacial zone are available it seems unlikely that these rates of degradation could continue for a long period. It may be that the 1986 and 1987 flood events were unusually erosive. Fenn and Gurnell (1987) present data from the Bas Arolla proglacial zone which indicates over 8 m of lowering of the channel bed at the snout in one year (July 1977 - July 1978)! This value is far greater than the values shown in Figure 6.18 and stresses the great variation that can occur. Detailed survey of long profile of the Bas Arolla proglacial zone during the study period suggests that between June 1986 and May 1987 there is little overall change in the profiles, although following the 1987 flood event considerable scour in the lower reaches was evident (Figure 6.18). This fully supports the evidence from the monumented cross-sections (Figure 6.15 and Figure 6.16) with regard to the erosive nature of the two events. In the light of these surveys where maximum deviation between profiles is approximately 2.1 m, the estimate by Fenn and Gurnell (1987) seems a little high.

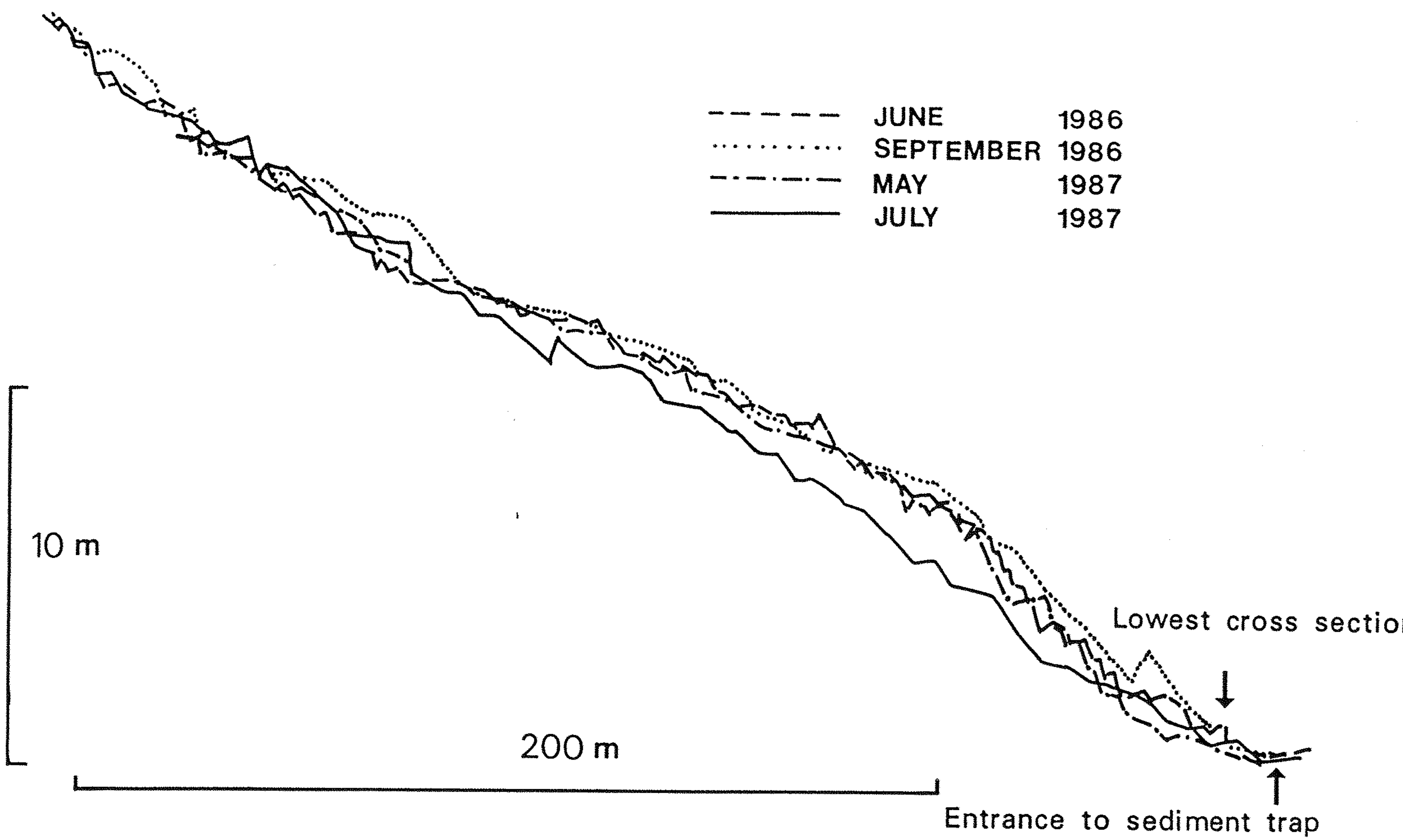


Figure 6.18 Long profile surveys of the Bas Arolla proglacial stream, June 1986 to September 1987.

6.7 Discussion and summary

Assessing the importance of gradual versus catastrophic change is not normally possible given the characteristic short-term length of geomorphic study. However, in proglacial mountain streams where high magnitude discharges are assured these effects can be observed, often within the same ablation period. Short-term changes in channel cross section prior to flood have been shown in this chapter to be generally an order of magnitude less than during flood flows (Figure 6.5). The direction of response is also interesting since the channel can shift dramatically from a slightly aggrading cross section to a state of massive aggradation or degradation.

The general channel response to flood flows in the Bas Arolla stream involved a switching of habit from a single thread, low sinuousity stream to a wider channel with 'bulbs' or 'nodes' of braiding. Following flood flows channel pattern rationalised by a reduction in the extent of braiding so that single thread channel was re-established by the end of the ablation season. This cyclical behaviour of the channel has been reported previously in proglacial streams (Fahnestock, 1963) and is common in fluvial systems responding to flood events (e.g. Ferguson and Werrity, 1983; Werrity, 1984). However, this is a simplification of the true nature of change because inevitably more complex patterns exist. Compare the 1986 and 1987 flood events (Figure 6.10 and 6.11), the 1986 event was characterised by a laterally shifting thalweg and relatively low sediment yield (1975 tonnes) from channel sources whilst the 1987 event was associated with thalweg scour and much higher channel sediment yields (4792 tonnes). The actual reason for this difference is difficult to determine, certainly the 1987 event was more prolonged but peak discharge was equivalent to the first event. This may explain the greater sediment yield from the second event and the fact that later in the 1987 ablation season sediment load was diminished and the stream channel switched from a laterally shifting to a scouring thalweg. Alternatively the effect of the flood may depend on the initial state of the fluvial system before flood flows and differences in response may relate to intrinsic threshold controls (Schumm, 1973).

Flood events are often assumed to leave a stable imprint on channel form (Gupta, 1983). However, the imprint of floods may be less marked than was initially appreciated. For example braiding in the central channel reach during the July 1987 flood seems to be an important feature (Figure 6.11) but examining the cross sections for this reach (54 to 79 m downstream from the upper cross section) in greater detail shows that there was a dominant thread of flow with only very shallow subsidiary channels (Figure 6.19). This is in contrast to more conventional braided stream cross sections where the flow is divided into numerous channels separated by bars or islands where the cross section is wide, flat-bottomed and shallow (Schumm et al., 1987, p. 155). The dominance of a single thalweg thread in the Bas Arolla cross sections is good evidence that the pattern of braiding is fickle. This was demonstrated between July 19th and 20th 1987 when the number of braid channels was reduced from 5 to 2. In 1986 this pattern was also evident but because the event was characterised by a laterally shifting thalweg, braid patterns tended to be more stable. Therefore pattern alone does not define the regime of the stream.

Spatial variations in channel adjustment are considerable during flood (Figure 6.15 and Figure 6.16). This is explicable in terms of the rapid spatial variations in material and geometric characteristics of the proglacial zone, since these properties will affect the hydraulics of flow and therefore the hydraulic

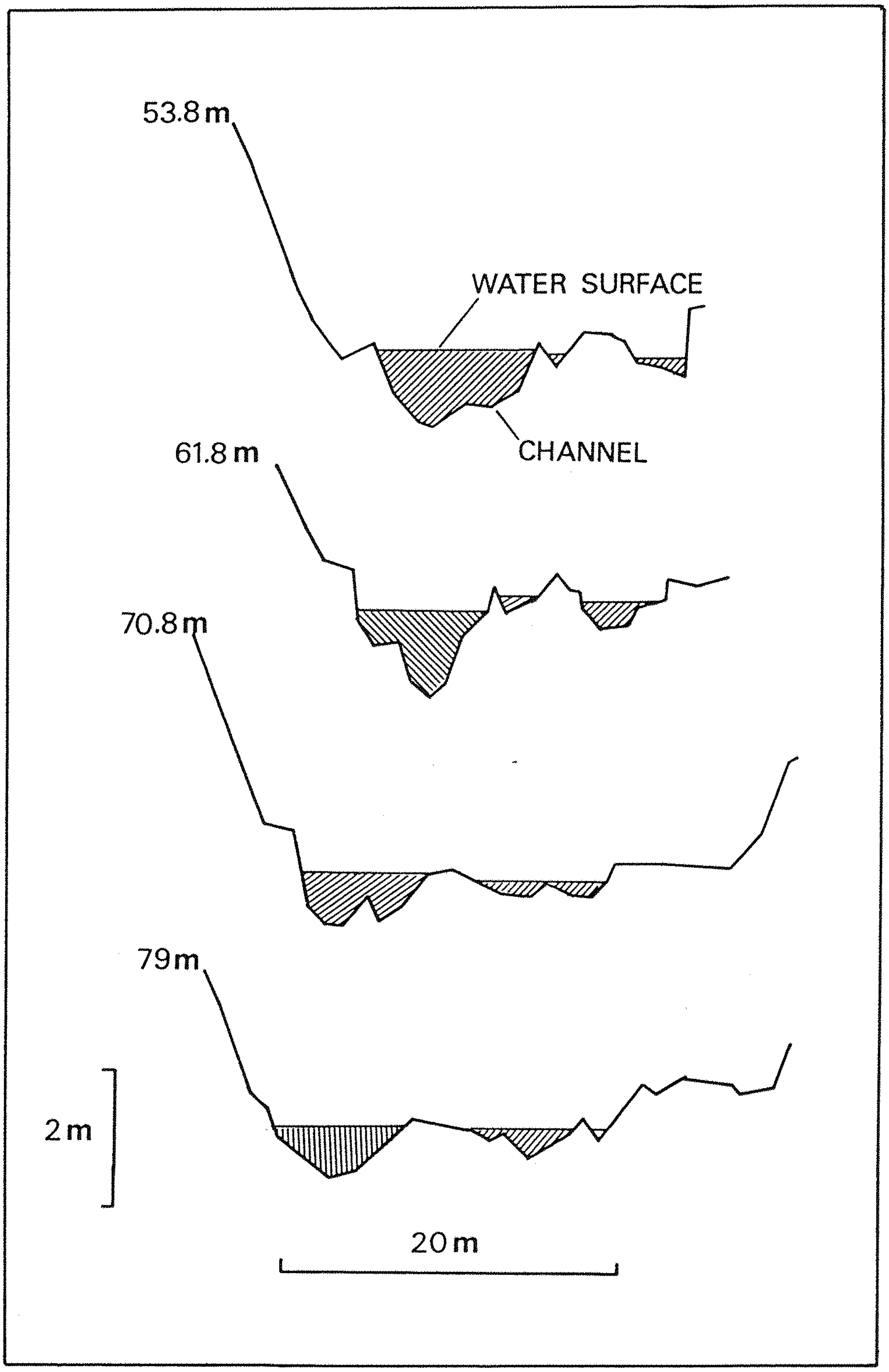


Figure 6.19 Valley train cross-sections following the July 1987 floods having the braided nature of the channel in the middle reach (53.8-79m downstream of the upper section) of the Bas Arolla Proglacial zone.

geometry of the channel (Ferguson, 1986b). Given variable bed material characteristics from section to section, the channel routes the discharge or the discharge forms the channel (Richards, 1977). Spatial variations in sediment transport should therefore be considered the norm in this type of channel (Pitlick and Thorne, 1987). The coarseness of channel materials in mountain streams largely prevents the establishment of significant hydraulic geometry relationships (Wasserman, 1988).

Observed sediment output from the proglacial zone reflected the characteristics of the model suggested by Trimble (1988). There are episodic outputs of sediment (Figure 6.5); highly variable rates of sediment gain and loss (Figure 6.15 and Figure 6.16); and the last large major flood has a large effect on the sediment budget in these mountain streams (Table 6.6). These effects are clearly illustrated by the two floods in July 1986 and 1987. The situation would be complicated by slope inputs into the sediment system since the linking between hillslope and channel systems has a profound effect on channel response (Nolan and Marron, 1985). However this is a very simple model and within any reach separate zones of sediment transport and storage will be present.

Chapter 7.

SEDIMENT CONTRIBUTION FROM VALLEY TRAIN BLUFFS

7.1. Introduction

Bank erosion has been the focus of a great deal of geomorphic research in recent years (Thorne, 1982) and is an important sediment-related problem which needs to be addressed by river engineers (White, 1987). Wolman (1959) initiated process orientated bank erosion studies and since then the mechanisms of bank failure have been explained (Thorne and Tovey, 1981); rates of bank retreat have been measured (Hooke, 1980); and contributions to fluvial sediment loads have been estimated (Kirkby, 1967). With the exception of the study of Arctic bank processes (Lawson, 1983) the majority of this research has been in temperate climates. Streambank erosion in mountain and proglacial streams has not been examined in detail.

Proglacial streams banks are likely to be susceptible to bank erosion because banks are generally composed of largely unconsolidated glacial and fluvio-glacial deposits. Vegetation cover is sparse and channels respond rapidly to changing sediment and water discharge conditions (Fenn and Gurnell, 1987). In proglacial streams Fahnestock (1963), Hammer and Smith (1981) and Gurnell (1982) all suggest bank erosion as an important sediment source but this was not measured directly. Part of the reason for this is that river bank erosion is usually associated with high discharge (flood) events and as a consequence is usually not observed and very rarely quantified. This is unfortunate because bank erosion is important in terms of channel change, valley train development, supply of sediment to the streamload (Thorne, 1981) and the generation of suspended sediment pulses (spikes) (Hammer and Smith, 1983).

In alpine areas proglacial stream channels are usually laterally confined to a narrow valley train which is actively reworked by fluvial processes (Fahnestock, 1963). Under these circumstances there is a large potential for sediment transport because steep bluffs along the margin of the valley train can be undercut, albeit sporadically, at high flows. During lesser flows the proglacial channel is confined within much smaller, quasi-stable banks composed of fluvially worked material. These banks are an ephemeral geomorphic feature since they owe their origin to the changing morphology of the channel. The large bluffs are generally 'cut' during flood events, are then abandoned, and remain uncoupled from the flow until the next large event, unless the channel remains 'fixed' at the bluff base. This distinction between the two bank types is important since it defines a two step morphology to the valley train cross-section (Figure 7.1) with the large bluffs located at the margin of the valley train and smaller streambanks adjacent to the channel. In this section bank erosion is considered only in relation to valley margin processes (the bluffs). Channel bank adjustment was discussed in relation to channel adjustments in Chapter 6.

This Chapter describes bluff erosion in the proglacial zone of the Bas Glacier d'Arolla in 1986 and 1987. In particular bluff erosion is considered in the context of sediment supply to the stream channel (Section 7.4) and in the production of non-periodic sediment pulses (Section 7.5) which are characteristic of many proglacial streams (Fenn, 1983).

7.2 Form and Composition of valley margin bluffs

The importance of bank materials is that they control

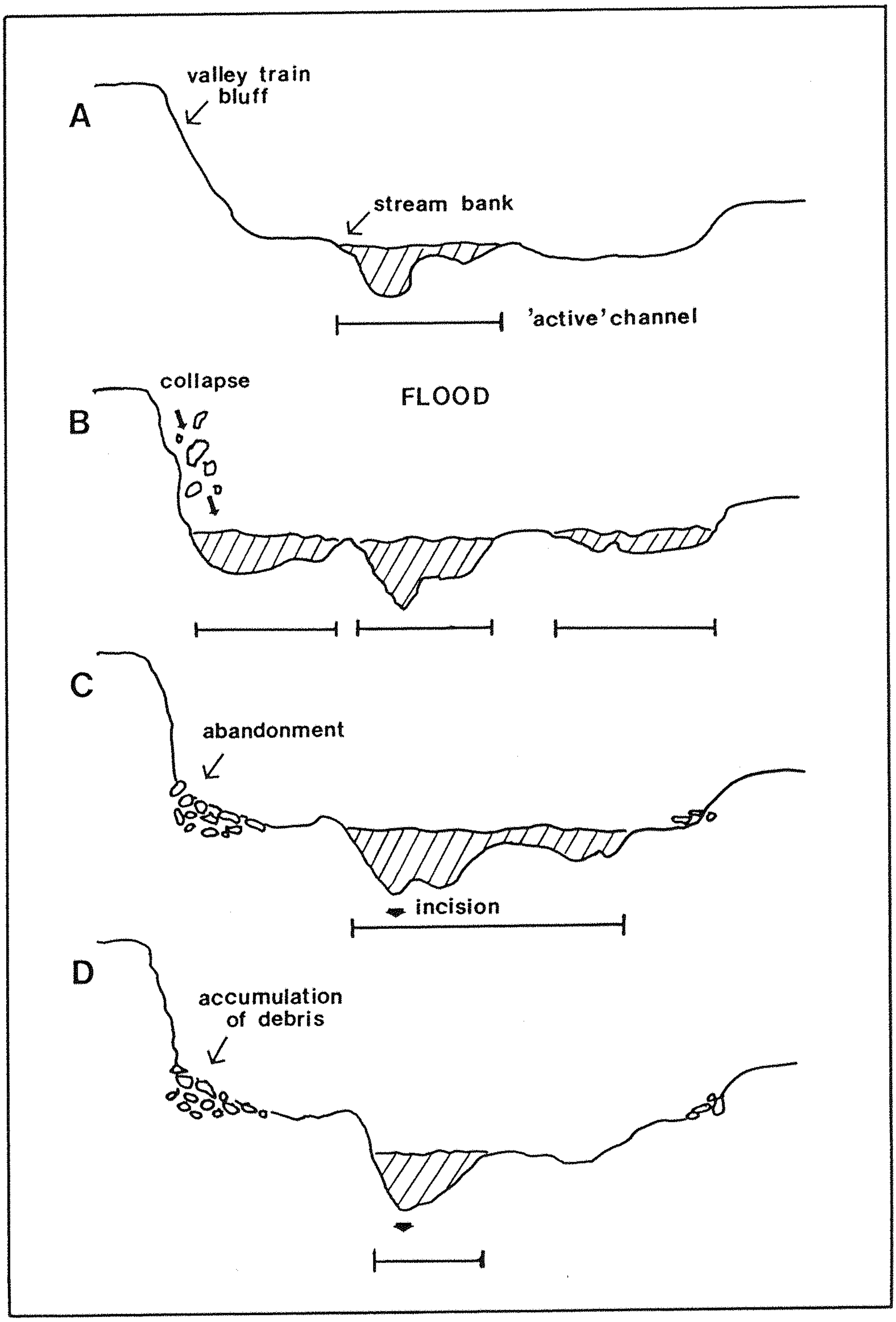
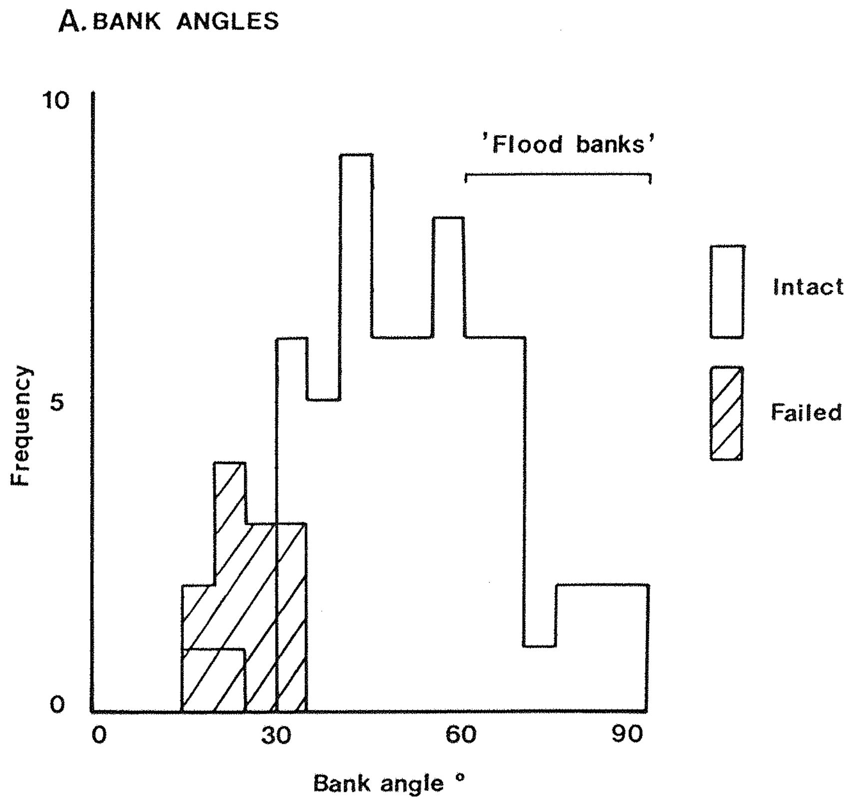


Figure 7.1 Relationships between valley train bluffs and channel streambanks in the Bas Glacier d'Arolla proglacial zone.

bank stability which influences channel cross-sectional shape and affects rates of channel migration. However, because of the wide range of particle sizes present in valley bluffs, only the smaller size distributions are used to approximate the bulk properties of the material (Whalley, 1975). This can be justified to some extent since the particle size of the matrix greatly influences bulk bank stability (Schumm, 1960; Goss, 1973) and the supply of fine sediment to the streamload. Difficulties in carrying out material testing on gravel and boulder materials makes bank stability analyses difficult. However, Whalley (1975) working in the same area as this study, on the origin of abnormally steep moraines, carried out a simple Culman stability analysis on very similar materials. Results suggest a theoretical moraine height far less than the heights observed. This lead Whalley to suggest that these slopes were abnormally steep, and had an internal cohesion which allowed them to attain steep angles of upto 70^{o} (which is about twice the angle of rest of the material (40^{o})). These angles are equivalent to the bank angles shown in Figure 7.2a for intact (recently cut flood banks) and failed bank faces.

Although there is considerable grain-size variation in their composition, bluffs are generally coarse with few fines (generally 9 % less than 4 phi, Figure 7.2b and Figure 7.3). Therefore the apparent cohesion of bluff faces is unlikely to be related to the presence of clay-sized material. Whalley (1975) suggested that the cohesive nature of the material was related to quartz cementation or the presence of packing which would also explain the sensitive nature of the materials.

A summary of the general bank characteristics of bluffs in the Bas Arolla proglacial zone (1986) is given in Table 7.1. The majority of banks seem to be free from vegetation or partially vegetated. A large proportion of banks are undercut by fluvial action. Bank materials



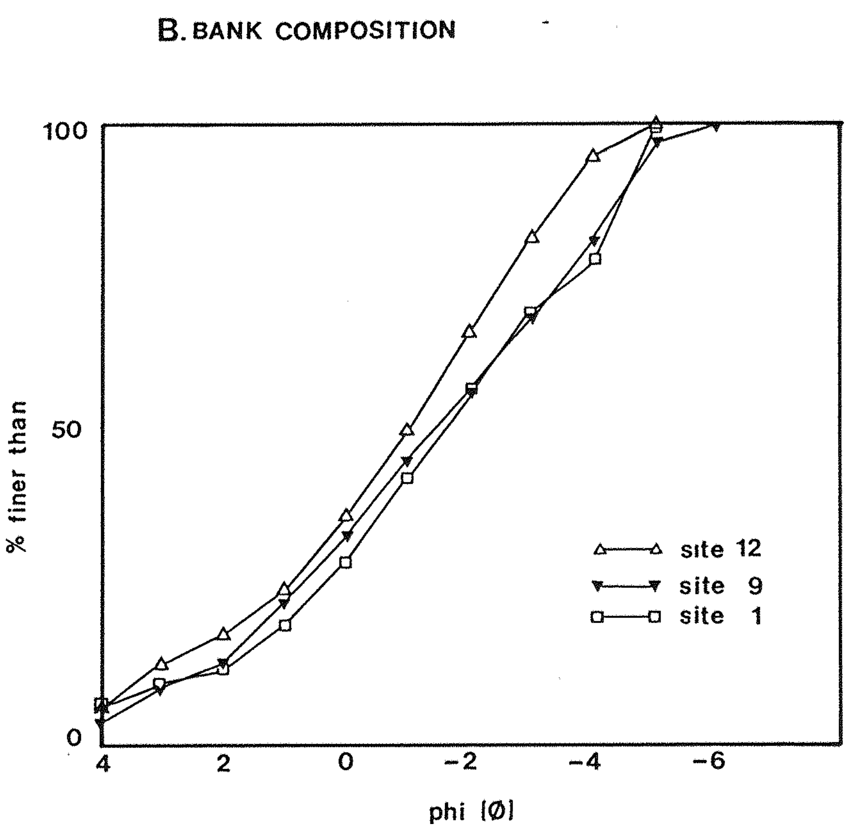


Figure 7.2 Characteristics of Bas Arolla proglacial valley train bluff. a) Bank angles b) matrix grain size comparison.

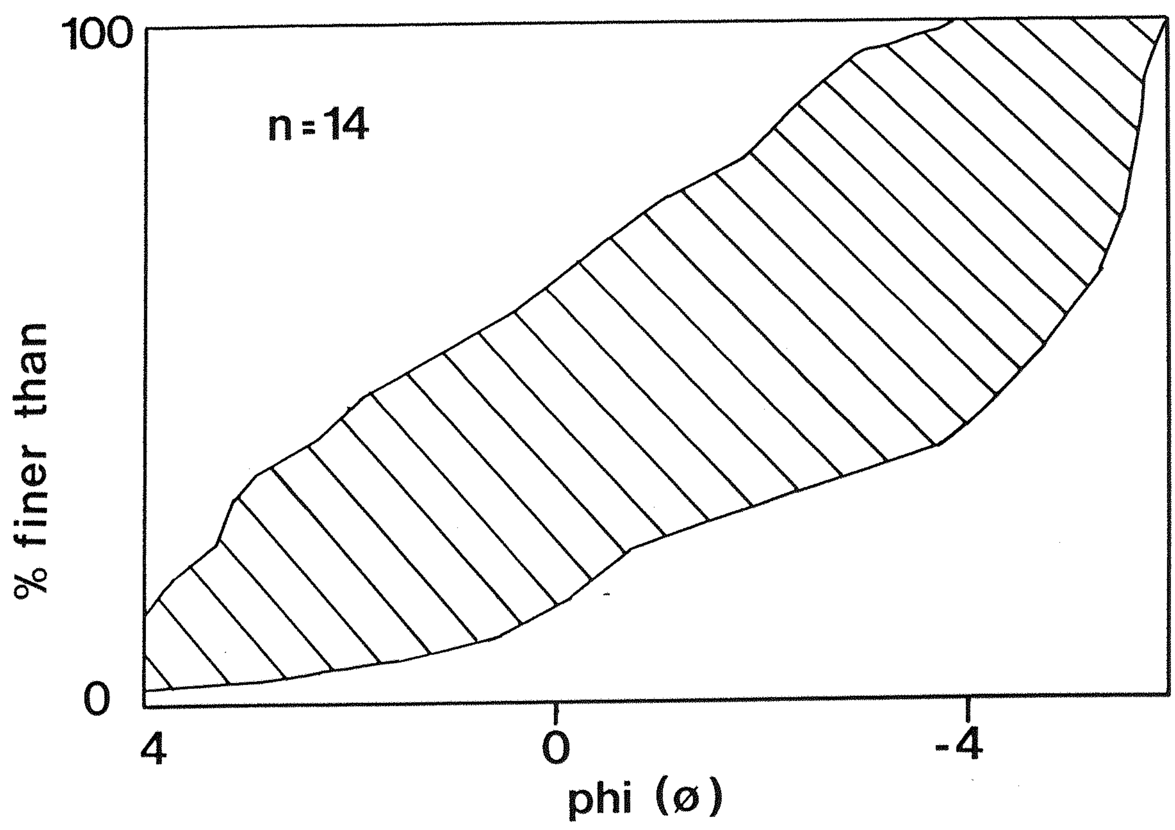


Figure 7.3 Envelope of grain size distributions of matrix samples from the 14 bluff sites studied in the 1986 ablation season.

Table 7.1 Bank Characteristics - 1986 Bas Arolla

	Vegetation free		Bank top	Totally vegetated	
<u>Vegetation</u> n = 62	(49) 79%		(6) 9.7%	(7) 11.3%	
	Freshly cut		Undercut	Not undercut	
<u>Under cut by stream</u> n = 62	(14) 23%		(25) 41%	(22) 36%	
	Till	Boulder gravels (colluvial)	Boulder gravels (fluvial)	Gravels (colluvial)	Gravels (fluvial)
<u>Bank materials</u> n = 62	(13) 21%	(10) 16%	(10) 16%	(11) 18%	(18) 29%
<u>Bank heights</u> n = 73	Range = 0.4 - 6.3		Modal class = 0.6 - 1.0 (26%)		

<u>Note</u>:
Bank materials divided into colluvial/fluvial components by virtue of the stratigraphy and fabric at the sampling sites.

are coarse, usually bouldery gravels. Bank heights range from 0.4 to 6.3 metres with the modal height 0.6 - 1 metre. Bank profiles usually consist of a steep upper face with a lower angled toe deposit at the base (Figure 7.4).

7.3 Measurements of bank erosion

Thorne (1981) suggests that bank erosion should be studied using on site measurements if detailed erosion data is to be obtained. This involves the measurement of sequences of change at specific locations. Two methods were used to obtain such measurements: 1) Detailed profile measurement on bluff faces referenced to datum pegs located well-back from the bank top (Figure 7.4). 2) Monumented cross-sections surveyed across the full valley train (locations and cross sections were discussed in Chapter 6).

These measurements were coupled with planform measurement of bank lengths to determine the rate and location of bank erosion in the proglacial reach. Erosion pins (e.g. Hooke, 1979) could not be used because they caused too much disturbance in the sensitive bank materials. Bank profiling is superior to pin measurements since it indicates areas of erosion and deposition over the full bank profile.

A total of 19 sites were surveyed in 1986 and 1987 and erosion rates calculated (Table 7.2). The two sets of data, July to September 1986 and May to July 1987, represent two parts to the melt-season and so define the seasonal sequence of bank change. The 1986 data define the readjustment processes following a large flood and the 1987 data document processes early in the melt-season leading up to the major seasonal flood. The flood of July 6th 1986 provided an excellent opportunity to observe bank collapse, to measure rates of bank

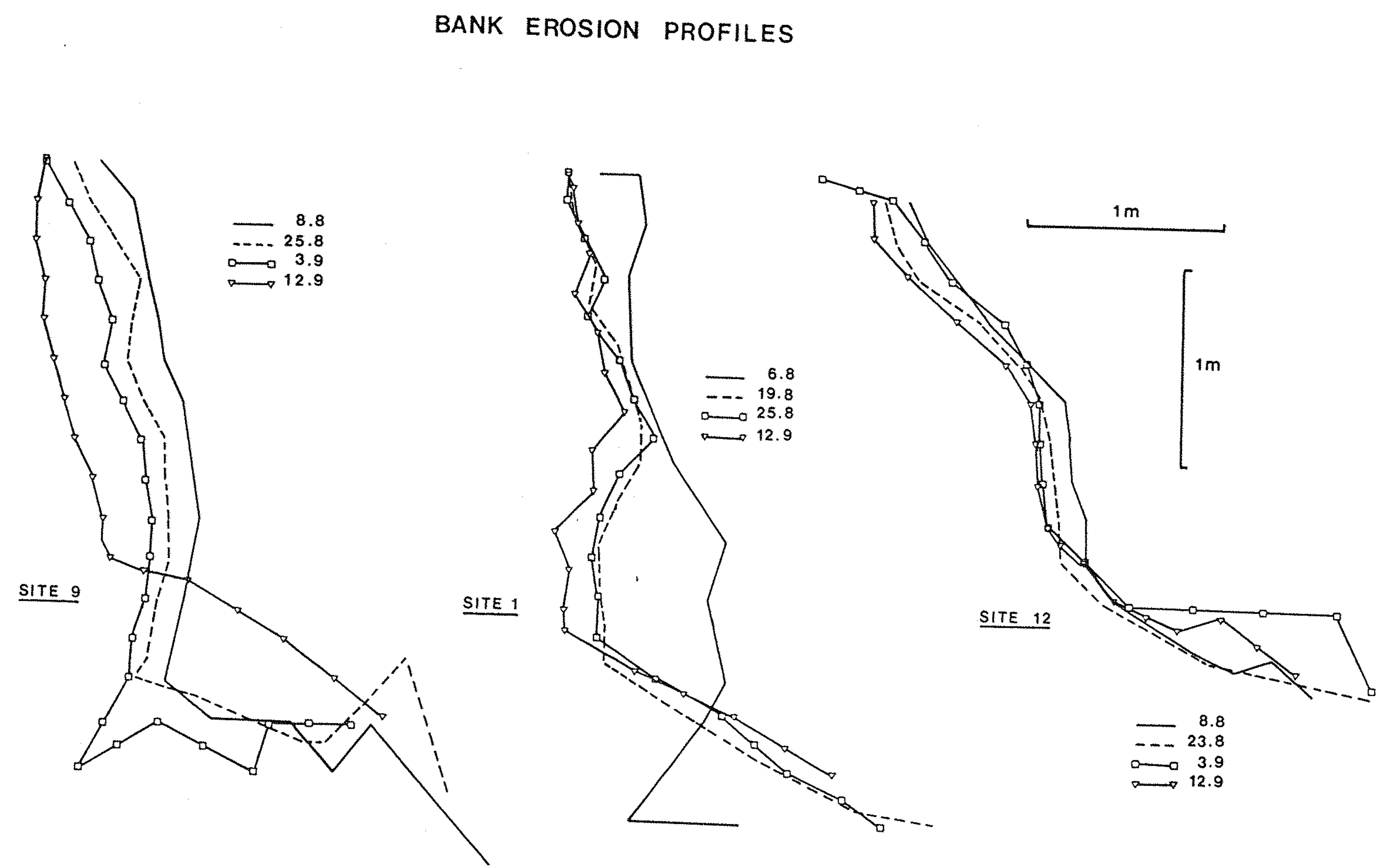


Figure 7.4 Examples of bank profile change at sites 1, 9, and 12 in 1986.

Table 7.2 Summary of bank erosion from surveyed profiles 1986 and 1987.

Site	Dates	Days	Average angle degrees	Height m	Bank material	Vegetated	Erosion m2	Deposition m2	Change m-2	Retreat m	Change m2 d-1	Retreat m d-1
1986												
1	6.8-12.9	37	80	3.3	b/bg	n	1.123	0.291	-0.832	0.84	-0.022	0.023
2	6.8-12.9	37	84	3.6	bg	t	0.210	0.551	+0.341	0.07	+0.009	0.002
	12.9-3.6									0.82		0.003
3	6.8-3.9	28	43	4.5	bg/g	v	0	0	0	0	0	0
4	6.8-3.9	28	35	4.2	bg/g	v	0.166	0	-0.166	0.15	-0.006	0.005
5	6.8-12.9	37	60	2.05	g	t	0.255	0.358	+0.103	0.32	+0.003	0.009
6	11.8-12.9	32	62	1.9	b/bg	n	0.024	0.088	+0.064	0.12	+0.002	0.004
7	11.8-12.9	32	61	1.4	g	n	0.227	0.011	-0.216	0.32	-0.007	0.010
8	8.8-12.9	35	95	3.6	b/bg	t	1.110	0.410	-0.700	0.52	-0.020	0.015
9	8.8-12.9	35	81	3.1	b/bg	t	1.040	0.410	-0.630	0.60	-0.018	0.017
10	8.8-3.9	26	45	5.0	bg	v	0	0	0	0	0	0
12	8.8-12.9	35	59	2.5	bg/g	t	0.343	0.062	-0.281	0.26	-0.008	0.007
13	8.8-12.9	35	60	0.8	g	n	0.076	0.007	-0.069	0.12	-0.002	0.003
14	8.8-12.9	35	50	1.05	g	t	0.154	0	-0.154	0.20	-0.006	0.006
1987												
B	2.6-9.7	37	58	1.9	bg	n	1.030	0	-1.030	0.38	-0.028	0.010
C	3.6-31.7	58	34	4.6	bg	t	2.930	0	-2.930	0.55	-0.051	0.010
D	3.6-31.7	58	44	4.1	bg	t	2.670	0	-2.670	0.56	-0.046	0.010
E	3.6-31.7	58	42	5.0	bg/g	t	1.490	0.740	-0.750	0.58	-0.013	0.010

Notes:
1. Bank materials: g = gravels; bg = boulder gravels; b = boulders
2. Vegetation: t = on top of bank; n = no vegetation; v = vegetated
3. Site 11 (1986) is missing because the survey points were disturbed
4. Site A (1987) is missing because the profile was destroyed by non-bank processes
5. Sites 3 and 10 were stable
6. Site 2 was the only site resurveyed after the winter of 1986/1987
7. Site B was destroyed due to channel migration on July 15th
8. The stream was close to the base of sites C and D in 1987

retreat during flood and to examine the processes of bank readjustment following flood.

On the 6th of July 1986 a small outburst flood from the Bas Arolla glacier caused extensive bank erosion along margins of the proglacial zone. Observations during this flood suggest that vertically cut banks, mainly composed of till, are formed by stream undercutting initiating buttress failure. Figure 7.4 shows 3 bank profiles which, following the flood, showed a progressive retreat of the bank face and accumulation of a basal toe deposit of large boulders. Fluctuations in the stability of the toe were brought about by scour in the main channel, entrainment of clasts from the bank base and settling of base deposits. Adjustment was strongly influenced by the form and composition of the bank and the proximity of the bluffs to the main channel. Table 7.2 illustrates some of these features. The greatest retreat rates were recorded for sites 1,8 and 9, all of which were very steep, boulder gravel banks which were adjacent to the stream channel. Sites 3 and 10 were low angle banks with no stream at the base and as a consequence they were stable. The rest of the banks represent variations between the two extremes. An estimation of at-a-site variation in bank yield was made by surveying the accumulation of toe debris at 6 localities on the same bank face over a period of 15 days (July 16th to 31st, 1987). Accumulation varied between 0.54 and 1.05 m^2 (Standard error 0.19 m^2, coefficient of variation 22%) which suggests that the selection of a single representative profile for erosion along an entire bluff should be undertaken with care.

There are 3 main mechanisms (scales) of failure: mass failure (buttress failure - during flood), large boulder failure, and subaerial erosion (Lawson, 1983). Mass failures occurred during flood conditions as a result of basal under-cutting and caused major adjustments in bluff morphology and channel cross profile. Boulder failures and sub-aerial erosion contributed to re-adjustment of profiles following mass failure. These processes were inter-dependent since sub-aerial erosion 'paves the way' for internal failures (e.g. subaerial weathering of the bank matrix may lead to clast failure). Large boulder failure was the dominant retreat mechanism in bouldery till banks, contributing to the higher retreat rates observed at these sites (Table 7.2).

Comparing rates of retreat between flood and re-adjustment phases (Table 7.3), flood retreat varied between 2 and 6.5 metres and retreat due to readjustment varied between 0.07 and 0.84 m. Comparison between rates at specific sites showed retreat due to flood to to be between 3.8 and 25 times greater than retreat during the readjustment phase. This emphasises the great importance of bluff erosion during flood. Differences in retreat rate are due to the different bank erosion processes that operate during flood and re-adjustment. During flood mass failure of banks is important but following flood clast failures and subaerial weathering dominate. Figures 7.5 and 7.6 show this effect for the July 15th - 17th flood, 1987. Before the flood (Figure 7.5a) sediment had accumulated at the bank base. Following flood (Figure 7.5b) material had been eroded from the basal sediment store and the bank had been cut back. Further bluff erosion continued in the form of small scale shallow slip failures (Figure 7.6c) because the channel remained at the base of the bluff.

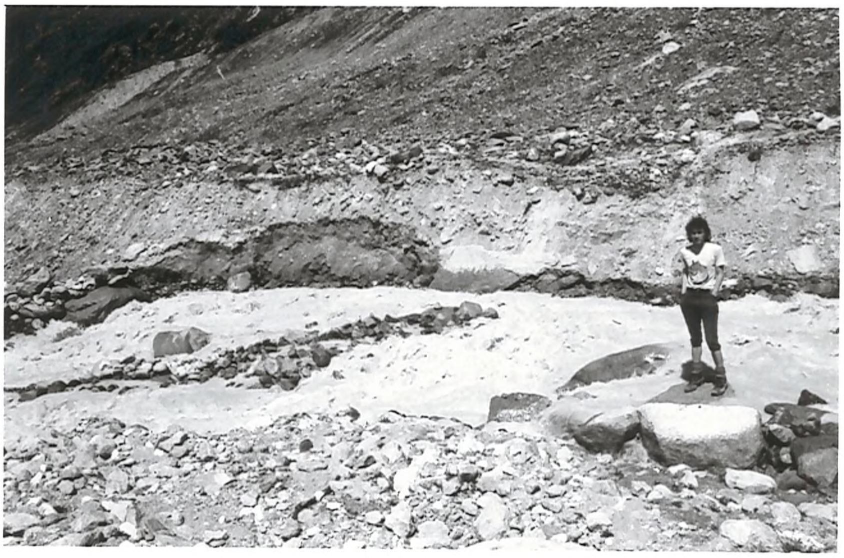
The association between channel movement and bank erosion is illustrated in Figure 7.7 where shifts in the thalweg during the July 6th 1986 flood are associated with zones of bank erosion. This example illustrates that where shifts in thalweg are great they are usually accompanied by bank erosion because within the narrow confines of the valley train flow readily impinges on

Table 7.3 Comparison of bank and bluff recession during the 1986 flood and rates of bank readjustment thereafter. Bas Arolla proglacial zone.

SITE	BANK RECESSION			RECESSION RATIO
	DURING m	FOLLOWING (36 days)	RETREAT m d^{-1}	(flood : adjustmant)
1		0.84	0.024	
2		0.07	0.002	
5	2.50	0.32	0.091	7.8
6	2.50	0.12	0.034	20.8
8	2.00	0.52	0.149	3.8
9	3.50	0.60	0.171	5.8
12	6.50	0.26	0.074	25.0

Figure 7.5 Removal of basal sediment store following the July 1987 flood. The upper photograph (A) shows the bank before flood and the lower photograph (B) the bank following the flood.

Figure 7.6 Sequence of bank change close to site C Bas Arolla proglacial zone (1987). Upper photograph (A) shows the site in mid-June, the middle photograph (B) was taken on July 18th and the lower photograph (C) on July 19th.



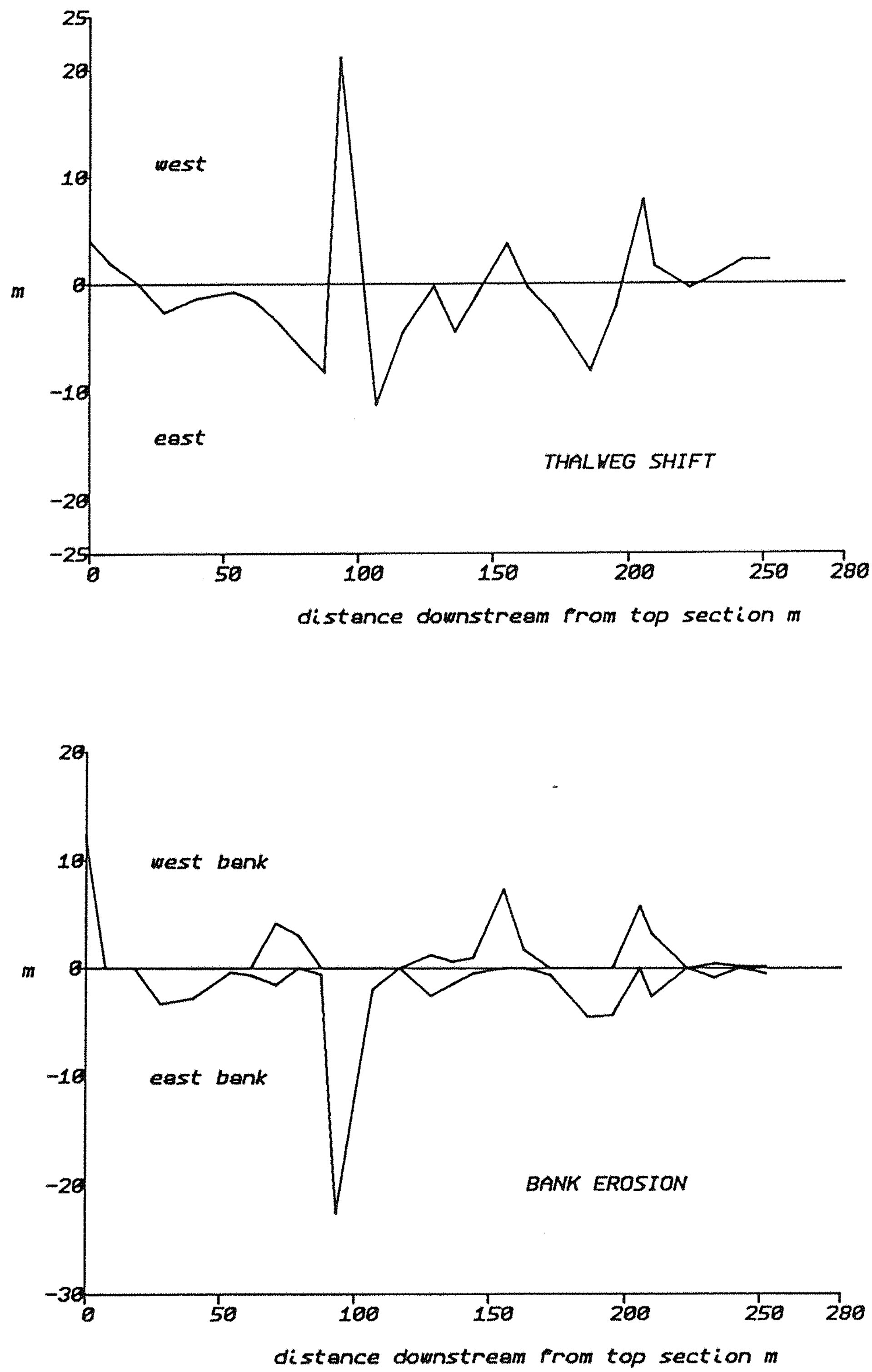


Figure 7.7 Relationship between shift in thalweg of the main proglacial stream during the July 6th 1986 flood and the areas of bank erosion along the valley train bluffs.

the valley margin bluffs. Bluff erosion and channel migration are controlled by the same process of bed material transport (Nanson and Hickin, 1986); which is why large boulders (rip-rap) are used to protect banks from erosion.

In order to evaluate the hypothesis that large boulders at the base of the bluffs provide protection, the movement of six boulders at the base of a bluff were monitored from June 3rd to September 5th 1987. Three of the boulders were lost from the site. These were the three smallest boulders (b axes 59 to 84 cm) and included the two located nearest to the stream channel, which were removed in the flood of 15th to 18th July, and a third from the bank toe, which was lost before the 5th of September, probably during the flood of the 24th of August. Prior to removal some of the boulders showed minor re-adjustments involving movements of a few decimeters. This demonstrates that for the majority of the time large boulders will remain at the bank base but during flood events some boulders (in this case up to 84 cm b axis) can be removed by flood flows. Removal may not necessarily involve entrainment by the flow.

Survey of a large boulder (1.6 x 1.4 x 1.2 m) deposited on the valley train June 8th 1987 (Figure 5.14b) demonstrated that the boulder became overturned and buried within the valley train gravels. From June 8th to July 11th the boulder remained perched on the gravels of the valley train, then in the high flows surrounding the flood event of the 15th to 17th July the eastern margin of the boulder was undercut, a scour hole developed and the boulder toppled in. The net result was transport of the boulder 3.7 m downstream and a lowering of the boulder by 1.54 m so that the boulder was virtually buried in the valley train gravels. Therefore, although the main flow did not transport the block directly, removal of the underlying gravels by scour resulted in movement. This is important in the context of bank protection, since removal of fine material or even small boulders from the bank toe can result in the removal of large blocks and in the exposure of the bank toe to stream action. This balance between supply and removal of material from the bank toe is known as basal endpoint control (Thorne, 1982) and is a key element in understanding bluff erosion processes.

In 1987 bank erosion observations focussed on the early part of the season to try and account for the influence of snow on bank processes. Observations of debris accumulation on plastic sheeting at the bank toe showed that after July 2nd little material was contributed. Prior to this date many clasts (b axes 10-117 mm) accumulated along with a small amount of fines. Material deposited on snow early in the season (Figure 5.12b, Chapter 5) consisted of much larger debris with boulders and soil aggregates up to 60 cm. For the sites studied, early season erosion rates were high although little of the eroded material accumulated at the bank toe (Table 7.2). Part of the reason for this was that sites B, C and D were all located on the channel margin, so that any deposition could enter the stream channel directly. This is misleading because although there was little sediment deposited in 1987 a large basal toe deposit was in existence from the previous year (Figure 7.6). This deposit was protected by the winter snowpack, which transported any sediment deposited during spring thaw away from the bank base towards the stream channel (Figure 5.12b, Chapter 5). Figure 7.8 shows the location of the seasonal snowpack associated with site E. The snow clearly protects the bank toe.

The observations described above suggest a two stage process of sediment yield. Bank processes during one season one season supplied material for removal in the following year. This has been noted by Miles (1976) in arctic and subarctic areas, by Koutaniemi (1984) in Finland and by Bathurst et al (1986a) in the Roaring

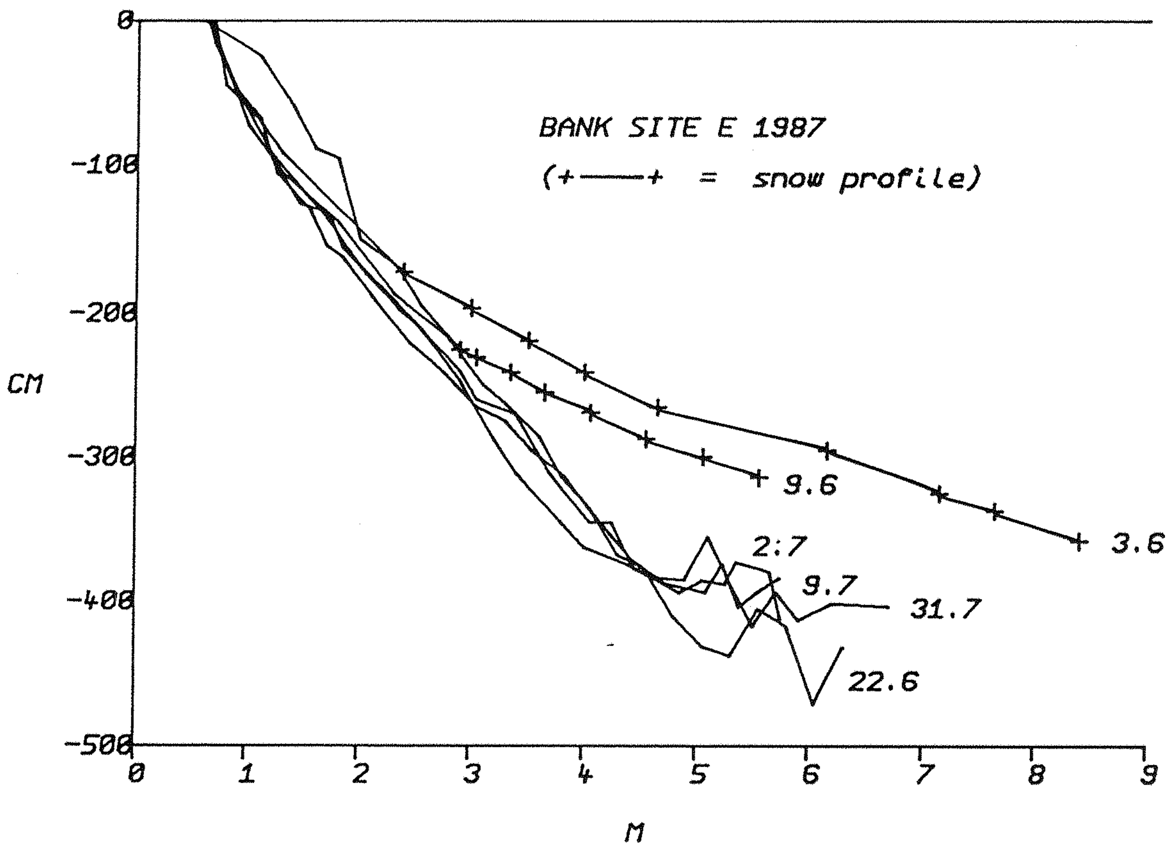


Figure 7.8 Bank profiles at site E, 1987 showing the presence of a basal snowpack covering the bank toe.

River, Colorado and seems widely applicable in streams which have a seasonal discharge regime. Figure 7.6 shows 3 views of the same bank taken in mid-June, 18th July and 19th July. Figure 7.6a shows the large accumulation of debris from 1986. This was removed during the flood of July 15th-18th (Figure 7.6b) and the bank underwent further minor failures in response to other minor fluctuations in discharge (Figure 7.6c). Minor adjustments of the type shown in Figure 7.6c are typical since the proglacial discharge regime is characterised by a summer flow peak and relatively strong diurnal fluctuations often accompanied by pulses of discharge. Fluctuations in discharge are a major cause of bank erosion (Wolman and Bush, 1961; Knighton, 1973) as long as the main channel is in close proximity to the bank base.

7.4 Contribution to the streamload

Bray (1987), in a study of channel change on the North Nashwaakis stream New Brunswick, suggested that 78% of the sediment load associated with a 1973 flood could be derived from channel erosion of a 200 m reach; similarly 93% of total sediment load resulted from erosion of bluffs on the Water of Deugh in S.W. Scotland (Kirkby, 1967);and Duijsings (1986), in his very comprehensive study of streambank erosion in a forested stream, found streambank erosion contributed 53% of the total sediment load. Therefore bank and bluff erosion can be a very important source of sediment.

Bank retreat rates for bluffs along the Bas Arolla proglacial zone (Table 7.2 and Table 7.3) are much greater than estimates of retreat rates from coarse banks in other areas: Hill (1973), Northern Ireland - gives 'bend' erosion rates of 0.03-0.06 m yr^{-1}; Knighton (1973) where 0.02-0.40 m yr^{-1}; Thorne and Lewin (1979), River Severn - 0.5 m yr^{-1}; Hooke (1980

), Devon Rivers (cohesive banks)- 0.08-1.18 m yr^{-1}; and Lawler (1984), South Wales - 0.02-0.31 m yr^{-1}. However, the Bas Arolla retreat rates are comparable to erosion rates on some of the large arctic and subarctic rivers with permafrost banks (e.g. Lawson, 1983, 1-30 m yr^{-1}). In relation to catchment size the Bas Arolla retreat rates are higher than the average (Hooke, 1980).

In order to calculate the amount of sediment yielded from the valley side bluffs, bank lengths were mapped and combined with representative profile measurements to determine the sediment balance for each bank segment (Table 7.4). Bank segment yields were summed to get the total balance for the proglacial zone for the period May 27th to July 31st 1987. Table 7.4 shows that five out of the twelve bank segments were stable or showed no measurable change. Of the remaining 7 segments, 6 showed a negative balance and only one positive. This positive balance site (Table 7.4, bank F) was adjacent to a small fan and received sediment in addition to that derived from bank processes. Results showed a contribution of 560.3 tonnes from the stream banks with 108.1 tonnes deposited at the bluff base. This yielded a balance of 452.2 tonnes removed from the bank toe. Thus by the 31st of July the bank toe area only stored 19.3% of the sediments eroded during the observation period in 1987. Given the variability of bank toe deposition rates presented earlier, these estimates are likely to vary by approximately 22%.

7.5 Sediment pulses from bank collapse

This section attempts to assess the role of bluff failure in generating suspended sediment pulses. An example of a bluff failure observed in the field is presented to illustrate the general mechanisms involved:

Table 7.4 Calculation of 1987 bank sediment yields, Bas Arolla proglacial zone

Bank segment	Bank length m	Days	Erosion tonnes	Deposition tonnes	Balance tonnes	Percent stored	Comments
A	77.4	60	--	--	--	--	Stable
B	25.2	60	132.3	0	-132.3	0	
C	10.4	60	--	--	--	--	Stable
D	43.5	60	250.5	0	-250.5	0	
E	40.0	60	--	--	--	--	Stable
F	21.7	60	14.0	36.8	+ 22.8	263	See note (1)
G	14.8	60	16.5	2.9	- 13.6	18	
H	37.4	60	--	--	--	--	Stable
I	45.2	15	5.7	0	- 5.7	0	
J	29.6	15	3.7	0	- 3.7	0	
K	47.0	60	137.6	68.4	- 69.2	50	
L	49.6	60	--	--	--	--	Stable
	TOTAL		560.3	108.1	-452.2	19	

Notes: Days = Number of days from first exposure of bank
Volume - weight conversion = 1.9 (determined from bulk density samples of bank materials
(1) sediment was deposited at the base of this bank by material transport over the bank top
The bank segments listed in this Table do not correspond exactly with the sites listed in Table 7.2

1. *Time of collapse*

The collapse occurred at 16.00 on July 5th 1986. The collapse lasted approximately 4 seconds from the time of initial movement.

2. *Material supplied*

Volume of material supplied to channel = 20.4 m^3
Volume weight conversion = 1.9 (Bluff bulk density)
Mass of sediment supplied = 38.76 tonnes

3. *Proportion of material going into suspension*

Using the characteristic grain-size sample for the matrix of the bank (Figure 7.2b, Site 9).
Silt and clay fraction = (3.7%) 1.43 tonnes
Silt, clay and fine sand fraction = (13.3%) 5.16 tonnes
(Because this is based on a matrix sample this is likely to be an over-estimation of the amount of fine material).

4. *Local instantaneous peak concentration in the stream*

Discharge = 5.07 $m^3 s^{-1}$
(This is divided by 2 since sediment is introduced to only half of the stream) Therefore average peak suspended sediment concentration =

$$\frac{\text{load / bank length}}{\text{(discharge / 2) x duration of collapse}}$$

Which gives peak concentrations for:
Silt and clay fraction = 6410 mg l^{-1}
Silt, clay and fine sand fraction = 23131 mg l^{-1}
Background concentration = 9 - 13000 mg l^{-1}
However this material would rapidly become mixed in the flow and the peaks would soon be damped.

5. *Estimating peak concentration downstream*

The decay of the peak downstream is given by:

$$\frac{\textit{peak concentration at origin}}{\textit{distance downstream} \times 0.049}$$

The 0.049 is a factor estimated from the decay of sediment pulses determined in small scale experiments performed in the Bas Arolla proglacial stream (see below). Beyond 50 metres the peak concentration is halved because complete mixing is now assumed to have occurred. Results indicate the following peak concentrations (mg l^{-1}):

Distance downstream	*50 m*	*100 m*	*215 m*
Silt and clay	*1308*	*654*	*316*
Silt, clay and fine sand	*4720*	*2360*	*1141*

The accuracy of these calculations is undermined by several factors. The first problem is in estimating the volume of material supplied to the channel since an unknown volume is contributed below the water surface (sub-aqueous bank failure). However, this will be accounted for to some extent in the bank surveys that sandwiched the event. Defining the amount of material going into suspension is also difficult so two thresholds are used. The rationale for selecting the two size thresholds was that the turbidity meter only effectively monitors the silt size class but the fine sand size class was commonly observed to be in suspension. Indeed in sand-bed streams the proportion of suspended sediment in the banks correlates well with that in suspension (Pizzuto, 1984). This would be true in this example because of the high discharge and the turbulent conditions of flow. Estimated peak concentrations do not allow for the high concentrations of material already in suspension (9 - 13000 mg l^{-1}) which may reduce the capacity of the flow to carry the additional load. The decay factor shown in step 5 of the calculations was determined for small-scale experimental sediment pulses over short reaches, for small volumes of sediment, at low discharges. Therefore at high discharges where the capacity for transport is greater this simple decay factor may not be accurate. Furthermore the grain-size characteristics of the pulse will have considerable influence over the transport - distance decay function, although the same materials were used in both experiment and observed collapse.

Observations lend support to these calculations since average mainstream sediment concentrations were 9 - 13000 mg l^{-1} with some peaks of 34000 mg l^{-1} which compares favourably with predicted peak concentrations of between 15410 - 36131 mg l^{-1}. The largest pulses received at the measurement station during this period were only minor but the record is interrupted by purging of the sediment trap. However, given the high background concentrations it is unlikely that the pulses would be easily distinguishable, especially after passing through the sedimentation traps of the water intake. Because bank collapse only supplies a small amount of material of appropriate size to contribute to the suspended load, the remaining material will presumably be deposited on the bed where it will be stable or be transported as bedload. For this particular event (assuming that all the coarse sediment delivered to the bed is transported as bedload), 33.6 - 37.3 tonnes were removed as bedload which is equivalent to approximately 40% of the capacity of the water intake gravel trap. Since there was no appreciable accumulation of a bank toe deposited following this event it is likely that most of this material was removed as bedload. Boulders released directly into the flow are also likely to be transported since they will have lower thresholds of movement than those already sitting on the bed.

7.6 Sequence of bank change and sediment supply

Because bank erosion and the supply of sediment to the stream channel is so closely related to the proximity of previous high discharge events (Knighton, 1973); the susceptibility of banks to erosion (Lawler, 1984); and the presence of seasonal snowcover (Koutaniemi, 1984) it is often difficult to assess the importance of bank erosion measurements if the complete geomorphic history is unknown. In the proglacial environment geomorphic processes are strongly conditioned by the alpine seasonal cycle with runoff typically restricted to the period May to October. Early season runoff is dominated by snowmelt but later in the season glacial melt is more important.

Banks adjacent to the proglacial stream become snow-bound over winter and only emerge during spring melt. The combination and timing of the disappearance of the snowcover, peak nival runoff, peak glacial melt and occurrence of seasonal flood events largely determine the nature of the bank erosion processes.

Figure 7.9 attempts to summarise the seasonal sequence of bank change along the valley train of the Bas Arolla glacier. During May discharge is very low and snow infills the valley train. As thaw begins banks are exposed and begin to thaw. Thawing and the high pore water pressures in the thawed sediments encourages increasing bank increases as thaw progress. At this stage much of the eroded sediment is transported away from the bank toe by supra-nival pathways towards the stream channel. During June, banks are now virtually fully exposed snowmelt is at peak and discharge increases, usually culminating in a nival flood. Bank retreat rates tend to be relatively high. In late June banks begin to stabilise due to the absence of snowmelt and sediment begins to accumulate at the bank toe. July is usually characterised by a large flood event which

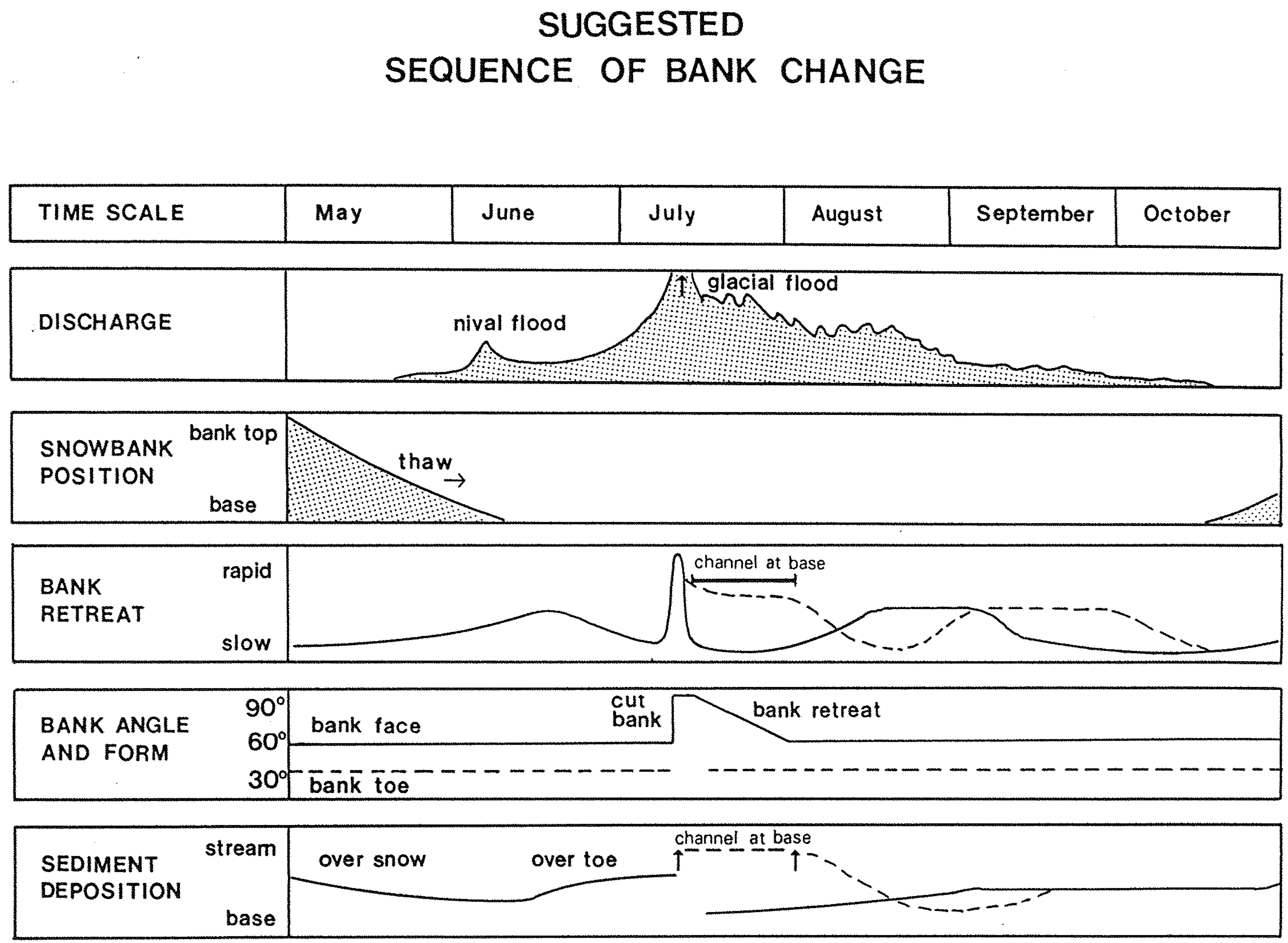


Figure 7.9 Suggested sequence of bank change for bluff along Bas Arolla proglacial zone.

results in widespread bank erosion. This results in cut banks which gradually re-adjust unless the stream channel remains at the bank base, in which case vertical banks and rapid rates of retreat are maintained. During August re-adjustment of the bank profiles after the July flood leads to sediment accumulation at the bank toe so that the banks begin to stabilise. September is a period of relative bank stability with minor readjustments in the profile. By October bank activity is at a minimum. Bank retreat may pick up very briefly at the end of October due to frost action on the exposed banks but this is usually shortlived because of the return of the snowcover. This sequence of events is not based on field measurements but it appears but appears to fit field observations. It is a conceptual model that does not conform to other studies which suggest that frost action is the strongest control on bank erosion (Lawler, 1986).

Figure 7.10 is a composite of retreat measurements for selected sites in 1986 and 1987 and attempts to summarise the sequence of bank retreat. The sequence is similar to that outlined in the conceptual model of Figure 7.9. Initially the 1986 flood event produced massive retreat rates, which were followed by a lag period of no response and then increasing adjustment in the bluffs until the middle of September. No observations were carried out over the winter but resurvey of bank 2 in early June suggested relatively low retreat rates during this period. Following snowmelt, retreat rates were very high and gradually reduced thereafter. The effect of the 1987 flood event was not very pronounced at these two sites.

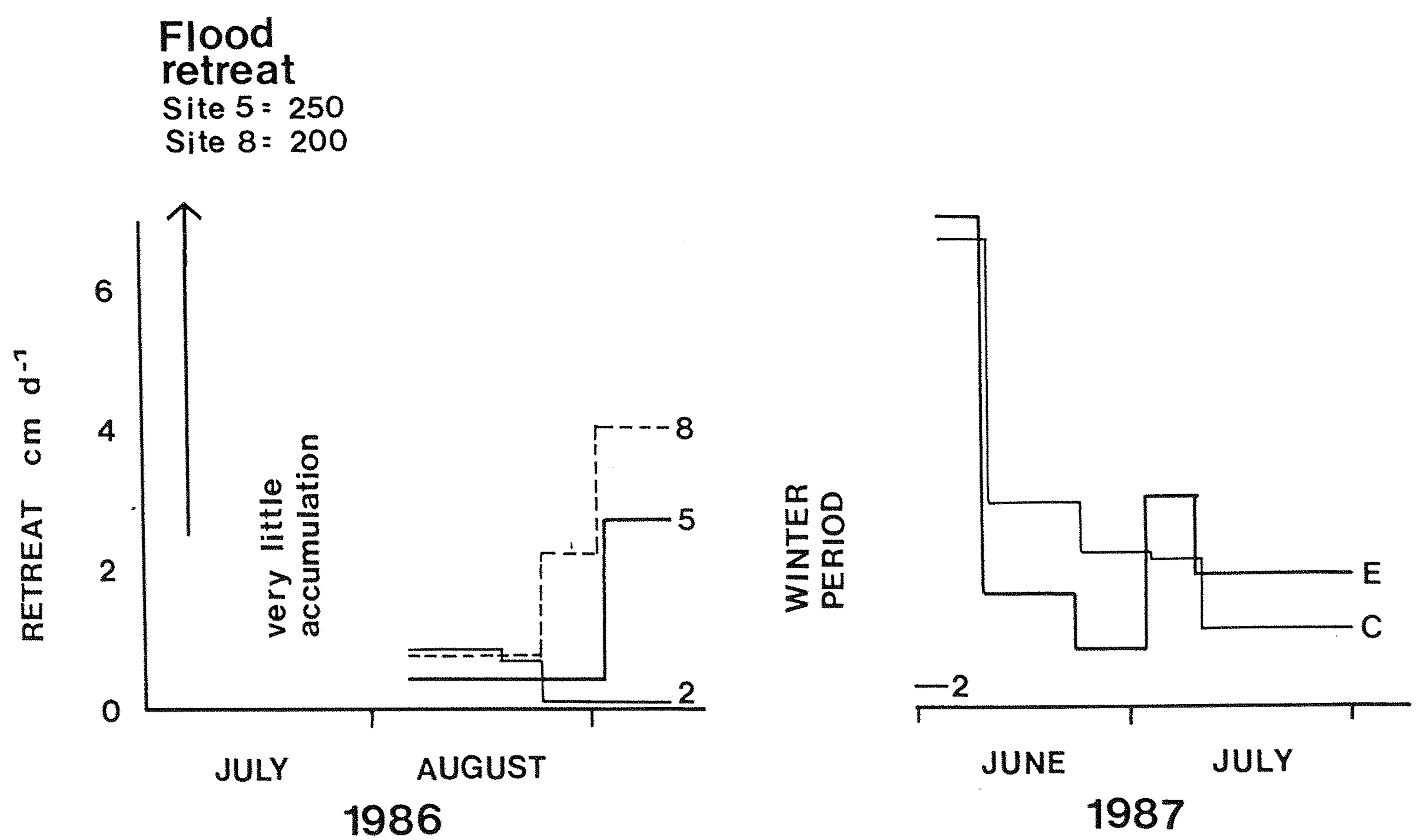


Figure 7.10 Diagram showing bank retreat rates at selected sites for 1986 and 1987.

7.7 Summary

Stream bank erosion is an important process contributing sediment to the load of proglacial streams. Measurements of bank profile retreat at 19 sites along the proglacial melt-water channel of the Bas Glacier d'Arolla provide a basis for the development of a conceptual model relating bank erosion and adjustment to sediment supply (Figure 7.9). The dominant mechanisms causing retreat are stream undercutting initiating buttress failure of non-cohesive boulder and gravel banks, and the collapse of large boulders. Bank profiles are substantially modified. The upper bank retreats in parallel , maintaining steep bank angles (up to 70^{o} or more). The lower bank consists of accumulated toe deposits at an angle of approximately $30\text{-}35^{o}$. This form is characteristic except at sites where there is under-cutting or in situ-basal protection. Accumulation of boulders at the bank toe exerts a strong basal endpoint control, and a state of impeded removal exists. As a consequence sediment supply from bank erosion will also be greatly diminished. Relative stability is maintained as long as discharge in the proglacial channel is incompetent to move boulders or the flow stage is too low to breach basal boulder accumulations. This simple model is complicated by the presence of seasonal snowpacks at the base of the bank, which assist in transmitting sediment beyond the bank toe. The significance of bank erosion is evaluated in the context of sediment supply to the valley train and, as a mechanism responsible for generating sediment pulses by bank collapse. Results show that large sediment pulses are unlikely to originate from bank collapse.

Attempts to control bluff erosion in the Bas Arolla proglacial zone by the hydro-electric company have been restricted to the lower 60 m of the channel immediately above the meltwater intake. Two approaches have been used: shaping the bank into a stable form using heavy machinery, and facing the banks with large boulders. The latter method has only been partially successful since the large boulders used for facing rest on the fine gravel bed of the main channel. Removal of the channel fines by development of a scour hole at the bank toe undermines the large facing blocks leading to bank instability. This is the most common mechanism for initiating failure in stream banks (Galay et al., 1987). Therefore, facing of streambanks is insufficient, and the toe must also be protected by providing a 'launching' apron (Galay et al., 1987). Following erosion at the toe, sediment supply may be enhanced in this lower channellised reach because, in an attempt to constrain flood flows, bank top elevations are often increased, providing a greater potential for sediment supply. For the 1987 sediment budget period the bluffs supplied 452.2 tonnes of sediment.

Chapter 8.

SEDIMENT FLUSHING EVENTS

8.1 Introduction

Time series of suspended sediment concentration and discharge observed in proglacial streams show clear diurnal cycles (Gurnell, 1987). This characteristic of proglacial stream time series is disrupted by non-periodic suspended sediment discharge events (Østrem, 1975). These non-periodic events take two principal forms:

1) Meltwater flood events (high magnitude - low frequency).
2) Pulses - short-lived bursts in suspended sediment concentration (low magnitude - high frequency).

Meltwater floods and sediment pulses are two ends of a spectrum of sediment transport events. These vary from small sediment pulses which bear no association with discharge, through events related to releases of small to large water pockets, to major outburst events. This chapter assesses the origins of these events and their importance to the seasonal sediment budget. In addition, because the chapter deals with events which are glacio-hydrological in origin, some of the links that exist between the glacial and proglacial water and sediment systems will be considered.

Meltwater floods are generated by water escaping from ice-marginal lakes, glacio-volcanic activity and the drainage of internal water bodies in the ice (Sugden and John, 1976). Of these three mechanisms the drainage of ice-dammed lakes has been frequently discussed (e.g. Collins, 1979; Shakesby, 1985) but the drainage of englacial water bodies less so. This is not surprising since Haeberli (1983), in his review of the frequency and characteristics of glacier floods in the Swiss Alps, suggests that 60-70% of observed floods are caused by outbursts of marginal glacier lakes or sudden breaks in ice dams, and the minority, 30-40%, by ruptures of englacial water pockets.

The flooding discussed in this chapter is of the latter type and involves the drainage of stored water from the Bas Glacier d'Arolla. Meltwater floods are important in the Swiss context because historically glacier flooding has been responsible for many catastrophes (Haeberli, 1983). In assessing the effect of these meltwater flood events, Haeberli (1983) makes the distinction between 'unusual' and 'annual' events, since on some glaciers events occur annually (e.g. Gornersee, Collins, 1979) whilst on others non-periodic events occur. In addition, a third category - periodic - should perhaps be added to these since some glaciers exhibit longer term cycles in meltwater release e.g. Grimsvötn, Iceland (Björnsson, 1974). Floods from the Bas Glacier d'Arolla would, therefore, seem important since in this part of Switzerland meltwater flooding is common place (Hagen, 1944; Beecroft, 1983; Haeberli, 1983) and most floods are non-periodic, originating from the rupture of water pockets which cannot be predicted.

The terminology of meltwater flood events is confusing since these events have also been termed meltwater outbursts and jökulhlaups. These terms are avoided in this chapter since they infer a catastrophic release of water from the glacier e.g. a glacier outburst flood is a ....'sudden and often catastrophic release of bodies of water impounded by ice' (Young, 1980); a Jökulhlaup is a ....'catastrophic glacial outburst flood' (Jackson, 1979). Use of the jökulhlaup is particularly misleading since this is generally reserved for the sudden and rapid draining of glacier dammed lakes (Paterson, 1981). Without detailed study of each meltwater flood the exact mechanisms cannot be reliably established and genetic terminology should be avoided.

Although frequently reported, the origins of meltwater floods and their contribution to sediment yield have not been evaluated in any detail (Young, 1980). Observations characteristics of meltwater floods include: a sudden release of water (Haeberli, 1983); large and variable discharges (Beecroft, 1983); high sediment transport rates (Østrem et al., 1967); extensive channel erosion (Hewitt, 1982); and triggering of debris flows and slope failures (Jackson, 1979). Observations of these events are rare (Young, 1980) but from the limited information available sediment transport rates are high e.g. Beecroft (1983) for the Tsidjiore Nouve glacier, Switzerland estimated that 19% of the ablation season suspended sediment yield was transported in a 3 day period and Østrem et al., (1967) on the Decade glacier, Baffin Island suggested that 60% of the entire seasons suspended sediment output was transported in a single day event. Haeberli (1983) estimated that the ratio of debris to water in the total volume of material moved was commonly 0.35 but could be as high as 2.4, at which stage the flood is more like a low density debris flow (Davies, 1985). Given these characteristics meltwater floods would appear to be very important for the sediment budget of the proglacial zone (Johnson and Power, 1985).

A suitable context for evaluating the impact of meltwater floods would be in terms of 'work' and 'effectiveness' (Wolman and Gerson, 1978; Newson, 1980). The work of the flood can be evaluated in terms of the sediment transported during the event and the effectiveness in terms of the geomorphological change brought about by the event. The effectiveness concept can be interpreted in terms of the sediment budget as a means of identifying the main source, transport and storage elements during flood. However, evaluating floods in terms of work and effectiveness is not enough, a third dimension 'response' needs to be added. In the case of the proglacial environment the response to flood normally manifests itself in adjustments in the channel system (Figure 8.1). It is hypothesised that in response to increased sediment load and discharge imposed by the flood, the proglacial fluvial sediment system temporarily adjusts by increased braiding and sinuousity and a reduction in channel slope (Figure 8.1). In the aftermath of the flood, sediment transport declines to levels below those prior to flood; braiding and sinuousity are reduced as the channel rationalises; and slope increases in response to a single thread channel incising in to the valley train. Sediment transport rates are lower than before flood due to exhaustion of stored sediment or reduced availability of sediment.

In an attempt to evaluate the importance of these events the first part of this chapter describes the characteristics of three meltwater floods from the Bas Arolla glacier, discusses their origins and quantifies the contribution of the events to the overall seasonal sediment budget (Section 8.2).

In the second part of the chapter small-scale, short-duration sediment pulses in the suspended sediment series of the Bas Arolla proglacial stream are discussed (Section 8.3). These pulses although several orders of magnitude smaller than meltwater floods occur frequently (Østrem, 1975; Bogen, 1980; Gurnell, 1982) and represent the opposite end of the magnitude- frequency spectrum. These events have been variously described as pulses, bursts and flushes, all of which suggest a rapid short-lived release of sediment. In Section 8.3 small scale field experiments illustrating the transport of sediment pulses in proglacial streams (section 8.3.1) provide an introduction to the discussion of the origin of natural sediment pulses and how observations of pulse

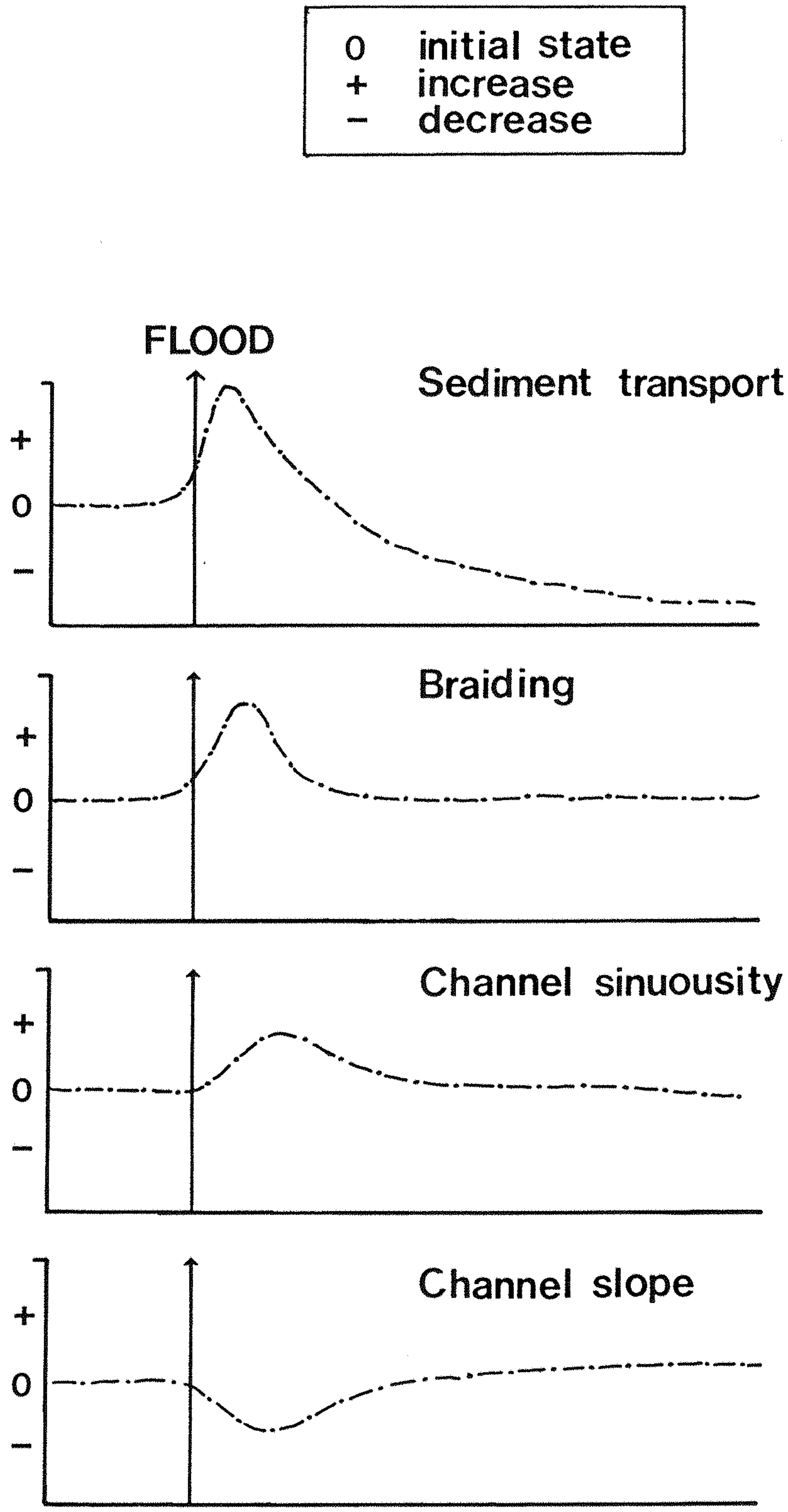


Figure 8.1 Hypothetical response of the proglacial sediment system to flood.

activity can be used to make inferences about glacial hydrology (Section 8.3.2).

8.2 Flood events in the Bas Arolla proglacial zone

During the two years of study (1986 - 1987) three major flood events were recorded in the proglacial zone of the Bas Arolla Glacier. In 1986 a single flood event occurred (5th - 6th July) and in 1987 two floods occurred (15th - 19th July and 24th - 25th August). The two July events originated in the glacio-hydrological system of the Bas Glacier d'Arolla but the August 1987 event differed in that it was generated from a manual water release from the Haut Glacier d'Arolla catchment. The flood waters from the Haut Glacier d'Arolla were released from a high level gallery which transmits water from the Zermatt valley over to the storage reservoir (Lac des Dix). This is not normal practice by the hydro electric company but was necessary since the capacity of the upper galleries was exceeded due to high summer melt and intense prolonged rainfall of the previous few days. In this respect the two July events should be regarded as natural in that they appear to be an annual phenomena but the August event is unusual and should therefore be viewed as an artificial anomaly.

8.2.1 Discharge Characteristics

It is notoriously difficult to estimate flood discharges in proglacial streams because of changes in channel cross-sections during flood, destruction of gauging apparatus and the inaccuracies involved in applying hydraulic reconstructions (Costa, 1983; Komar, 1987). Figures 8.2 and 8.3, based on records collected at the Bas Arolla meltwater intake show the seasonal and detailed discharge series for the three events. Seasonally the discharge cycles for both years are characterised by 3 main periods of discharge: an early season spike generally occurring in June, and two main periods of high discharge peaking in July and August. It is during these two latter seasonally-high discharge

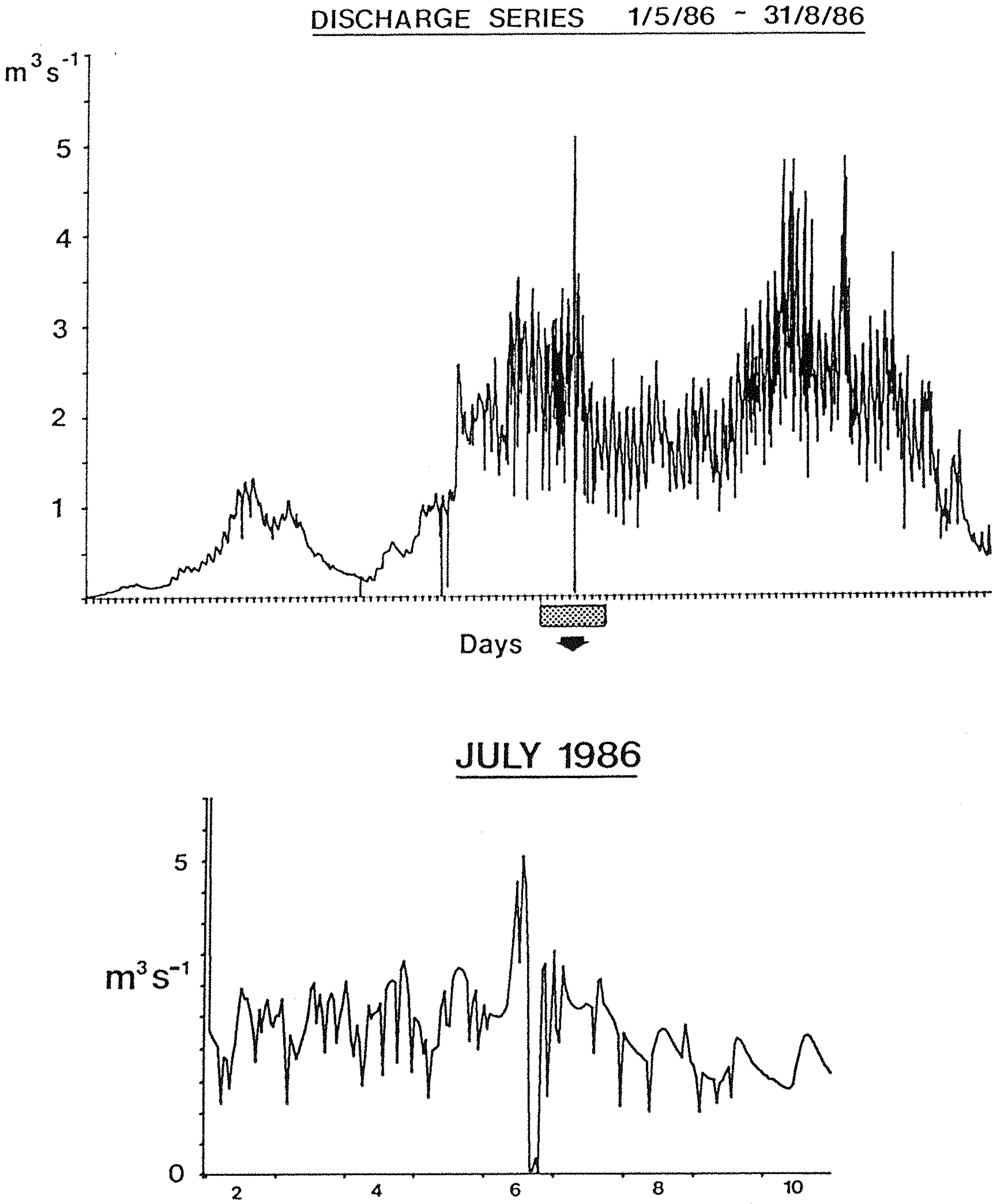


Figure 8.2 Seasonal and flood discharge series 1986.

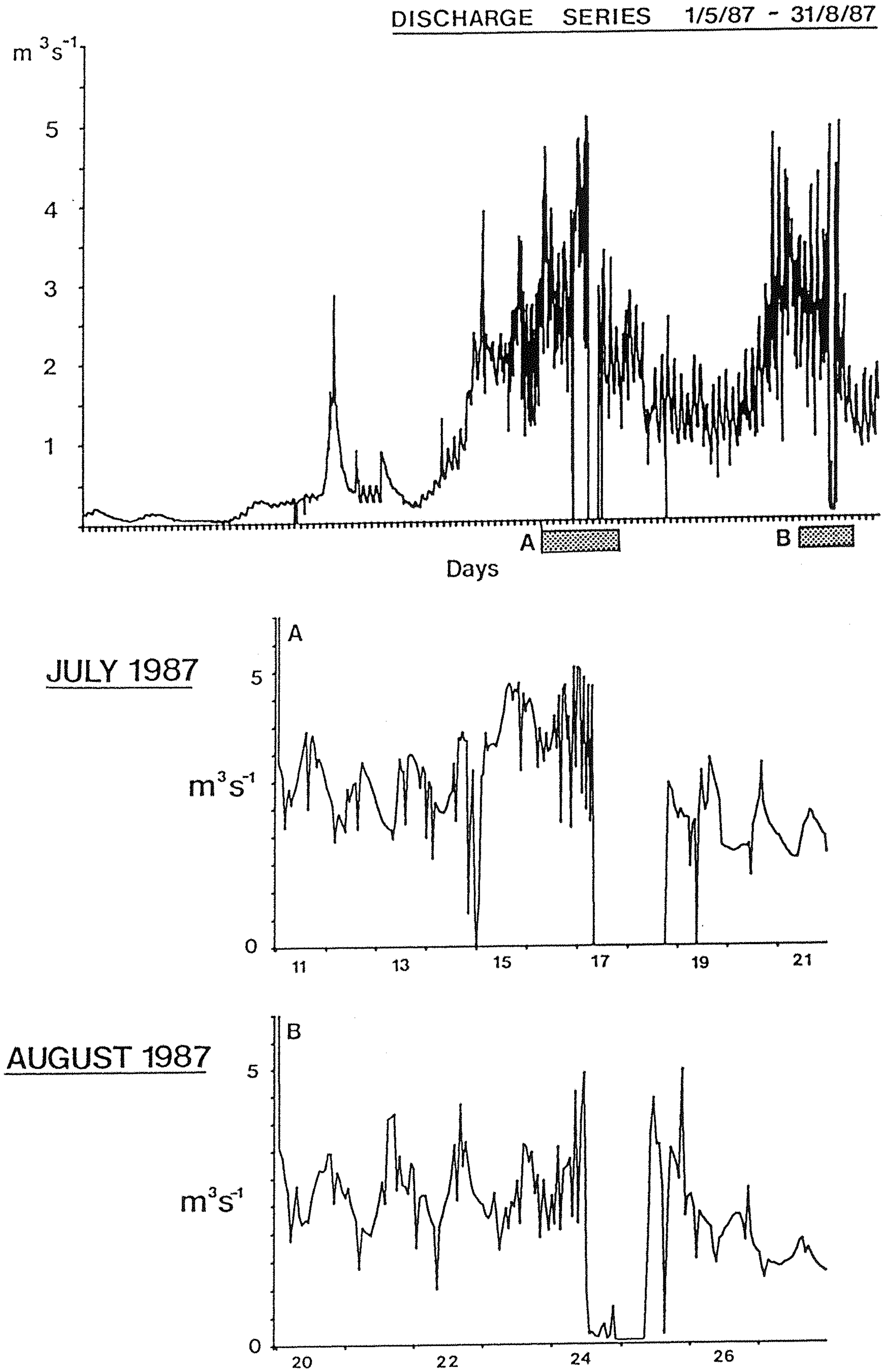


Figure 8.3 Seasonal and flood discharge series 1987.

periods that meltwater floods occur and are generally most effective (Haeberli, 1983). The main differences between the two years is that the 1987 seasonal pattern occurred slightly later in the year and its peak discharges were more pronounced. The series are incomplete during the periods of peak discharge because water was diverted away from the stage recorder to prevent damage to underground installations. However, peak discharge was estimated from the flow of water through the open meltwater intake. Under these circumstances the intake structure acted as a rectangular weir and, using floats to estimate velocity, discharge could be approximated. This could not done for the August 1987 event since this was outside the field study period, therefore discharge was estimated from computations of the excess water released from the Haut Arolla high level gallery (Chevalley, pers. comm.). Comparison of the timing and discharge values for the three events (Table 8.1) with other major Swiss glacier floods (Haeberli, 1983) shows that the three events occur in the two commonest months with the two natural events corresponding with the modal month (Figure 8.4a). The relationship between flood volume and average slope which defines the damage potential of the event (Haeberli, 1983) suggest that the events were unlikely to have been a major hazard (Figure 8.4b).

The pattern of discharge leading up to the July 1986 flood (Figure 8.2) consisted of a series of broken diurnal cycles with a slightly increasing trend in discharge. This was followed by a very rapid rise on the 6th of July (which is truncated at 5 $m^3 s^{-1}$ by overflow at the intake structure) to an estimated peak discharge of 7-9 $m^3 s^{-1}$. Following the event on the 6th of July, flow was still high but steadily declined in the following days. In July 1987 the flood period was more complex (Figure 8.3). Prior to the flood, discharge was reasonably high with a peak discharge of about 3 $m^3 s^{-1}$ which was maintained until the

Table 8.1 Characteristics of the Bas Arolla meltwater floods

CHARACTERISTIC	JULY 1986	JULY 1987	AUGUST 1987
Nature of the event			
Peak discharge m3 s-1	7 - 9	10	20
Period of flood	July 5-7th	July 15-19th	August 24-25th
Rainfall mm	40.3 in 14h	15.6 (17th) 20.1 (18th)	Prolonged intense storms
Flow pathway	Snout zone east margin	Snout zone east margin over the snout	Rear of glacier east margin
Release of meltwater from HEP scheme?	No	No	Yes
Peak suspended sediment concentration	13509 - 34890 (6/7)	31108 - 39476 (15/7) 19085 - 27112 (18/7)	Not sampled
Electrical conductivity	rise on (6/7)	slight rise 2200h (15/7) High on 19/7	No significant change
Pre-conditions			
Constricted englacial and subglacial drainage?	likely	likely	No
High ablation rates?	Yes	Yes	Yes
Rainfall or previous storm activity?	Yes	Yes	Yes
Water storage in glacier snow cover?	No	No	No
Observed events			
High water levels close to glacier surface?	Yes	Yes	Unlikely
Sudden drainage of surface water?	Yes	Yes	?
Uplift of glacier?	?	May be	?
Increased seismicity at the bed?	?	?	?
Fracturing of glacier surface?	?	Yes	?
Large suspended sediment loads?	Yes	Yes	Yes
Increase in mainstream electrical conductivity?		Yes	No
Sudden changes in subglacial and proglacial discharge?	Yes	Yes	Yes
Shift to marginal drainage?	Yes	Yes	Yes

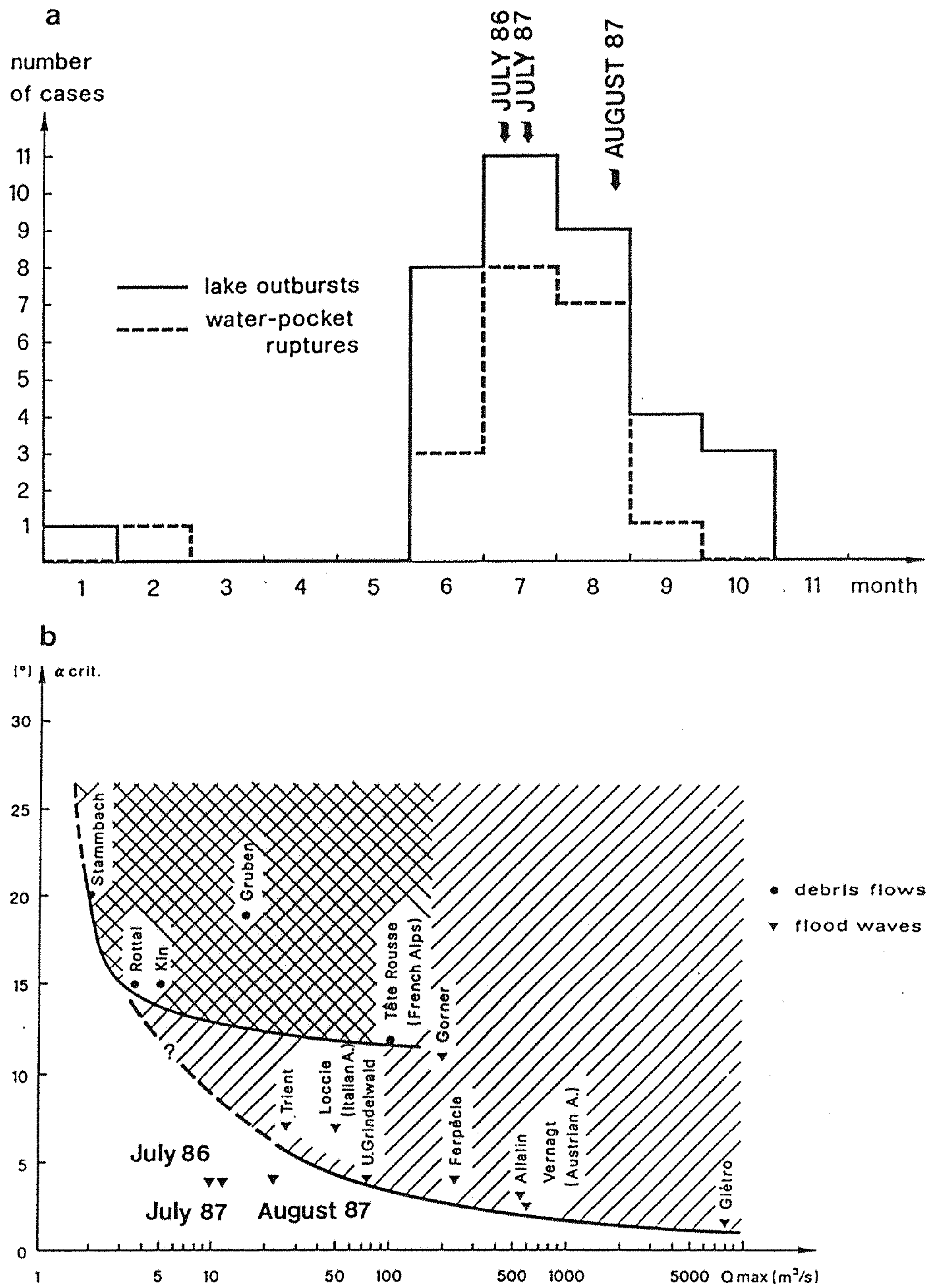


Figure 8.4 (a) Seasonal distribution of historical glacial floods in the Swiss Alps. (from Haeberli, 1983) showing three Bas Glacier d'Arolla floods. (b) Extent of damage along the glacier stream for sufficiently documented glacier floods in the Swiss Alps
α crit = average slope between the place of rupture (usually the glacier terminus) and the limited of recorded damage, (from Haeberli, 1983). Three Bas Glacier d'Arolla floods 1986-1987 are plotted.

evening of the 15th of July. At 2200h on the 15th flood waters were rapidly released from the glacier causing the discharge record to be truncated (the dip in the record). The estimated peak discharge using the weir and float technique was 10 $m^3 s^{-1}$. Following this initial flood wave, in the days between 16th and 19th July, discharge remained high and flood waters were continually diverted away from the stage recorder. A second flood wave occurred at 1000h on 18th July and peak discharge was estimated to be 10 $m^3 s^{-1}$). After 19th July discharge declined steeply. The August event in the same year showed a pattern of high (2.5 $m^3 s^{-1}$) irregular discharges for the days before the event, followed by rapid truncation of the discharge on the August 24th (Figure 8.3) as waters were catastrophically purged from the Haut Arolla catchment. Water was diverted past the stage recorder for almost 24 hours on 24th to 25th August. After the event discharge declined rapidly. Because the event was unobserved, discharge estimates could not be checked by the weir and float method. However, based on the observation that the downstream effects of the flood were greater than the two July events (causing erosion and deposition problems up to 5.5 km downstream) then the Grande Dixence estimate of peak discharge of 20 $m^3 s^{-1}$ (Chevalley, pers. comm.) appears reasonable. All three peak discharge estimates are greater than 2.45 $m^3 s^{-1}$ which was estimated for a 'minimal outburst' at the Bas Arolla in 1974 (Haeberli, 1983).

8.2.2 Description of the three Floods and discussion of likely origins

Determining the origin of meltwater floods is fraught with difficulty since the processes governing the release of the flood are acting within the glacier. The inaccessibility of the glacial hydrological system means that field observations, indirect measurements and theoretical deductions are often used to characterise glacier hydrology (Rothlisberger and Lang, 1987). The following discussion uses field observations and measurements to infer the characteristics of the three floods. The August 1987 flood is discussed on the basis of reconstruction of the event from the available geomorphic evidence 14 days after its occurrence.

Events of July 6th 1986 - Heavy overnight rainfall of the 5th/6th July produced extensive runoff in the Bas Arolla glacier catchment with high flows in valley side tributaries and supraglacial streams. Peak discharge occurred on the 6th between 1300h -1500h and was estimated to be 7 - 9 $m^3 s^{-1}$. The main proglacial stream emerged from the snout as an up-welling, suggesting that the flow was driven by a considerable pressure head. Main stream flow was augmented by two tributary streams; a large stream on the eastern margin and a smaller tributary to the west (Figure 8.5a). Both tributaries emerged from portals in the ice about 50 m to each side of the main (central) stream outlet. All three outflow streams had very high suspended sediment concentrations (13509 - 34890 mg l^{-1}) and were rich in organic matter and ice debris. Flow eroded a large channel along the eastern margin of the glacier and caused extensive bluff erosion along the margins of the valley train (Chapter 7). Observations suggested that vertically cut-banks, mainly composed of non-cohesive till, were formed by stream under-cutting initiating buttress failure. All channel bar features were 'drowned out' as the flow spread across thee full width of the valley train. Flows were so high that water over-topped the meltwater intake structure and flowed around the sides (Figure 8.6a). Flow around the intake caused large-scale gully erosion removing 175 tonnes of debris. The gravel trap over-filled with sediment and cobbles (up to 20 cm diameter) were transported over the sides of the trap. The trap filled with boulder gravels with the largest boulders commonly between 30 and 60 cm diameter.

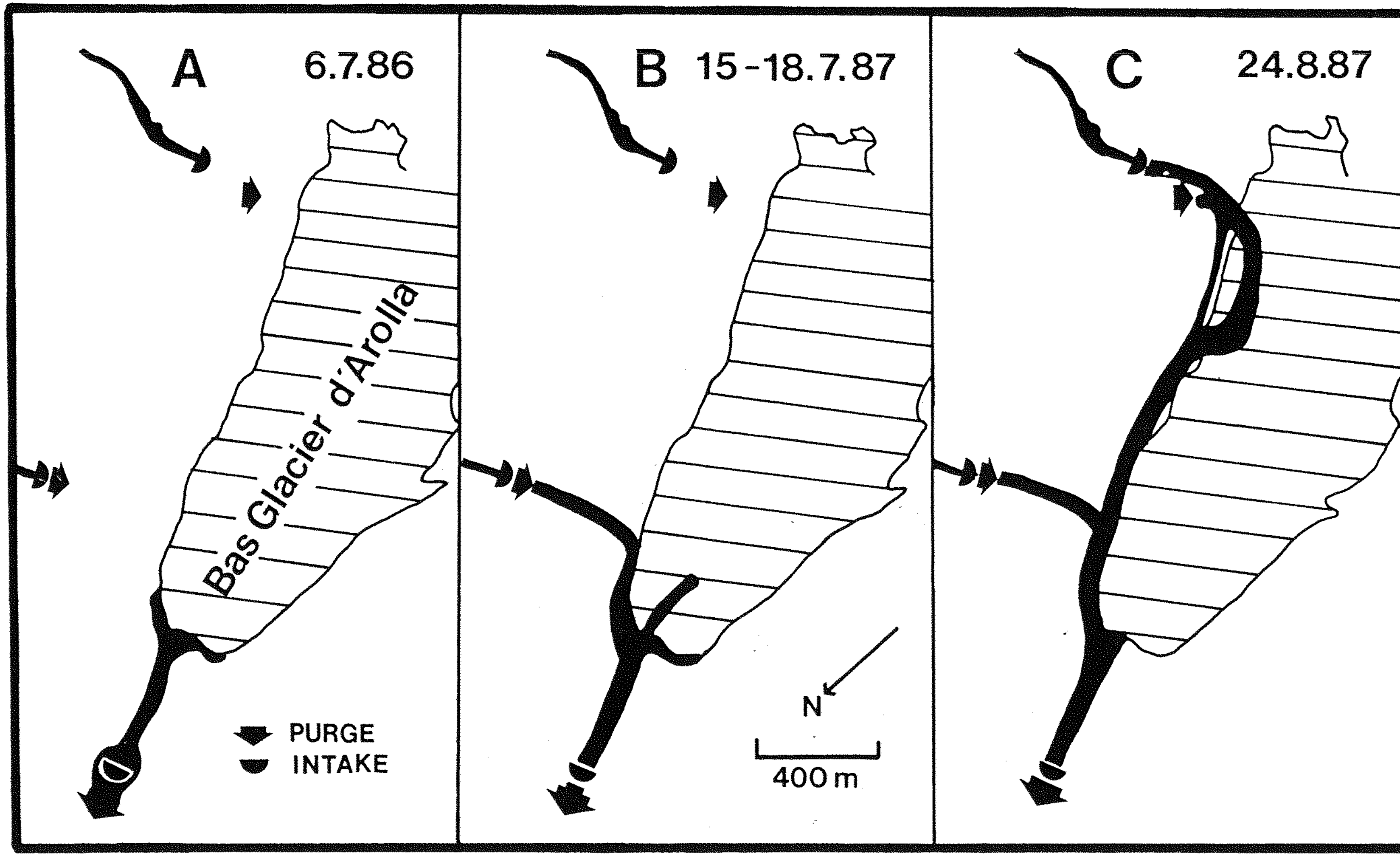


Figure 8.5 Pathways of flood water for the 1986 and 1987 Bas Glacier D'Arolla floods.

Figure 8.6 (a) Upper photograph - Bas Glacier d'Arolla proglacial zone during the 1986 July flood, 1615 h. Note by-passing of meltwater intake by flood waters. (b) Lower Photograph - showing the overflow channel (shown as a thread of white ice), at the snout of the Bas Glacier D'Arolla, associated with the flood of July 15th 1987.

Events of 15th - 19th July 1987 - The combination of high summer melt and intense rainfall caused a complex sequence of flooding at the Bas Glacier d'Arolla between 15th and 19th July. At 2150h on July 15th a large overflow from two active moulins cascaded over the snout of the Bas Glacier d'Arolla (Figure 8.5b). The catastrophic nature of the event caused the upper monitoring station sensor mount (constructed of dexion) to be bent at right angles. This supraglacial overflow event carved a new 150 m channel joining the two moulins to the snout of the glacier (Figure 8.6b) and caused a large volume of supraglacial debris to be transported over the snout and deposited at the head of the proglacial zone. This probably caused the audible 'explosion' which accompanied the overflow event. One moulin was significantly enlarged during the event, and both displayed clear wash lines on their walls well above stream entry levels (Fenn et al., 1987). Comparison of channel capacity upstream and downstream of the moulins suggested that the overflow waters resulted from bottom-up back-filling of the moulins rather than top down over-filling (Fenn et al., 1987). This indicates a poorly developed glacial drainage system which was unable to transmit the water subglacially. The stress caused by the over-filling of the englacial cavities caused parallel arcuate fractures across the glacier. Suspended sediment concentrations in the main channel were very high (31108 to 39476 mg l^{-1}) and the main stream switched between braiding and down-cutting modes in response to discharge pulses. Large quantities of ice and organic material, were eroded from the margin of the glacier and transported by the flow. Sediment purged from the sediment trap during this period (2300h) consisted mainly of boulders (50 - 60 cm diameter) set in a sandy gravel matrix.

On 16th July large fresh vertical fractures were observed on the face of the glacier and their appeared to be a pronounced bulge in the glacier face which was not evident on 15th July. Tracing the fractures on the glacier surface, suggested that fracturing tended to focus on the two large moulins involved in the overflow event. The largest concentration of the fractures occurred on the eastern side of the glacier in the area where the main marginal stream emerged from the subglacial drainage system. Discharge was much reduced on 16th July, but short duration discharge pulses (5 minutes) continued throughout the day and the channel remained in a braided state.

On 17th July channel pattern changed little but the length of discharge surges increased to about 15 minute duration and sediment was occasionally discharged over the top of the water intake. Early on 17th July (0730h), marginal discharge channels were dry but by 0940h discharge had started again. Evidence of tributary erosion close to the glacier (Tributaries T1 and T2, Chapter 5.) was widespread, fence traps were destroyed and tributary tracers were found in the main channel.

Subglacial release of flood waters was finally achieved at 1000h July 18th. This second flood wave was released rapidly and may have been generated by hydraulic jacking (Rothlisberger and Lang, 1987) at the glacier bed. The upper monitoring station was washed away by the flow which is good evidence of the rapid release of water.

The release of these flood waters caused large-scale proglacial channel change and sediment movement. High rates of sediment transport and channel movement continued throughout July 18th. Large amounts of organic material and ice were transported during the flow and some large ice blocks (up to 70 cm diameter) were imbricated in the valley train gravels. The pattern of channel change during this period involved initial widening and braiding followed by down cutting and the transition to a single thread of dominant flow (Chapter 6).

Reconstruction of the events of 24th August 1987 - Heavy rain storms and high summer melt in August 1987 combined to produce exceptionally high runoff in the glacierized catchments of Southern Valais. This runoff is normally exploited by the Grande Dixence hydro-power scheme but in this event runoff exceeded the capacity of the collecting galleries. In order to reduce excess flow in the high level gallery between Zermatt and Arolla, water was discharged from the Haut Arolla purge outlet (Figure 8.5c and Figure 8.7). In addition, runoff from the the Haut Arolla basin could not enter the already full high level gallery and therefore by-passed the meltwater intake and flowed down towards the Bas Glacier d'Arolla. Discharge from the Haut Arolla basin and from the purge outlet combined and flowed down towards the Bas Glacier d'Arolla (Figure 8.5c). The flow caused large-scale gully erosion leaving only large boulders lining the channel bed (Figure 8.7 (1) and Figure 8.8a). From the bottom of the gully the flow crossed the moraine at the rear of the Bas Glacier d'Arolla, depositing boulder sized material up to 1 m diameter. At this point the flow divided; the main part continued over the ice, flowing down glacier parallel to the margin. A secondary flow travelled along the glacier margin (figure 8.7).

Where the main flow travelled across the glacier surface a flood channel (Figure 8.8b and Figure 8.7 (2)) was incised (depth 1-1.5 m, width 4-7 m). Large angular moraine blocks and minor amounts of gravel were deposited in the channel but they may have accumulated after the passage of the flood wave. Flow continued down glacier into a series of crevasses which were transverse to the direction of the flood (Figure 8.7 and Figure 8.9a). This had a pronounced effect on flow, with debris (mainly gravels) being deposited in the

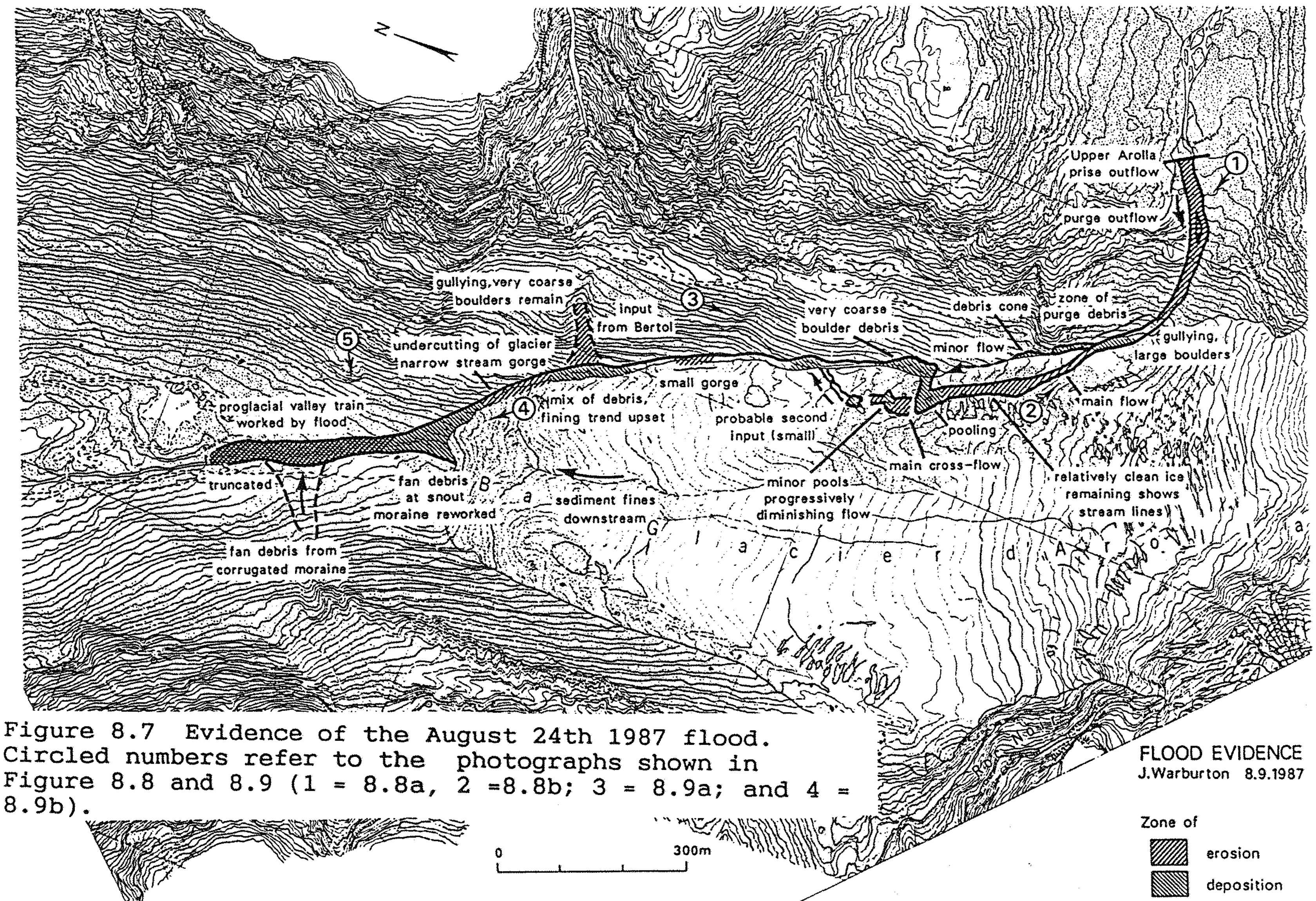


Figure 8.7 Evidence of the August 24th 1987 flood. Circled numbers refer to the photographs shown in Figure 8.8 and 8.9 (1 = 8.8a, 2 =8.8b; 3 = 8.9a; and 4 = 8.9b).

Figure 8.8 (a) Upper photograph - View looking down the gullied area below the Haut Arolla meltwater intake. The last 1987 flood channel can be seen continuing over the glacier surface. (b) Lower photograph - View looking up the August 1987 flood channel which was cut into the glacier surface (rucksack for scale).

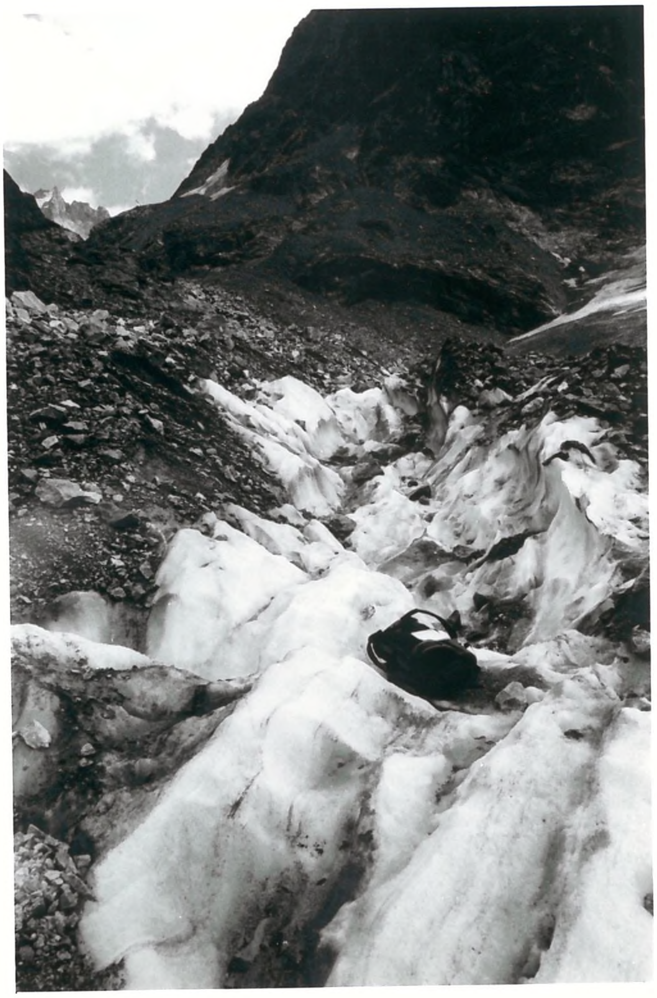


Figure 8.9 (a) Upper photograph - View showing the August 1987 flood channel on the surface of the Bas Glacier d'Arolla. The channel appears to end in a series of crevasse depressions in the centre of the photograph. (b) Lower photograph - Gullying and undercutting along the margin sof the Bas Glacier D'Arolla by the August 24th 1987 floodwaters.

crevasse depressions and sections of intervening ice being swept clean of debris. The net effect of the crevasses was to create ponding which reduced flow velocity. Ponded water spilled laterally off the glacier surface, via crevasse lines, to the glacier margin (Figure 8.9a and Figure 8.7). Flow was channe lised along the margin of the glacier, depositing a large volume of debris which showed a pronounced downstream fining in boulder sizes. The width of the flood channel in this area was restricted to between 5 and 15 metres. Along the glacier margin, flow under-cut both the lateral moraine and talus slopes. Deposition, in the form of small boulder clusters behind large stable boulders, was common and large organic debris, from old engineering works, formed log jams. Channel slope in this reach was between 5^o and 10^o This section ended at the confluence with the Torrent de Bertol stream (Figure 8.7). At this point an input of sediment from this stream deposited a large cone of debris which was subsequently incised and truncated by the main marginal flow.

Beyond the confluence with Torrent de Bertol, the channel steepened greatly ($25\text{-}30^o$) and turned with a northerly trend, towards the North West along the margin of the glacier. Flow continued towards the snout of the glacier in a series of step-pool plunges. Gullying at this point caused undercutting of the glacier and exposure of the underside of the glacier (Figure 8.9b and Figure 8.7). Flow emerged from the steep section and deposited a large fan (Figure 8.7) before joining the main proglacial stream which emerged from the snout of the Bas Glacier d'Arolla. Flow continued down the valley train, reworking the full width and causing excessive bank erosion along both margins. A large debris fan, deposited on the western side of the proglacial zone, was truncated at the base by stream flow at the margin of the valley train (Figure 8.7).

The geomorphic effectiveness of the three events is best evaluated in terms of the extent of their influence, the sediment sources activated during the event and the landforms produced (Newson, 1980). Although essentially qualitative, this approach is useful when combined with quantitative measurement of sediment yield, since it allows the controls on sediment delivery to be established. The July 1986 flood was confined to the immediate valley train and involved contributions from subglacial and proglacial sediment sources with minor amounts of sediment eroded from the side slopes. The most significant feature of the flood was the gullying around the meltwater intake structure. Bank erosion was widespread and the channel switched to a braided regime. The influence of the July 1987 flood was more widespread, involving glacial and proglacial areas, with water flowing over the snout of the glacier. A greater variety of sediment sources were involved including supraglacial, subglacial and proglacial sources with a minor contribution from the side slopes. Channel change during this period exhibited a complex response, moving from an initially braided channel via incision to a dominantly single thread stream. The August 1987 flood was very different because the release of water was from the Haut Glacier d' Arolla. Sediment sources in all areas of the glacierised catchment were involved. Sediment was eroded as a result of gullying, from erosion of supraglacial moraine, from talus along the glacier margin, from side slopes, subglacial sources and within the proglacial zone. Channel pattern altered radically and bank erosion in the valley train was widespread.

Borrowing the terminology from Newson (1980) and applying it to the glacierized catchments these floods can be classified in terms of 'channel' floods, 'slope' floods or 'glacier' floods. In these terms the July 1986 flood would be a 'channel' flood the July 1987 a 'Channel-glacier' flood and the August 1987 event a 'channel-glacier-slope' flood. This is a useful approach in the context of sediment budget studies because it recognises the linkages between sediment stores and transport processes.

Given the differences in the character of the three floods it is important to determine the origins of the events. The August 1987 flood was generated by the release of water from the Haut Arolla catchment and meltwater gallery. The two July events however originated from the glacio-hydrology system and are more difficult to interpret. Table 8.1 outlines the characteristics of the three floods along with check-lists for the pre-conditions necessary for flooding and observations diagnostic of meltwater floods. The August 1987 flood is included in the table to aid comparison between managed and natural flood events. Apart from contrasts in discharge the main difference between the three events is that the August 1987 event occurred later in the season, it was generated by release of water from the hydro-electric scheme and it involved flow over the upper portion of the Bas Glacier d'Arolla. The two July events were very similar, the only significant difference being the snout overflow event during the 1987 flood.

In attempting to elucidate the origins of the July flood events it is initially assumed that the events originate from the rupture of internal water bodies within the ice. This assumption is reasonable for the Bas Glacier d'Arolla since no ice dammed lakes are present and a previously documented flood in 1974 (Haeberli, 1983) originated from such a source. The main pre-conditions (Table 8.1) required for meltwater flooding have been identified by Rothlisberger and Lang (1987). These include: the need for a restricted englacial and subglacial drainage network; extreme melt rates; storm precipitation and water storage in a glacier surface snowcover. For the July floods all these preconditions were met except for the presence of a glacier surface snow-water store (Table 8.1). The August 1987 event was unlikely to be associated with a restricted internal drainage network at that time in the ablation season and particularly since the event followed the July flood event. On these grounds an outburst origin for the August event could be rejected even if the origin of the event was not known. Given that the the pre-conditions of Rothlisberger and Lang (1987) were met for the two July events, an outburst origin for the events seems likely.

Field observations during the two July events (Table 8.1) can be attributed to what is termed the 'spring event' (Rothlisberger and Lang, 1987) and are similar to the characteristics of 'mini-surges' (Humphrey et al., 1986). These involve the following sequence of events: Firstly, following winter the internal drainage of the glacier has degenerated and most of the water passages in the glacier are considerably reduced in size (Shreve, 1972); Secondly, the capacity of the meltwater system is insufficient to transmit the supply of water when runoff increases in early summer (Hooke et al., 1985); Thirdly, as a result high water pressure results which acts at the bed to influence movement of the glacier. The water pressure may rise to equal or exceed the ice over burden pressure (Sharp, 1988). If the drainage network is poorly connected to the bed, local pressure points develop which act as hydraulic jacks driving the ice upward and forward; Finally, glacier movements of this kind allow new passages to open up and a new drainage network to rapidly develop. These events are short-lived because once the drainage connects with the bed numerous water links develop, the stored water rapidly drains (outbursts) and the pressure in the system is lost.

The significance of the observed events listed in Table 8.1 should now be apparent. High water levels close to the glacier surface indicate a meltwater drainage system full of water and one which is poorly connected to the bed. Sudden drainage of stored water results when the connection is made with the bed and water drains as a flood wave. Glacier uplift should be apparent if the hydraulic jack mechanism operates. This would be accompanied by seismicity involving ice-quake activity (Holmlund and Hooke, 1983) near the bed in response to the stress set up by the pressure within the glacier. The release in this pressure will result in surface fracturing of the glacier. Flushing out of stored water along newly opened subglacial passageways will lead to sudden changes in discharge and shifts in the drainage pattern. In the early stages of the event marginal drainage at the snout may occur because the main subglacial channel will still be poorly developed so forcing excess water to escape by marginal routeways. Connection with the bed should also be reflected in a change in the proglacial stream suspended sediment concentration and electrical conductivity. Suspended sediment concentrations rise rapidly since rock flour accumulated at the glacier bed is flushed out. Electrical conductivity should also rise because water is draining from subglacial cavities which should be relatively high in solutes.

For the two July events, where observations were available, most of the above criteria occurred (Table 8.1). Of the two events, the July 1987 flood was the most well-documented and can be considered in more detail. This event was remarkable in that on the 15th of July water overflowed the snout of the glacier. Flow was generated from a moulin 150 metres back from the snout. The moulin appeared to fill from the bottom up because the capacity of the supraglacial channel entering the moulin was less than that of the overflow channel below the moulin. Therefore, water was either supplied from subglacial drainage at the bed or from englacial drainage in the ice. If flow was from the bed then this implies that the subglacial drainage was temporarily blocked forcing water upwards. Alternatively the water could have entered the moulin englacially and then backed onto the glacier surface. Because there was no deposition of suspended sediment on the glacier surface, the water source would appear to be englacial and supraglacial (subglacial water would be laden with suspended sediment). This does not totally negate the influence of subglacial water pressure, because the overflow may be the result of displacement of englacial water as subglacial water rose in the drainage system. Fracturing was associated with this event but the main glacial drainage did not fully connect with the bed until the July 18th. On the 18th July a meltwater outburst occurred and water levels in the glacial drainage system fell rapidly. This event was probably associated with melting through of englacial cavities in the intervening 2 days since the initial overflow event.

8.2.3 Sediment contributions

Beecroft (1983) suggests that outburst floods are associated with high levels of geomorphic work, but can work can work satisfactorily explain the effectiveness of floods (Newson, 1980). For the three floods described here effectiveness and work characteristics can be compared.

The three floods have been described above as:

July 1986	'channel'
July 1987	'channel-glacier'
August 1987	'channel-glacier-slope'

Estimates of sediment transport during the three floods (Table 8.2) show marked contrasts in sediment yields measured at the Bas Arolla meltwater intake structure. The lowest total yield was associated with the July 1986

Table 8.2 Minimum Estimates of Sediment Transport during flood.

FLOOD EVENT	DAY	2 GRAVEL OUTPUT	3 SAND OUTPUT	4 SUSPENDED SEDIMENT OUTPUT	TOTAL SEDIMENT	MEAN HOURLY SEDIMENT OUTPUT	5 PERCENTAGE OF TOTAL SEASONAL OUTPUT
July	5-7/7	1696	179	776	2651	66.2	8.6
July	15/7	757	51	791	1599	66.6	
	16/7	567	89	222	878	36.6	
	17/7	1135	64	781	1980	82.5	
	18/7	2648	166	3166	5980	249.2	
	19/7	946	38	133	1117	58.8	
	15-19/7	6053	408	5093	11554	100.5	22.3
August	24-25/8	1892	77	3301	5270	114.6	10.2

Notes

1. All values in metric tonnes
2. Gravel output determined from purging of gravel trap
3. Sand output determined from purging of sand trap
4. Suspended sediment concentration refers to concentration at the end of the sand trap (Site D)
5. Percentage of the total seasonal sediment load which was removed during the period of flood discharge.

event. The July 1987 event had the highest yield. It is clear that all three events were very important for sediment transport and represent an important part of the seasonal sediment yield. The July flood contributed 8.65% of the total seasonal load in 2 days and the two 1987 events between them contributed 32.5% of the total seasonal load. The proportions of the load also varied. For example during the 4 days of the July 1987 flood coarse sediment (sand and gravel) and suspended sediment varied considerably with the greatest transport rates occurring during the initial overflow event and additional very high sediment transport when the main outburst occurred on July 18th. The August 1987 event also yielded a tremendous amount of suspended sediment, although the flow was directed over the glacier surface. This is interesting because the geomorphic evidence suggests that the effectiveness of this event concentrated at the glacier surface, but the sediment data imply a significant flushing of subglacial sediment, since this is the dominant source of fine sediment. Comparison between the 3 floods is difficult because they were so varied. However, an index of mean hourly sediment transport rate for each flood gives some idea of their relative importance. In this respect the most important is the August 1987 flood which transported 114.6 tonnes h^{-1}. The importance of outburst events is not to be under-rated by referring to mean hourly transport rates because sediment transport during the short-lived outburst on July 18th was very high (249.2 tonnes h^{-1}).

The recovery or response of the proglacial fluvial system during the aftermath of flood was hypothesised in the introduction (Figure 8.1). This simple model can be tested using data collected during the 1986 season. Figure 8.10 shows sediment transport fluctuations in relation to channel braiding, sinuosity and slope. Observations fit the model reasonably well, the only anomalies are a small secondary spike of braiding

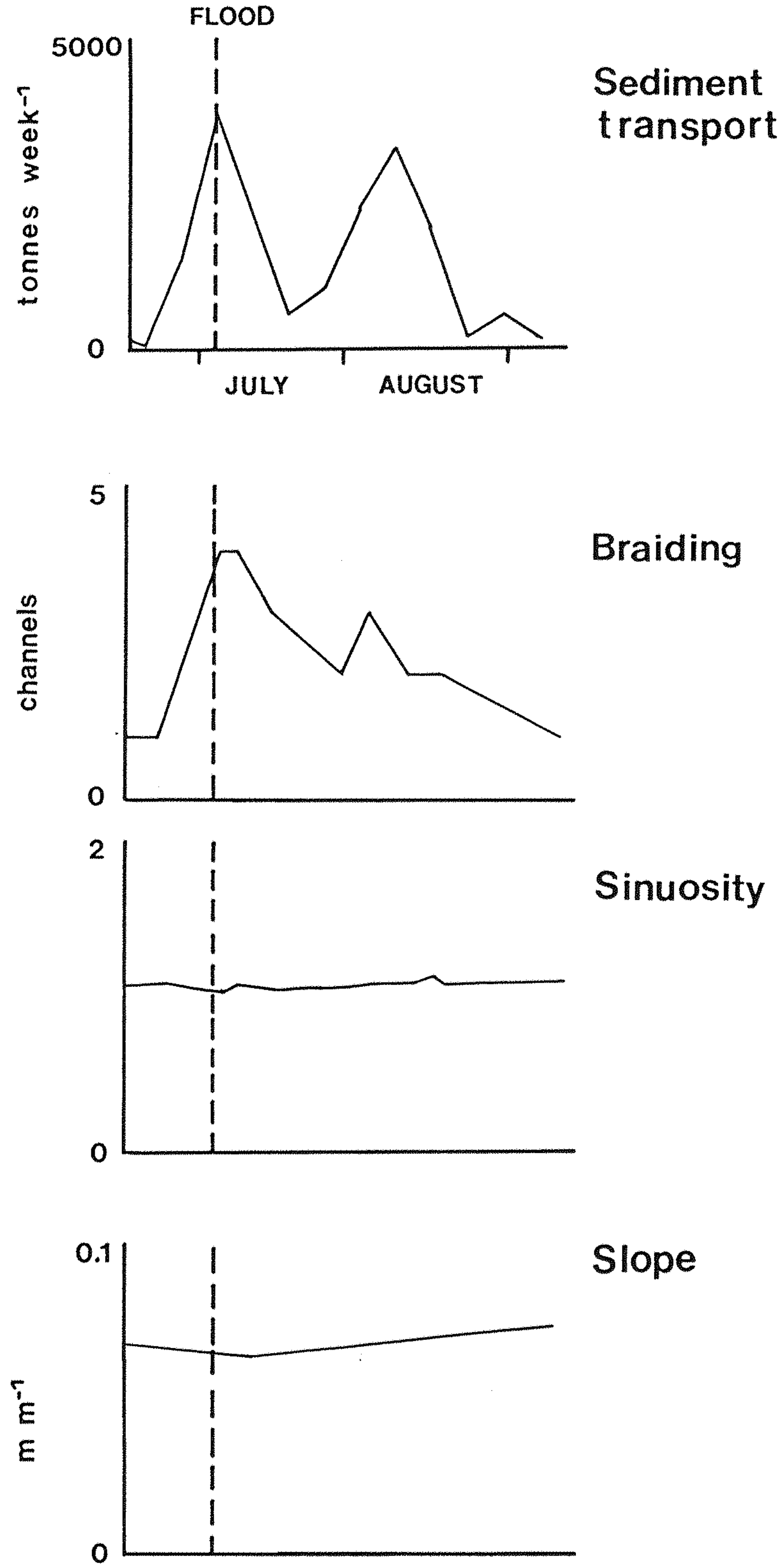


Figure 8.10 Response of the Bas Glacier d'Arolla proglacial fluvial sediment system to the July 1986 flood event.

associated with seasonally high sediment discharge in mid-August and the lack of response in sinuosity to the flood. The invariance in sinuosity is probably a function of confinement by valley train width. This suggests that the proglacial channel system responds rapidly to the flood event and shows a very short recovery period within the same ablation season. This dynamic response of the proglacial system was discussed in greater detail in Chapter 6 where it was shown that sediment transport (from the proglacial valley train) during the July 1986 and July 1987 floods yielded 1975 tonnes and 4794 tonnnes respectively.

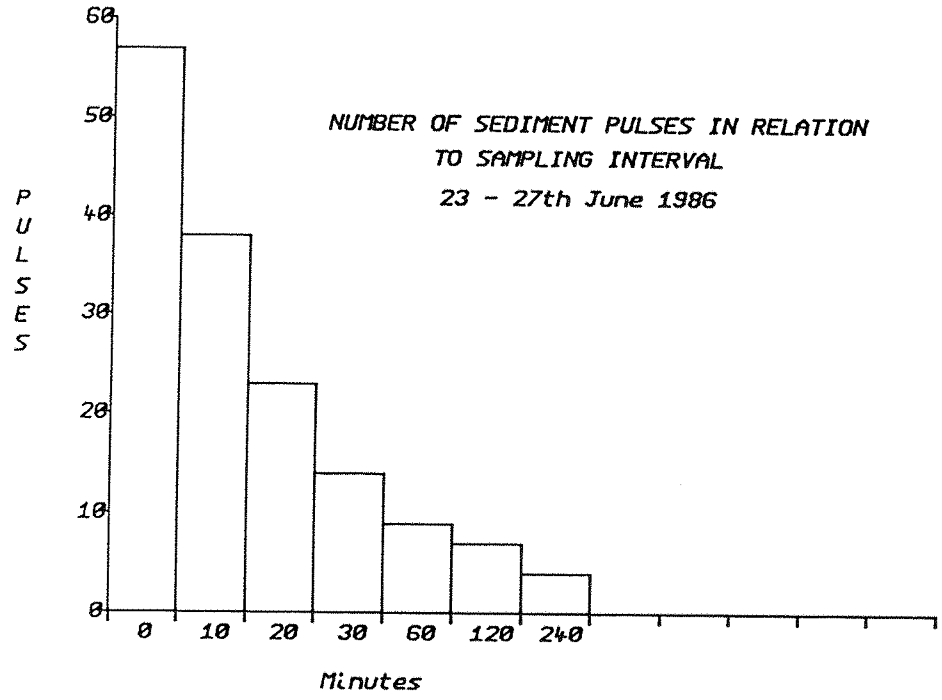
8.3 Suspended sediment pulses

In comparison with meltwater floods, pulses in suspended sediment concentration represent the opposite extreme of the magnitude / frequency spectrum. These low-magnitude high-frequency non-periodic pulses in suspended sediment are a characteristic of many proglacial stream suspended sediment series (Fenn, 1983). The term 'sediment pulse' suggests a rapid short-lived burst in suspended sediment concentration. Sources of such bursts, and accompanying discharge spikes, have been linked with: tapping of new subglacial sediment sources; rapid changes in glacial motion; purging of sediment traps from tributaries high in the catchment; failure of channel banks; failure of valley train bluffs; channel erosion; inputs from tributaries and hillslopes. Identifying these pulse events is important in the design of sampling strategies. For example Figure 8.11 shows results of a simulated sampling experiment on two continuous suspended sediment series June 23 - 27th 1986 and July 25th to September 11th 1986 from the Bas Glacier d'Arolla. Results indicate that considerable detail will be lost if a continuous record of suspended sediment is not monitored. Adopting a hourly sampling strategy on the June series will result in a loss of detail of 84% in the number of pulses identified. For the July series the effect is not as pronounced and the loss of detail is only 64%.

This section attempts to isolate the main controlling factors governing the form and origin of these pulses. Investigations were undertaken to determine:

1) The nature of sediment pulse transport in the proglacial channel.

2) Whether the form of the sediment pulse was diagnostic of the source.



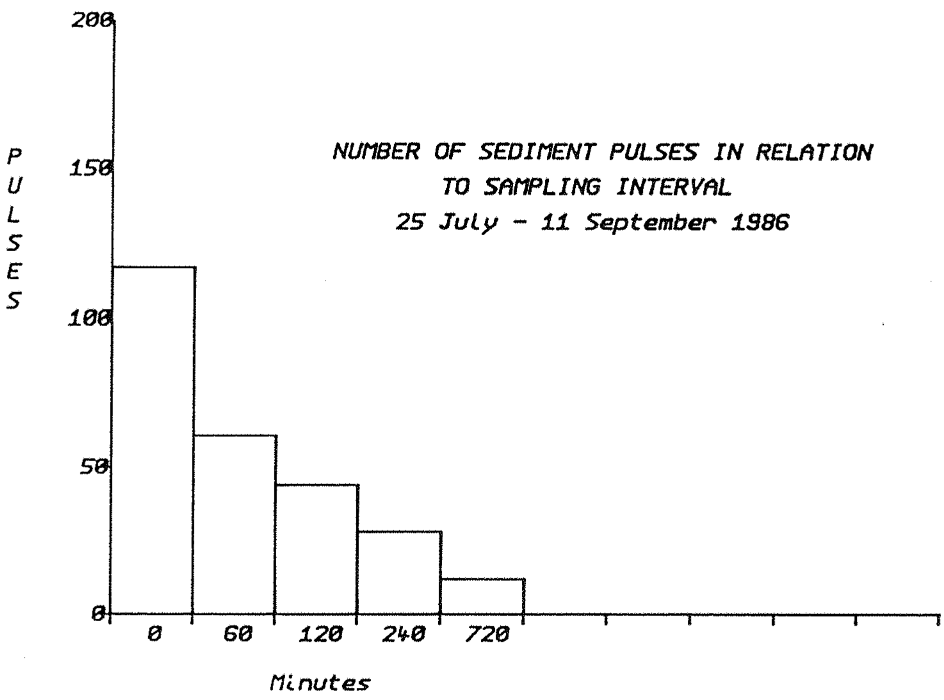


Figure 8.11 Simulated sampling experiments on two continuous suspended sediment series, July 23rd - 27th and July 25th - September 11th 1986, from the Bas Glacier d'Arolla proglacial stream.

8.3.1 Suspended sediment pulse experiments.

Experiments were carried out in order to determine the nature of sediment pulse transport in the proglacial channel of the Bas Arolla stream. A simple method for simulating sediment pulses involved the dumping of a known volume of sediment into the proglacial stream channel. Downstream of the introduction point, suspended sediment concentration was monitored using a Partech suspended solids meter. Sensors were not calibrated for actual loads, therefore results are presented as relative values of suspended sediment concentration. Care was taken to ensure that the same sensor, with constant settings, was used throughout the experiments and, to guard against confusion due to natural pulses, background concentrations were monitored continuously. Discharge was approximately stable for the duration of the experiments. Sediments were primarily collected from streambanks but some finer material was also collected for use in comparison with the passage of coarse sediment (bank materials) and fine sediment pulses.

Results demonstrate the basic transport characteristics of sediment pulses. Figure 8.12 shows that sediment introduced at progressively greater distances from the monitoring station showed an increased time to peak and a flattening (reduction in peak) of the pulse curve. Except for pulses measured over small distances (e.g. 10m) these pulses are characterised by approximately symmetrical rising and falling limbs. Varying the volume of sediment introduced over a fixed distance produced interesting results (Figure 8.13a). At the Haut Arolla site, as expected, the size of the pulse increased the greater the volume of sediment added. At the Lower Arolla site, over a shorter distance of 10m, sediment pulses were very similar except that when the 4 bucket volume was used the response was 3 times greater than the other pulses. This suggests that over such a

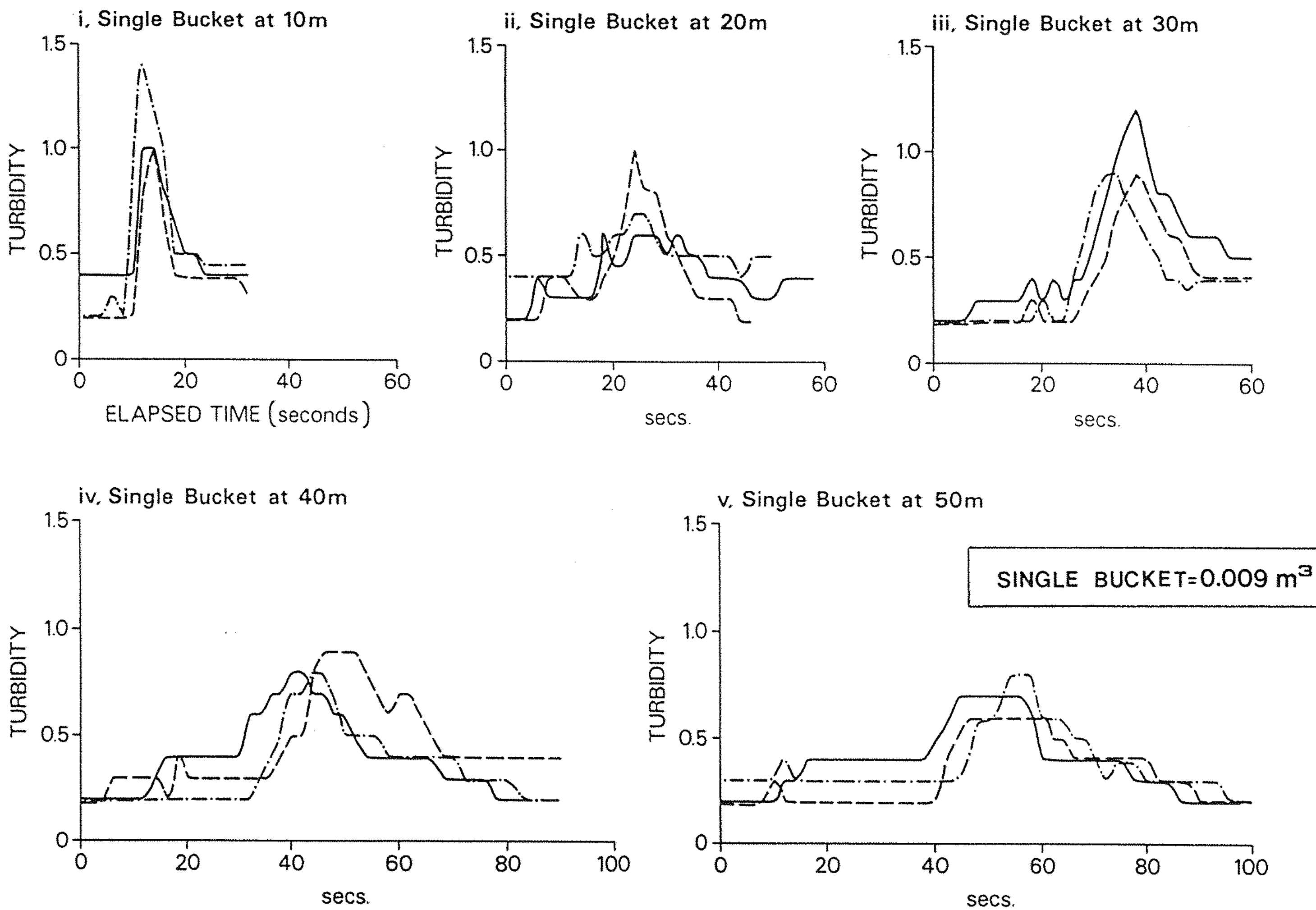


Figure 8.12 Suspended sediment pulse experiment showing progressive reduction in peak magnitude over distance.

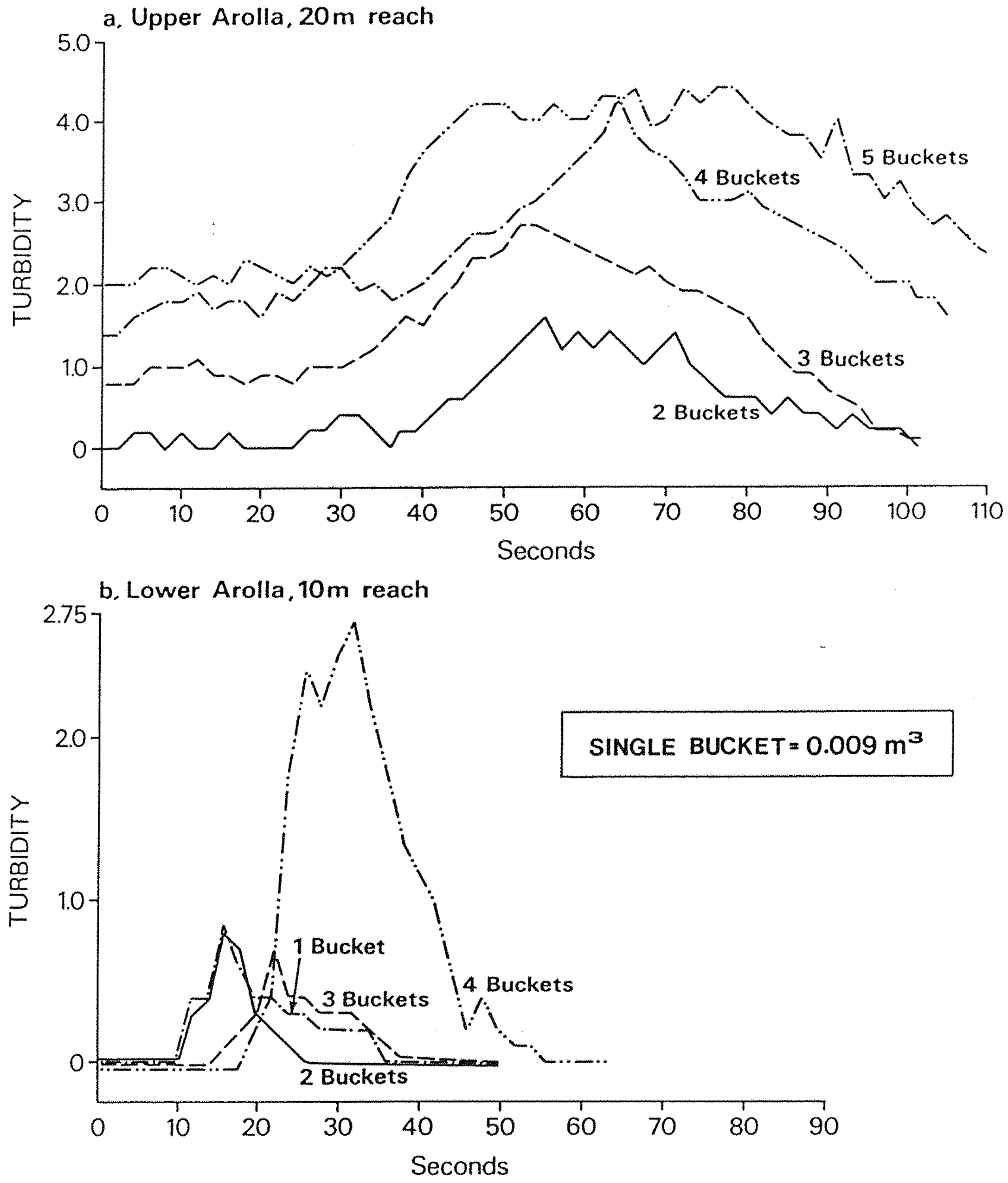


Figure 8.13 Suspended sediment pulse experiment showing the effect of varying the volume of sediment introduced into the channel.

short distance inadequate mixing occurred (Figure 8.13b). The nature of the pulse also varied with sediment size: in general fine sediment pulses were larger and recorded slightly earlier (Figure 8.14). Important variations did however exist and for apparently similar sediment (same source) a two-fold difference in the size of the pulse was recorded and in one instance a coarse sediment input produced a greater turbidity peak than a fine input. Comparison between sediment and solute (salt) showed the salt pulses to be recorded in advance of the sediment (Figure 8.15b). In addition, the solute curve was smooth as opposed to the erratic nature of suspended of the suspended sediment pulses. This probably reflects differences in the nature of transport: solutes are rapidly mixed and are less susceptible to hydro-dynamic perturbations during transit down the channel. However in one example (Figure 8.15b) the solute peak arrived after the suspended sediment pulse, which reflects the highly localised nature of sediment and solute flows in high roughness proglacial channels. In the last set of experiments (Figure 8.15a), sediment streaming in the main channel was demonstrated. At the Lower Arolla site sediment introduced at different locations in the channel, 40 m upstream of the recorder, produced marked differences in the magnitude and timing of pulses. Similarly, when sediment was introduced at both sides of the channel at once, a multi-peak response was recorded downstream (Figure 8.15a). However rather than receiving the anticipated 2 peaks, 3 peaks were recorded. The presence of a third peak may reflect a combination of the 'tails' of the two main pulses, which produce an apparent 3rd 'mixed' pulse. Pooling influences may also account for this, because sediment can be delayed behind obstacles and in channel eddies.

In conclusion, these experiments have demonstrated the variability of suspended sediment transport in proglacial channels. Travel distance, variations in sediment volume, sediment size characteristics and

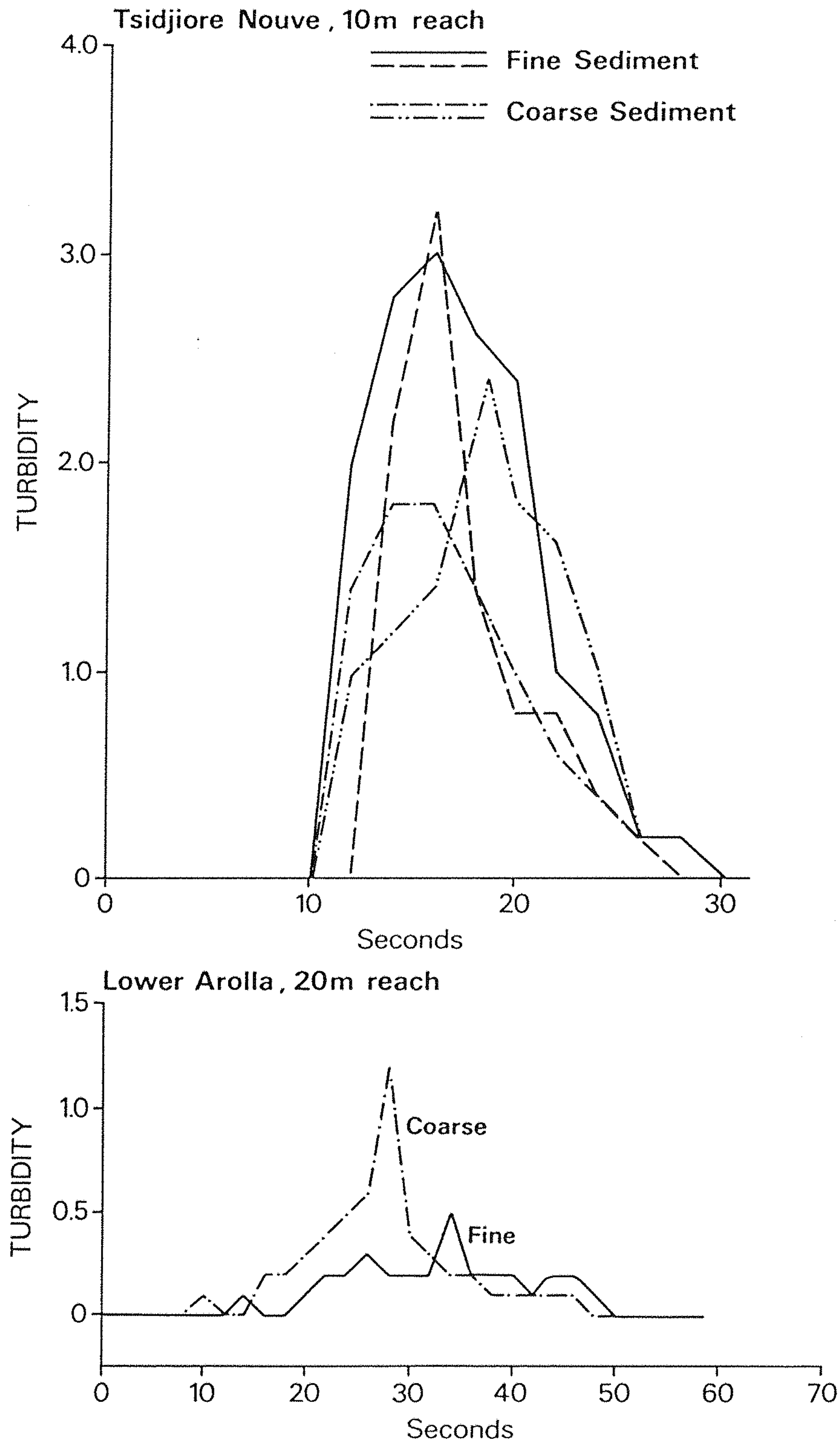


Figure 8.14 Sediment size effects and the generation of suspended sediment pulses.

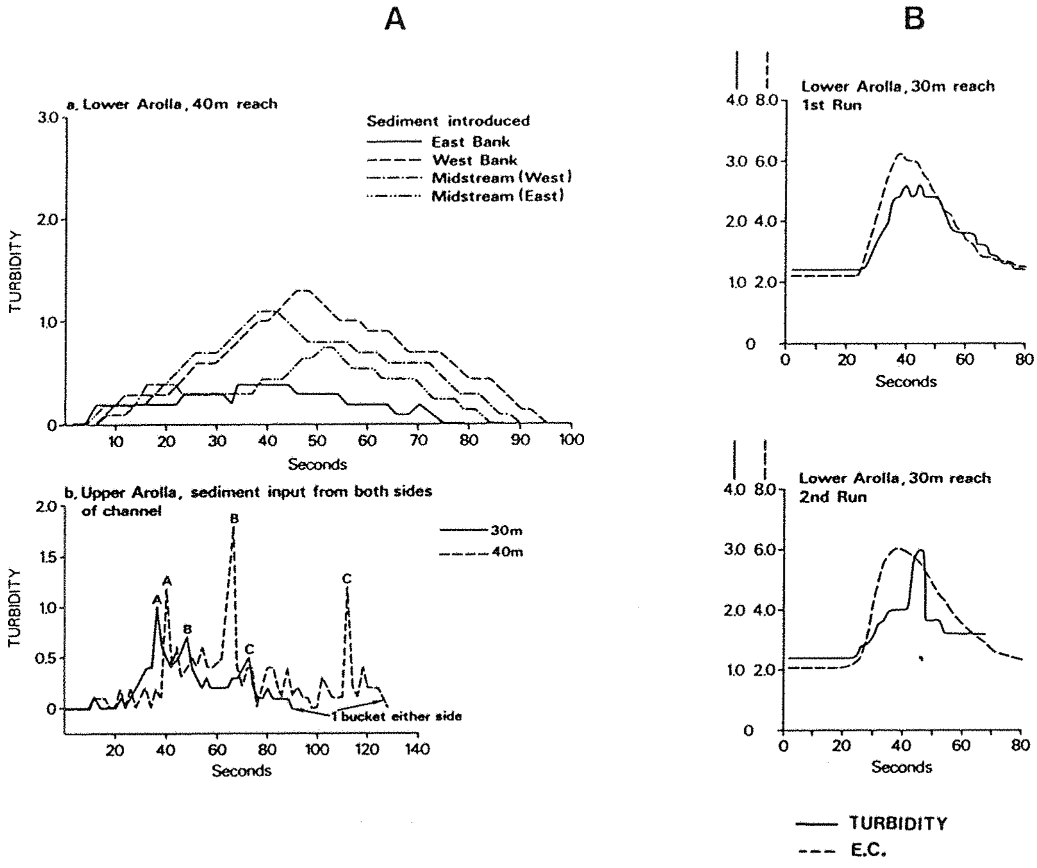


Figure 8.15 (A) Variations in suspended sediment pulses depending on the introduction point of the sediment upstream. (B) Comparison between salt and sediment pulse dynamics.

channel morphology all influence the magnitude, form and transport pathway of the suspended sediment. The premature arrival of solutes before sediment pulses is important since salt tracers are used to determine suspended sediment transport travel times. Experimentally some of the control was lacking due to local background fluctuations in suspended sediment concentration and discharge, variations in composition of sediment samples and inconsistencies in introducing sediment in the channel. In introducing the sediment into the channel the simulation was not accurate because sediment was delivered directly to the centre of the flow as a point injection whereas streambank collapse rarely occurs in this fashion. Scale is also important, because the study reaches were short (less than 50 m), sediment volumes were small (less than 70 kg) and the experiments were carried out over a restricted discharge range (0.6 - 1.0 $m^3 s^{-1}$).

8.3.2 The origin of sediment pulses

In Chapter 4 it was suggested that some of the large asymmetrical sediment pulses observed in the turbidity record were linked with purges from the Upper Arolla proglacial stream. This can be now be tested using the observations of suspended sediment concentration shown in Figure 8.16 for the 20th August 1986. On this day a heavy plant excavator (Swiss JCB) was employed to carry out channellisation works on the lower reach of the Bas Arolla proglacial stream just above the sediment trap. Measurements during the period of excavations showed that for the majority of the time there was little 'noise' in the record except when the excavations directly disturbed the channel bed and banks. Under these circumstances two sediment spikes were produced (Figure 8.16). Immediately following this disturbance, a large asymmetrical sediment pulse was observed moving downstream from the glacier (Figure 8.16) The contrast between channel pulses (perhaps indicative of bank

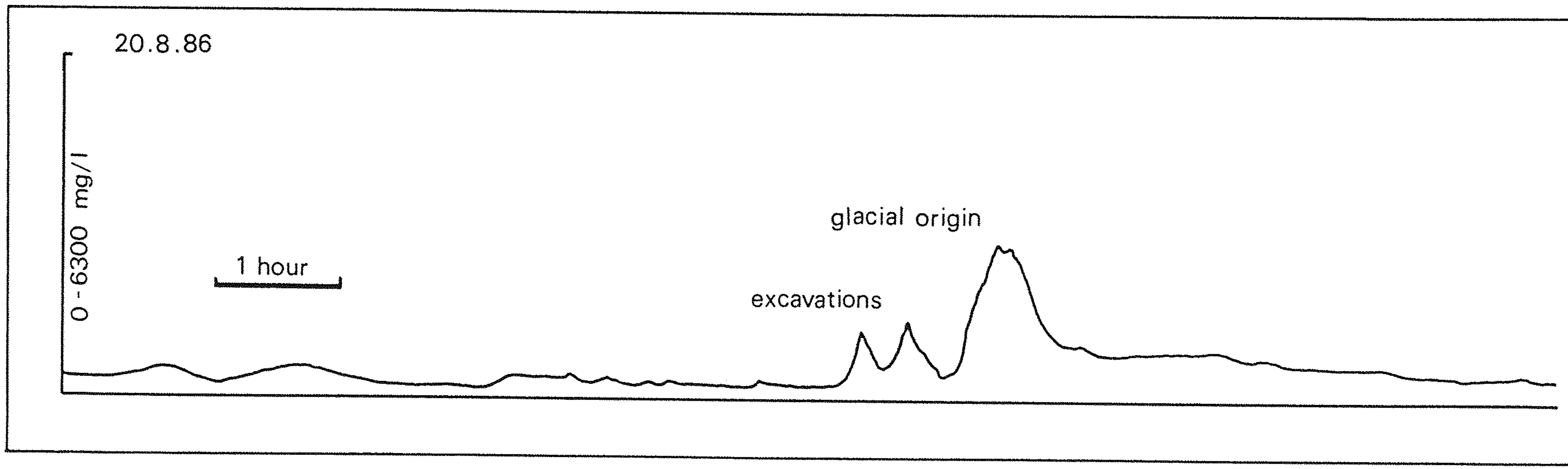


Figure 8.16 Comparison between suspended sediment pulses from the glacier and the proglacial zone. Pulses from the proglacial zone were generated by the action of an excavator (Swiss JCB) working on the Bas Arolla proglacial stream channel.

collapse) and glacial pulses can be clearly seen. Glacial pulses tend to be large and asymmetrical with a steep rising limb and a gradual falling limb whilst proglacial pulses are more likely to be smaller and symmetrical as shown in the sediment pulse experiments.

Figures 8.17 and 8.18 show 24 hour segments of turbidity trace in August and September which were characterised by a series of suspended sediment pulses. The times of purging of the Haut Arolla intake are plotted on Figures 8.17 and 8.18 and reveal close correspondence with the sediment pulses. Virtually every pulse or pulse complex is linked with a purge. This relationship is intriguing since it suggests that material purged from the upper glaciated catchments is very rapidly routed through the Bas Glacier d'Arolla. The characteristic form of these pulses consists of a very steep rising limb and a more gradual recession curve (Figure 8.17 A). Some pulses show more complex multi-peak forms (Figure 8.17 B). A number of pulses are unrelated to Haut Arolla purges (Figure 8.17 C). The magnitude of the pulses also vary considerably (compare pulses D and E in Figure 8.18). Therefore, pulse properties vary but all the pulses have a characteristic steeply rising limb.

In order to test this relationship further, cross association analysis (Davis, 1973) between the Haut Arolla purge series and the presence/absence of purges in the Bas Arolla suspended sediment series shows the best match between the two series at lag 0 with the second best match at lag 1. This supports the hypothesis that the release of purged sediment and water generates suspended sediment pulses on the Bas Arolla proglacial stream (Figures 8.17 and 8.18).

Analysis of the full suspended sediment series for 1987 identified a total of 256 pulses, 91 of which had a marked asymmetrical form like the pulses shown in Figures 8.17 and 8.18. Of these 91 pulses, 85% were matched

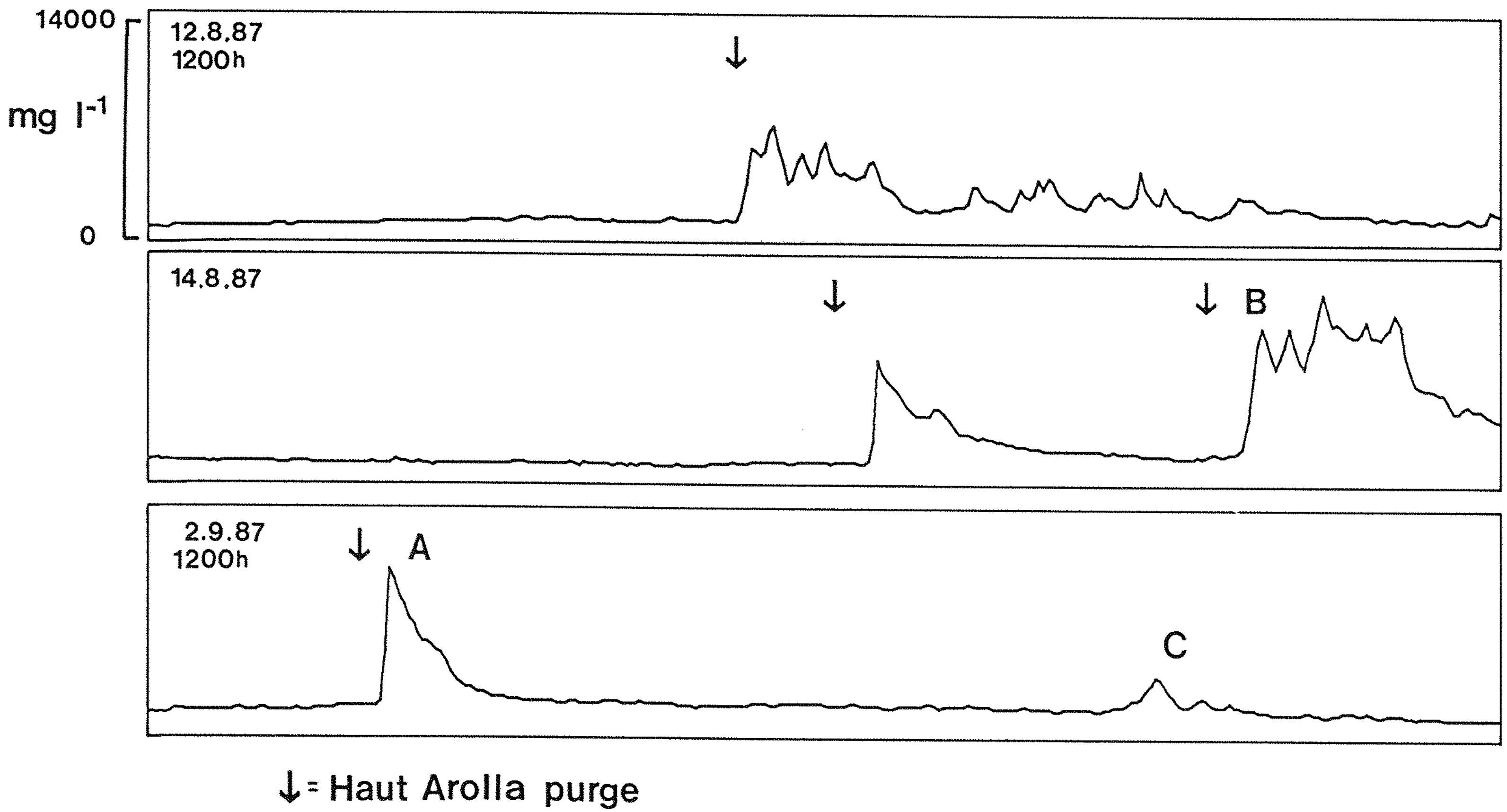


Figure 8.17 Segments of suspended sediment series showing sediment pulse events in the Bas Glacier d'Arolla proglacial stream and their correspondence with Haut Arolla purging. (Each segment is 24 hours long and starts at 00.00h unless stated otherwise).

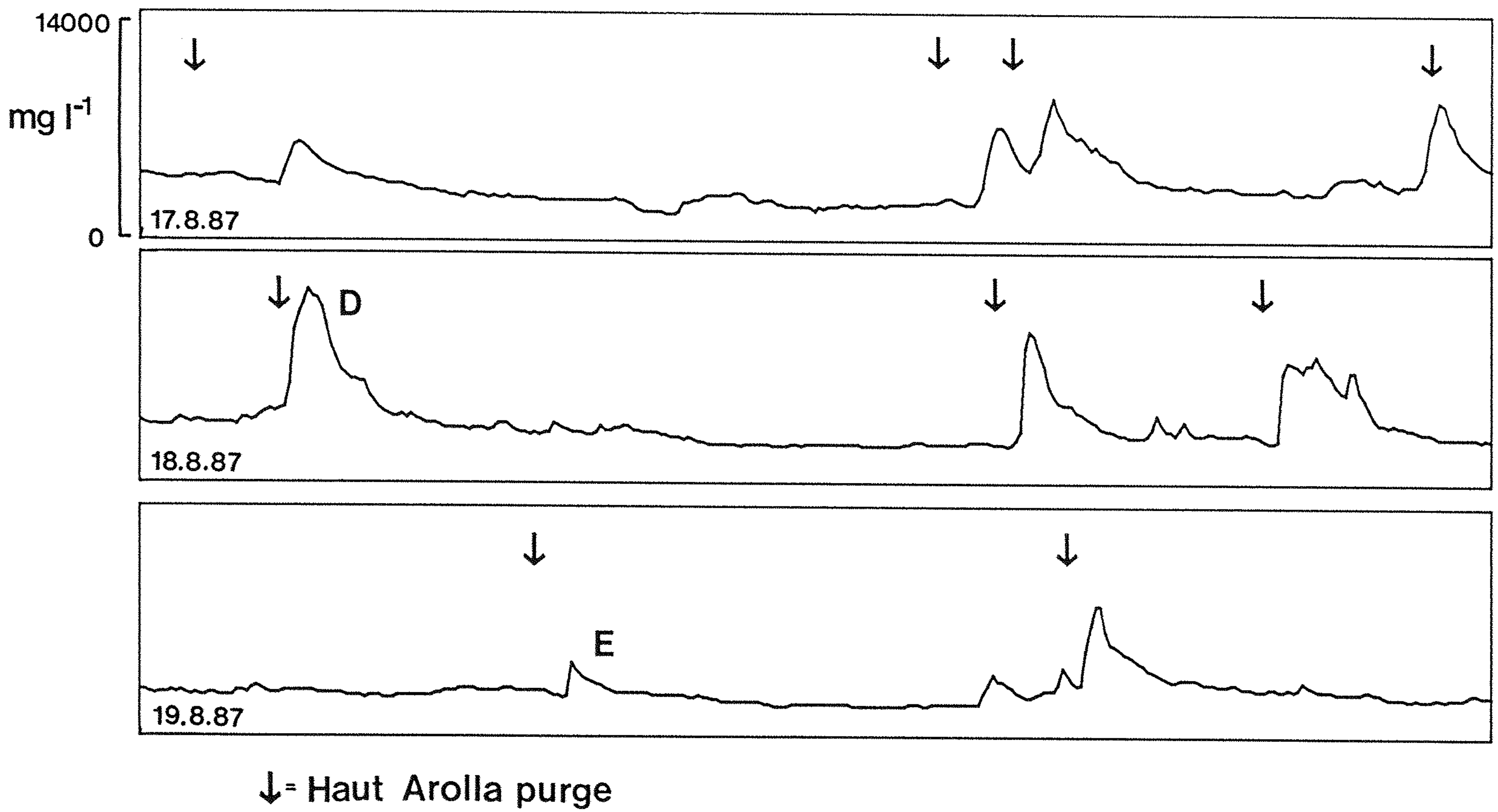


Figure 8.18 Segments of suspended sediment series showing sediment pulse events in the Bas Glacier d'Arolla proglacial stream and their correspondence with Haut Arolla purging. (Each segment is 24 hours long starting at 00.00h).

with Haut Arolla purges. Only 13 purges did not produce a sediment pulse on the Bas Arolla suspended sediment trace. Before 1st July Haut Arolla purges had no effect on the Bas Arolla suspended sediment series. This was because most of the purged debris from the Haut Arolla was discharged over snow and so did not enter the Bas Arolla drainage system. Out of the 13 purges that did not produce a sediment pulse, 5 were thought to be of this type. The remaining 8 purges could have entered the Bas Arolla glacio-hydrological network but due to temporary blockage in the internal drainage had no discernible effect on the suspended sediment series.

Travel times for the pulses varied (Figure 8.19) between 11 and 50 minutes. No systematic pattern in variation occurred over the melt season, which suggests that the internal drainage of the Bas Glacier d' Arolla is relatively open from the start of July onwards, although during July travel times tended to be longer than in August. The wide distribution of travel times probably indicates temporary changes in the subglacial passageways which become blocked by sediment or may be only active at certain discharges. A comparison of the pulse travel times (Figure 8.19) (between Haut Arolla purge and the recognition in the Bas Arolla suspended sediment series) and the travel time for water routed through a hypothetical tube-like conduit running the full length of the glacier (Humphrey et al., 1986) indicates that the travel time for a water pulse is 17-18 minutes (flow velocity 1.014 m s^{-1}, calculated from the Gauckler-Manning-Stickler formula, Rothlisberger, 1972) which is in agreement with the distribution of travel times shown in Figure 8.19. The majority of travel times ranged from 10 - 50 minutes suggesting a very 'open' glacial drainage system with little delay in the transmission of water and sediment.

Care should be taken in interpreting the origin of the sediment in the pulses because the water/sediment mix

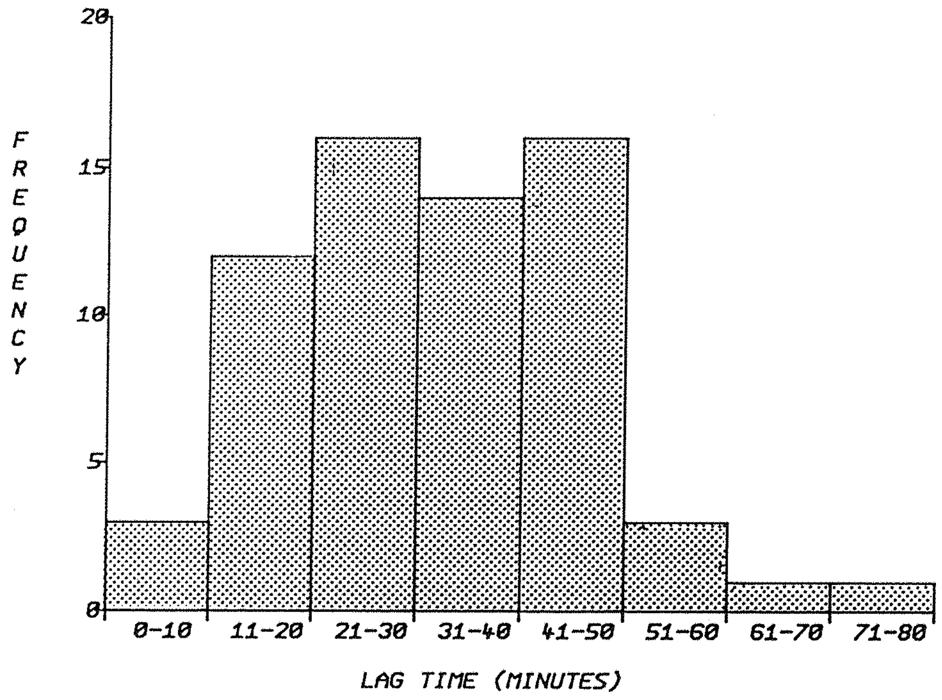


Figure 8.19 Travel times of suspended sediment pulses between the Haut Arolla sediment traps and the Bas Glacier d'Arolla proglacial stream.

purged from the Haut Arolla may not necessarily be transmitted through the Bas Arolla glacier together sediment may be deposited inside the glacier and the water pulse continues down glacier. If this was the case sediment would be entrained from the rock flour store at the glacier bed before the pulse re-emerged in the lower proglacial stream. Sediment tracing studies would be useful in answering the question.

Although artificial pulses are characteristic of many of the suspended sediment flushes from the Bas Glacier d'Arolla natural pulses also occur (e.g. Gurnell and Fenn, 1985). These natural pulses or flushes have been attributed to tapping of sediment sources within the glacier system and from inputs of proglacial material to the meltwater stream (Bogen, 1980; Hammer and Smith, 1983; Richards, 1984; and Gurnell, 1987). At Bas Arolla few of the sediment pulses are proglacial in origin, the vast majority are flushes from the glacier. A preliminary calculation suggests 1903 tonnes of sediment are contributed in 39 suspended sediment pulses, generated by purging of the Haut Arolla meltwater intake during the 1987 sediment budget period.

8.4 Summary

Meltwater floods are characteristically high magnitude low frequency events. Three events of this kind occurred in the Bas Arolla basin in July 1986 and July and August 1987. The 1986 event was the least effective of the floods, it did the least work and only affected the immediate proglacial zone. The July 1987 flood consisted of a series of events involving an overflow event on the 15th followed by an outburst on the 18th. The outburst was particularly important for sediment transport but overall the effectiveness of the flood was confined to glacial and channel areas. The August 1987 flood was an unusual event in that its origins were partly artificial in that a substantial proportion of the discharge was released from the hydro-electric scheme. Nevertheless the impact of the flood caused greatest disturbance to the landscape, mobilised the greatest proportion of sediment stores and produced the greatest hourly sediment transport rate. The events are very important for the evolution of the proglacial landscape (Johnson and Power, 1985) with the two events in 1987 contributing about one third of the seasonal sediment yield with a large proportion derived from the proglacial valley train. For example, in the 1987 sediment budget period (May to July) 22% of the total sediment yield was contributed from the valley train during the meltwater outburst flood July 15th to 18th (Chapter 9).

Recovery following the July 1986 flood was gauged by assessing the change in proglacial channel parameters following the event (Figure 8.10). Based on the trend in these data recovery seems to be complete by the end of the ablation season. Therefore for this event recovery was rapid and the 'memory' of the geomorphic system relatively short.

For hydro-electric companies these events are important because localized erosion around meltwater intakes, channel change and the routing of excess flood-waters all pose management problems. These conclusions only apply to the magnitude of the observed events (7-20 $m^3 s^{-1}$). Larger floods may act very differently, so it is best to 'expect the unexpected' (Young, 1980).

In addition to these high magnitude low frequency events, proglacial stream suspended sediment series are characterised by frequent short bursts or pulses of concentration. Small-scale experiments demonstrate the variability in suspended sediment transport in proglacial channels. Sediment pulses follow a predictable distance-decay trend down channel and such trends are consistent over a range of pulse sizes. Observations of bank disturbance and the concurrent passage of a sediment pulse from the glacier suggest that it is unlikely that large scale pulses originate from bank collapse. Correspondence between Haut Arolla sediment trap purging and the occurrence of large scale sediment pulses in the Bas Arolla suspended sediment series suggested a causal link between the two events. Based on the travel time of these pulses the Bas Arolla glacial drainage network appears to be very efficient in transmitting pulses and therefore probably consists of a large conduit system (Rothlisberger, 1972).

Analysis of both high and low magnitude sediment transport events provides an interesting insight into the glacio-hydrological system of the Bas Arolla glacier. This insight albeit cursory is useful in suggesting linkages between the proglacial and glacial sediment and water systems.

Chapter 9

CONSTRUCTION AND APPRAISAL OF A PROGLACIAL FLUVIAL SEDIMENT BUDGET.

9.1 Introduction

In the preceding chapters a variety of sediment monitoring techniques have been used to investigate various components of the proglacial fluvial sediment system. This chapter attempts to integrate these components into a sediment budget framework.

In this study sediment budget is defined as the quantitative description of sediment movement through the proglacial fluvial system. This involves the measurement of fluvial sediment transport and the exchange of sediment, sometimes via non-fluvial processes, with fluvial sediment stores. In glacierised catchments and mountain catchments as a whole measurements of this kind are severely lacking. Exceptions include the 'base-line' fluvial sediment transport monitoring programmes carried out on Norwegian proglacial streams (Østrem, 1975; Kjeldsen, 1981) and a series of less sustained case studies. This is not to say these case studies have not met with success. Many of them have made valuable contributions to the study of proglacial fluvial sediment transport e.g. Fahnestock (1963), Østrem (1975), Hammer and Smith (1983) and Gurnell and Fenn (1984). Of these studies Hammer and Smith (1983) provide the most explicit attempt to assess the relationship between proglacial fluvial sediment transport and glacio-fluvial sediment supply. It was concluded that bedload and suspended load were transported in approximately equal amounts. Dissolved loads were minor (less than 3%). Variation in local sediment supply produced scatter in the rating curves.

Analysis of fluvial sediment supply revealed that for suspended load, 6% was derived from supraglacial and high englacial debris, and 47% was derived from each of subglacial and channel bank sources. For bedload, 46% was from supraglacial and high englacial sources and 27% each was derived from subglacial and bank sources. Although useful, these computations are based on matching of sediment source characteristics with measurements of suspended sediment transport and so they must be considered tentative.

What this study does suggest is that an important contribution to the load of the meltwater stream was derived from proglacial sources. This is important because 'baseline' fluvial sediment transport measurements have been used as a first approximation of both erosion and denudation rates or amounts (Walling and Webb, 1983). In glacierised basins measurements of this kind have also been used as an inferential tool in understanding processes acting at the glacier bed (Collins, 1979). Without due acknowledgement of the possible modification of meltwater stream sediment loads by processes acting in the immediate proglacial zone, such approximations and inferences are to be regarded with some scepticism. This is not to dismiss the usefulness of streamload sediment yields, because they provide a valuable integrated estimate of catchment-wide erosion rate estimate. It is particularly when the sediment study focus changes from quantifying outputs to examining internal sediment fluxes that total streamload data become a questionable basis for inference (Trimble, 1981).

This study is concerned with fluvial sediment transfer processes in the proglacial zone of the Bas Glacier d'Arolla, and may appear comparable to the study of the Hilda Glacier, Alberta by Hammer and Smith (1983). In practice, this comparison is invalid because where Hammer and Smith only estimated contributions from sources by indirect means this study has measured erosion rates directly and has quantified rates of sediment transport and the rates of sediment addition to storage sites. In these respects this study has fulfilled the requirements of a sediment budget (Dietrich et al., 1982).

9.2 The sediment budget

Figure 1.2 (Chapter 1) provides an outline of a general sediment budget model for an alpine glacierised catchment and defines the proglacial fluvial sediment transfer sub-model. This section refines this sub-model and provides estimates of material transfers between the components of the model (Table 9.1).

Table 9.1 and Figure 9.1 provide a summary statement of the sediment budget for fluvial sediment transfer processes in the Bas Arolla proglacial zone for the period 25th May to 30th July 1987. Table 9.1 is divided into several sub-sets of processes: slopewash and splash; supply from hillslopes; material transfer in tributary streams; inputs from the glacier snout; and changes in sediment flux in the valley train. These elements define sediment fluxes within the sediment budget which can be evaluated against the material outputs from the basin (Table 9.1). Also included in Table 9.1 are estimates of sediment transport during particular events, namely the outburst flood of July 15th-18th 1987 and the cumulative effect of sediment pulses produced by purging of the Haut Arolla sediment trap (Chapter 8).

All material transfers shown in Table 9.1 refer to the total sediment moved (in tonnes) by the observed event (in the case of a slushflow for example) or the cumulative total sediment moved over the measurement period in which the individual material transfer processes were acting (e.g. the 67 days of valley train bluff erosion). Percentage contributions from each material transfer process are than calculated as a percentage of the total basin sediment output. Duration or action time (Table 9.1) refers to the number of days or fraction of a day over which each material transfer process was active. Errors associated with these

Table 9.1 Summary of Bas Glacier D'Arolla proglacial fluvial sediment budget, May to July 1987.

PROCESS	SEDIMENT SUPPLY (TONNES)	% OF TOTAL OUTPUT	DURATION ACTION TIME (DAYS)	WORK (JOULES)	POWER (WATTS)	SOURCE OF ERROR (%)
Supply to tributaries						
Slope wash and splash	0.076	0.00004	30	6.76×10^{-6}	2.6×10^{-12}	a (120)
Supply from slopes						
Supra-nival streamflow	0.39	0.002	0.003	7.65×10^{1}	2.55×10^{-3}	b
Rockfall	6.91	0.032	0.0001	1.05×10^{4}	1.05×10^{-3}	b
Slushflow	4.10	0.019	0.063	2.41×10^{3}	2.41×10^{-1}	b
SUB-TOTAL	11.4	0.053				
Tributary streams						
Bottle sampling	211.40	0.974	45			a b (71)
Traps and survey	8.75	0.04	45			a b
SUB-TOTAL	220.15	1.014		3.02×10^{5}	7.77×10^{-2}	
Snout zone						
Moraine deposition	265.9	1.23	67	1.57×10^{5}	2.71×10^{-2}	b
Rockfall	6.48	0.03	0.00006	4.45×10^{3}	8.90×10^{-2}	b
SUB-TOTAL	272.38	1.26				
Valley train						
Bluff erosion	452.2	2.09	67	1.77×10^{4}	3.06×10^{-3}	a b (22)
Channel erosion	4792.0	22.08	67	4.94×10^{5}	8.53×10^{-2}	a b
OUTPUTS						
Gravel and sand traps	14842	68.41	67			b
Suspended sediment (turbidity)	6598	30.41	67			b
SUB-TOTAL	21440	98.82		1.47×10^{7}	2.54	
Solutes	255	1.18	67	1.75×10^{5}	3.02×10^{-2}	a c
TOTAL	21695	100.00				
Outburst event						
Gravel and sand traps	6461	29.78	4.79			a b
Suspended sediment (turbidity)	5093	23.48	4.79			a b
SUB-TOTAL	11554	53.30		7.93×10^{6}	1.92×10^{1}	
Haut Arolla sediment pulses	1903	8.77	2.6	5.97×10^{6}	2.65×10^{1}	b

NOTES:

Source of error: a = statistical - sampling error
b = field measurement error
c = laboratory - analytical error

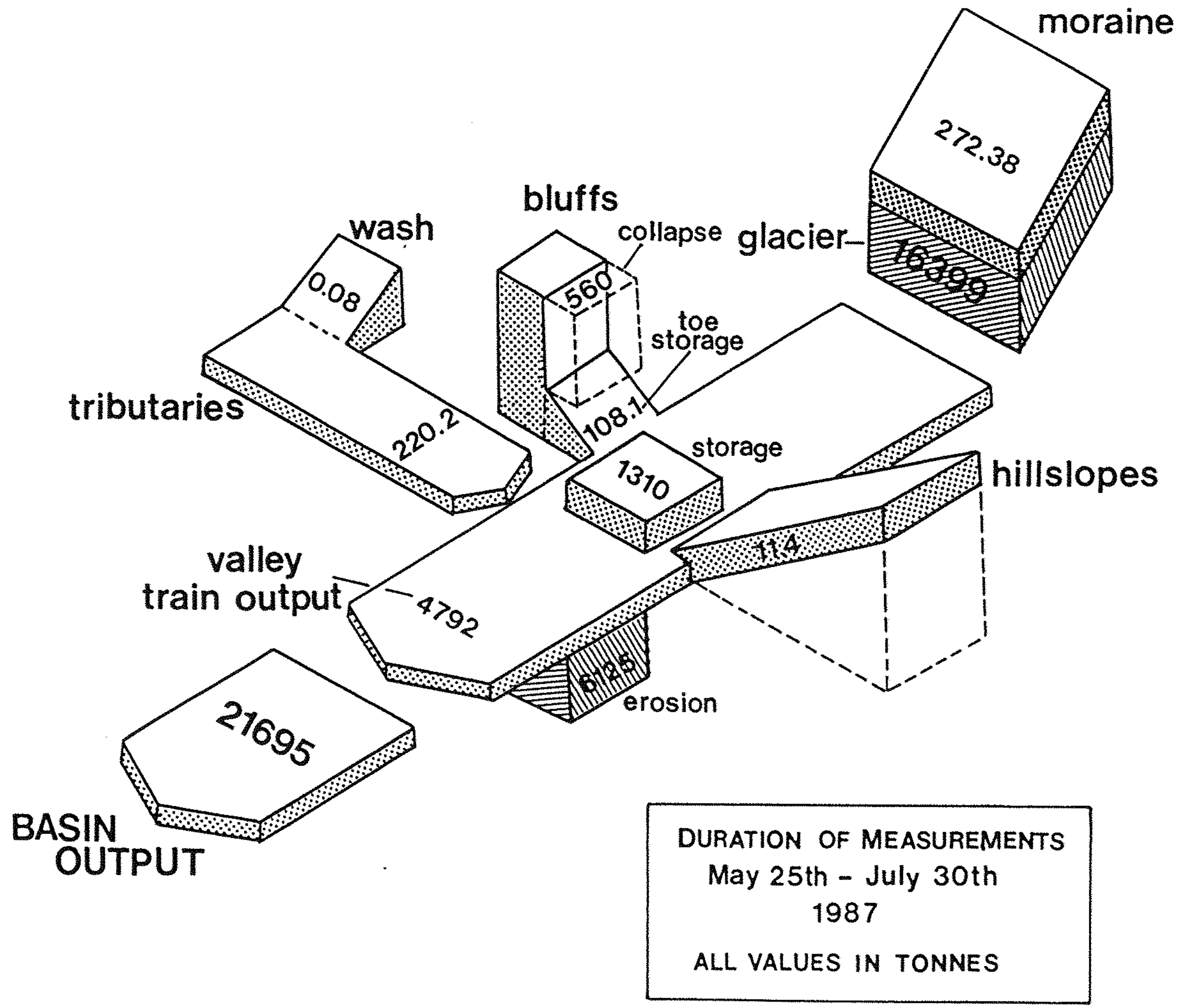


Figure 9.1 Summary of the Bas Glacier D'Arolla proglacial fluvial sediment budget, May - July, 1987.

estimates of sediment supply are difficult to quantify, so the last column in Table 9.1 indicates the likely source of the error. Three categories of error are distinguished: statistical - sampling error which can usually be expressed as a percentage error; field measurement error; and laboratory - analytical error.

The energetics (work and power) of these processes can also be estimated. In physical systems this is most suitably measured as work or power (Caine, 1976; Gregory, 1987):

WORK = force x distance (joule - J)

POWER = the rate of doing work (watt - $W = J\ s^{-1}$*)*

The capacity to perform work will be dependent on the the available potential energy in the environment and the rate at which this is translated into kinetic energy. This approach has considerable promise for integrating studies in physical geography (Gregory, 1987) and is readily applied to hillslope and fluvial processes (Caine, 1976). However, this has rarely been applied.

This section describes the components of the budget and the limitations of the approach. Components are discussed separately so that the strengths and weaknesses of inter-relationships, specified by the exchange of sediment within the budget, can be assessed. In this way an appreciation of the working linkages in the budget can be determined.

9.2.1 Components

Recalling the initial research questions from Chapter 1, two of the four questions set out related directly to the formulation of the sediment budget model. These were:

1) How is the sediment in a proglacial stream modified by contributions from sediment sources along the valley train and what are the proglacial fluvial sediment sources?

2) What are the major transport processes, storage elements and linkages in the proglacial fluvial sediment budget?

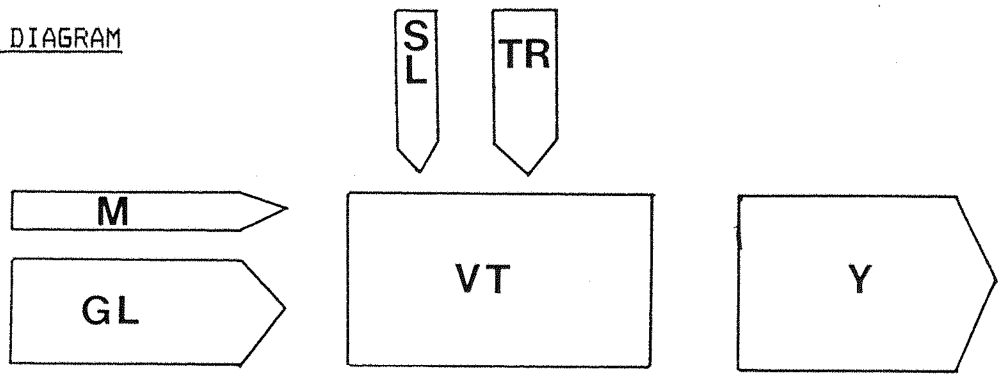
The first question relates to the relative proportions of load contributed from glacial and proglacial sources and can be calculated from the budget estimates in Table 9.1.

Figure 9.2 sets out the procedure for calculating the relative proportions of glacial and proglacial sediment contributions. The calculations are based on the basic sediment balance equation (Figure 9.2, (1)). The top section of Figure 9.2 defines the relationships between the components in the sediment balance equation. All components have been measured in this study apart from the glacial inputs. Given that this is the only unknown then it can be estimated by rearranging equation (1) to give equation (2). The glacial contribution can be added to the moraine contribution (to estimate the total glacial input (equation 3) and this can be subtracted from the total basin yield to estimate the contribution from proglacial sources (equation 4). These calculations indicate that that 23.2% of total output is derived from proglacial sediment sources. Therefore, the assumption that river reaches downstream from glaciers receive negligible contributions from proglacial sources (Bogen, 1980) is not valid in the case of the Bas Arolla proglacial zone under the conditions pertaining during the study period.

The detailed transfer processes and linkages within the budget (Figure 9.1) provide a insight into the

Figure 9.2 Sediment balance calculations for the Bas Glacier d'Arolla proglacial fluvial sediment budget, May - July, 1987.

DEFINITION DIAGRAM



Where: SL = slope inputs
M = moraine inputs
VT = valley train store
TR = tributary inputs
GL = glacial inputs
Y = yield

Sediment Balance Equation

$$Y = SL + TR + M + GL\ (+/- \triangle VT) \qquad (1)$$

Given that GL is the only unknown:

$$GL = Y - (SL + TR + M\ (+/- \triangle VT) \qquad (2)$$

Therefore, total glacial contribution:

$$= GL + M \qquad (3)$$

and the total contribution from proglacial sources is:

$$= Y - (GL + M) \qquad (4)$$

Calculations

The unmeasured contribution from the glacier is given by equation (2):

GL = 21695 - (11.4 + 220.15 + 272.38 + 4792)
= 16399 (76.8%)

Using equation (4) the proglacial contribution to load is:

= 21695 - (16399 + 272.38)
= 5024 (23.2%)

relationships between proglacial geomorphology and sediment transport. The most striking thing about valley train processes is the overwhelming importance of channel erosion and storage (Figure 9.1). Channel erosion accounted for 95% of the sediment supplied from proglacial sources. Of this amount just under 10% was derived from erosion of the valley train bluffs the remainder being attributable to channel erosion. This implies a deeply incised channel. However, this may only be a temporary phenomena because the budget may represent an early season 'cut' budget, whilst later in the season compensating 'fill' might dominate. This state can be partially explained by the recent July 15th to 18th flood and the 'incisive' nature of this event (Chapter 6). Supply from tributaries represented only 4.4% of proglacial sediment contributions with direct contribution from slope processes amounting to 0.5%. This suggests a poor connection between the hillslope and channel sediment transfer processes. The main addition to streamload in the proglacial environment was contributed from the valley train. As shown in Chapter 6 the majority of this sediment was supplied during the relative short duration of the July 15th-18th meltwater outburst flood. This emphasises the episodic nature of proglacial sediment supply. Indeed the outburst event was of major significance in the overall budget since it was responsible for 53.3% of the total material transported during the budget period. In addition the cumulative effect of small flushes of sediment caused by purging of the Haut Arolla sediment trap was important since it accounted for almost 9% of the total budget. The contribution from solutes was small (1.2%), but it represents long distance transport whilst coarse fraction components may well approximate to within-basin relocation of sediment.

A comparison of these estimates with those of Hammer and Smith (1983) indicates that in the Bas Arolla proglacial system during the study period the contribution from glacier sources was far greater. The importance of channel erosion as a source of sediment contrasts with the conclusion of Maizels (1978) who suggested, for the Bossons glacier Mont Blanc, that 16% of the fluvio-glacial debris eroded from the glacier was deposited in the valley train (i.e the proglacial zone was aggradational). These differences can be attributed to 'real' contrasts in catchment/historical aspects of the three proglacial systems, but should also be partly attributed to artifacts of the measurement techniques and sediment load expressions employed in each of the studies.

9.2.2 Limitations

The precision of a sediment budget depends on the efficiency with which material transfer processes are calculated. The accuracy of the budget is a broader concept which embodies the extent to which the budget is representative rather than just precise. At this level the limitations of the budget owe as much to concept, scale and system boundaries as to measurement technique. Because the proposed budget is constructed from an amalgam of process rate estimations which all show varying action times (often only a single occurrence - one event) errors are difficult to quantify. Generally, for continuous processes statistical errors in relation to spatial sampling or variation over time can be calculated (e.g. slopewash and bluff erosion, Table 9.1). However, if only a single event occurred (mainly slope events) during the period of observation, the major source of error is likely to be in the measurement of the event. The only other additional source of uncertainty in the erosion process estimates given in Table 9.1 stems from laboratory analytical errors (e.g. the determination of total dissolved solids in the calculation of solute transport). Based on this three-fold classification, for the processes in this sediment budget, the major sources of precision error are indicated in Table 9.1.

The greatest uncertainty in the data surrounds the estimation of hillslope processes and tributary sediment contributions. These processes are discussed in detail in chapter 5, but it is worth noting here that their contribution to the overall budget was so small that large errors in their calculation would probably had little effect on the overall structure of the budget. Errors in calculating moraine accumulation may also be significant because repeated survey in the moraine accumulation zone was not possible! However the estimates obtained compare favourably with other estimates of moraine accumulation in the Val d'Hérens (Small, 1987), and once again they are small enough (only 1.26% of the sediment supply over the budget period) to induce only very limited instability in the overall budget.

In addition to limitations in estimating transport processes, spatial and temporal design limitations are also inherent. It has been continually stressed that this is a *proglacial fluvial* sediment budget (with proglacial defined strictly with respect to distance from the ice front) and only involves the estimation of fluvial sediment transport processes and processes contributing sediment to fluvial sediment stores (channels). Therefore, the small contribution from slope processes does not suggest low levels of colluvial activity on hillslope: it just suggests that slopes are contributing relatively little sediment to fluvial sediment sources at the present time in this area.

Temporal constraints are also inherent in this approach because the budget has been constructed for a 67 day period from May 25th until July 31st, 1987. As has already been noted, this means that this is an early to mid season sediment budget and the processes acting in the first half of the season are unlikely to mirror

those later in the year (e.g. the sequence of bank change, Chapter 7). This is especially true in 1987 when a second large flood event on August 24th produced extensive erosion. Based on the geomorphic evidence following this event, the western slope of the Bas Arolla proglacial zone was substantially altered by slope movements, with the truncation of the toe of a large fan at the margin of the valley train suggesting a large contribution from slope sources to fluvial sediment yield (Chapter 8). This makes the point that there is intermittency in the proglacial fluvial sediment system. This intermittency may at first be triggered by the exceedance of a threshold (e.g. slope failure) but once exceeded, energy levels remain high (e.g. continued adjustments in a failed mass of sediment ensure that slope sources continue to deliver sediment to the valley train). Such an event may result in the 'Hurst effect' (Church 1980) which is a non-periodic grouping of similar values over a period of time.
The present sediment budget is thus of considerable inherent interest and significance, but cannot be readily be extrapolated to other locations or scales in time or space.

9.3 Energetics of budget processes

In the physical environment every process requires energy for its maintenance. In mountain environments the high relief and steep slopes which characterise these areas provide a considerable source of potential energy for erosion and sediment transport (Barsch and Caine, 1984). It therefore follows that energy fluxes in areas of high relief form a useful base for characterising the geomorphic systems of such areas and for comparing processes (Caine, 1976). Indeed the geomorphological application of the technique has its roots in mountain geomorphology, from the early applications of Jackli (1957) and Rapp (1960), who expressed work as the mass of sediment transported through a vertical distance in unit time, to Caine (1976) who elegantly expresses sediment movement in terms of physical work, with units in Joules. Surprisingly, this approach has been rarely applied and has found little support elsewhere in physical geography. However, if the prediction of Gregory (1987) is to be believed, a revolution, with physical geography based upon energetics, may be in sight.

This section attempts to discuss the sediment transport processes that constitute the sediment budget model (Section 9.2) in terms of magnitude and frequency and energy characteristics: work and power.

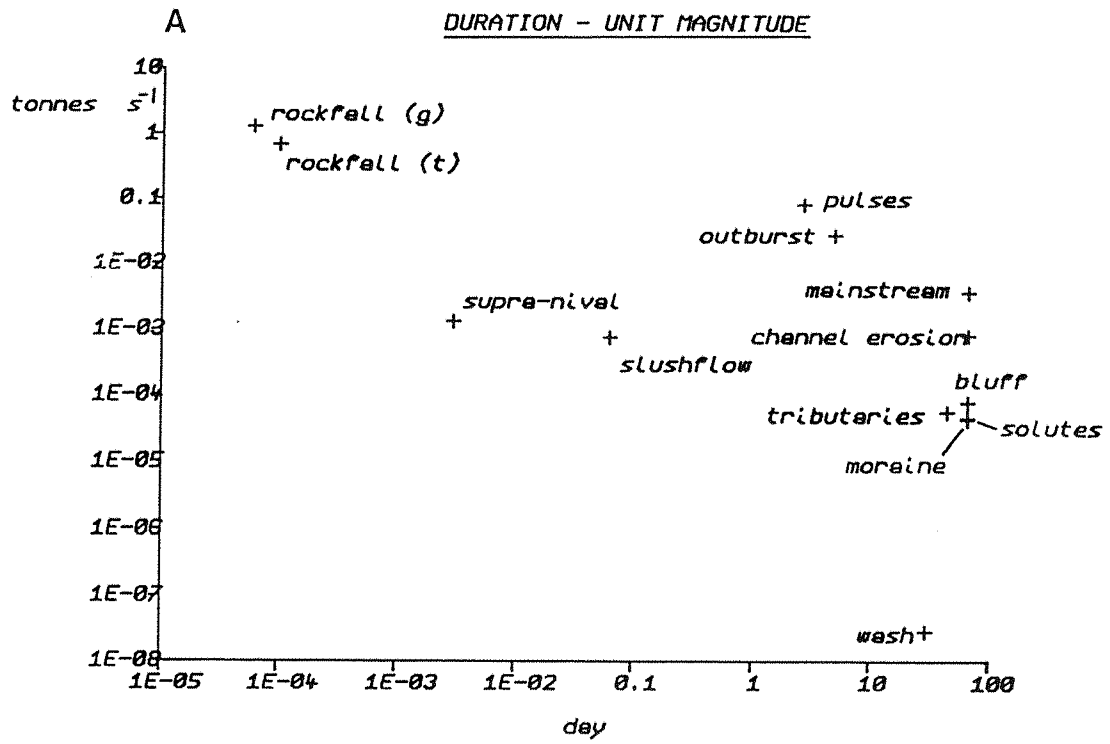
The use of energy concepts in describing and interpreting processes and their linkages is particularly applicable to the present sediment budget study, since the first step in the energy approach is the need for a system definition. Proglacial systems offer an excellent opportunity for testing the energy approach, because they are characterised by processes which exhibit extreme arrays of frequency and magnitude. The concepts of frequency and magnitude are fundamental in interpreting the effects of energy events because the total amount of work done and the instantaneous power expenditure depends on these properties. Work done varies greatly with the magnitude of the single event and the cumulative effects of several events, since the existence of thresholds regulates the intensity of change. In terms of the proglacial fluvial sediment budget, the most important question is the same as that posed by Chorley and Kennedy (1971, p. 197) which is "...... to determine the magnitude of the energy events which are most effective in changing natural systems".

The approach adopted here first considers the frequency and magnitude characteristics of the sediment transport processes (Figure 9.3) and secondly the magnitude of these processes expressed as energy events (Figure 9.4 and Figure 9.5).

Wolman and Miller (1960) recognised the relationship between magnitude and frequency and attempted to determine whether the cumulative work of small frequent events compared with the work accomplished in a single large event. In this context, the data collected in the Bas Arolla proglacial fluvial sediment budget Figure 9.3 successfully convey the essence of the magnitude frequency argument. Figure 9.3a is a plot of the frequency of the event expressed in days plotted against the event sediment yield divided by the duration of the transport event in seconds. Events of 67 day duration represent continuous processes since they operated for the full period of field study (e.g. mainstream discharge). At the other extreme only one rockfall from the sideslopes deposited sediment in the valley train and this only lasted 10 seconds.

It is worth remembering that these are processes acting within the proglacial fluvial sediment system and they therefore involve both fluvial sediment transport processes and processes which deliver sediment to fluvial sediment stores. Processes such as rockfall and supra-nival flows have low frequency of occurrence in the fluvial sediment budget but are much more widespread in the slope environment in general. Examining the distribution of processes in Figure 9.3a shows that slope processes tend to have high unit sediment-yield magnitudes but are short-lived. More continuous processes (to the right of the diagram) are the streamflow- dominated sediment transport mechanisms, but their unit sediment-yield magnitudes are low. Flushing events, small scale pulses and outbursts are relatively short-lived events but tend to be high in magnitude. The close proximity of these two events on the graph clearly illustrates that for this subset of processes (flushes), high frequency pulses are of a slightly greater unit magnitude than the much larger total magnitude single outburst event. One minor inconsistency in this approach is that moraine transport and bluff erosion are treated here as continuous processes, but in reality they consist of a series of small events. A similar argument could be in relation to sediment transport in the main channel, since individual particles will be transported along semi-continuous pathways.

Replotting these data, but using the total mass of sediment accumulated over the 67 day budget period (Figure 9.3b), produces an inverse trend in the relationship shown in Figure 9.3a. This clearly shows that the more continuous low magnitude processes exhibit a dominance over the low frequency high unit magnitude processes. It also reflects the results of the sediment budget (Figure 9.1), since the most important processes are the main channel flows of the valley train whilst the tributary sources are inferior in their role in the budget. An anomaly in both the diagrams in Figure 9.3 is the position of slopewash in the relationships. Although continuous, the magnitude of the contribution of slopewash to the fluvial sediment budget is minor.



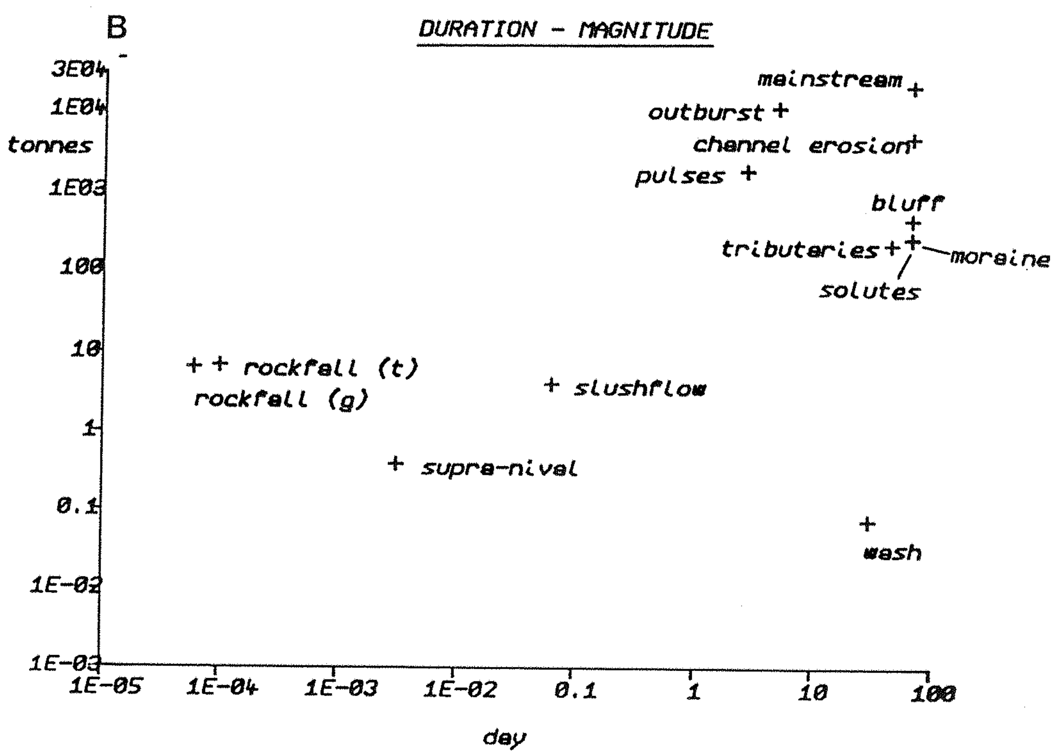


Figure 9.3 Duration - magnitude relationship between Bas Arolla proglacial sediment budget processes. (A) Duration - unit magnitude relationship. (B) Duration - magnitude relationship. (May - July 1987).

Because slopewash is only considered where it directly contributes sediment to the tributary streams, this under-estimates the importance of this process because it acts over a much larger slope area. One complication of this simple approach is that due to the existence of thresholds, large material fluxes do not always generate marked changes in the landscape. The opposite is also true, since small material fluxes may cross a threshold. In this latter case geomorphic memory and recovery concepts are important (Wolman and Gerson, 1980) since these set the stage for subsequent processes. In this respect geomorphic events need to be viewed as series, and in order to interpret the current level of geomorphic activity in a particular system the history of the system needs to be known. The paraglacial model of Church and Ryder (1972) is useful in this respect since it considers fluvial processes conditioned by glaciation and therefore includes a historical (recovery or memory) component in the estimation of paraglacial sediment yield.

The relationships shown in Figure 9.3 crudely define groupings of processes into subsets including: a mainstream channel subset; a subset comprised of bluffs, tributaries and moraines; and a subset of slope processes. These three subsets define three different morphological areas namely: the valley train; the margins of the valley train and the side slopes. A fourth set, defining sediment loss from inter-tributary areas, is probably represented by slope wash processes. This distinction into process-set environments defines a spatially stratified environment where the flows of matter between the subsets are not fully integrated (Thorn, 1987). A good example of this, in the alpine context, is from the Colorado Front Range, where Caine (1974) suggest a decoupling between the hillslope and fluvial sediment systems. As with the sediment budget itself, it must be stressed that the magnitude ranking of a particular sediment yield process will be sensitive to the location of the temporal and spatial boundaries of the system under consideration.

Extending this simple frequency - magnitude approach, the same data can be considered within an energetics framework. Here energy of material transfer processes is seen as the capacity to perform work. Under these circumstances the energy of physical systems is most suitably measured as work or power (Caine, 1976; Gregory, 1987).

The capacity to perform work will be dependent on the the available potential energy in the environment and the rate at which this is translated into kinetic energy. In alpine mountainous terrain this statement represents the physical realisation of the geomorphic catch-phrase 'high energy environment'. Gregory (1987) provides an excellent review of the applications of the energy approach in physical geography. What follows here is a specific example of this approach in defining the energetics of the proglacial fluvial sediment budget.

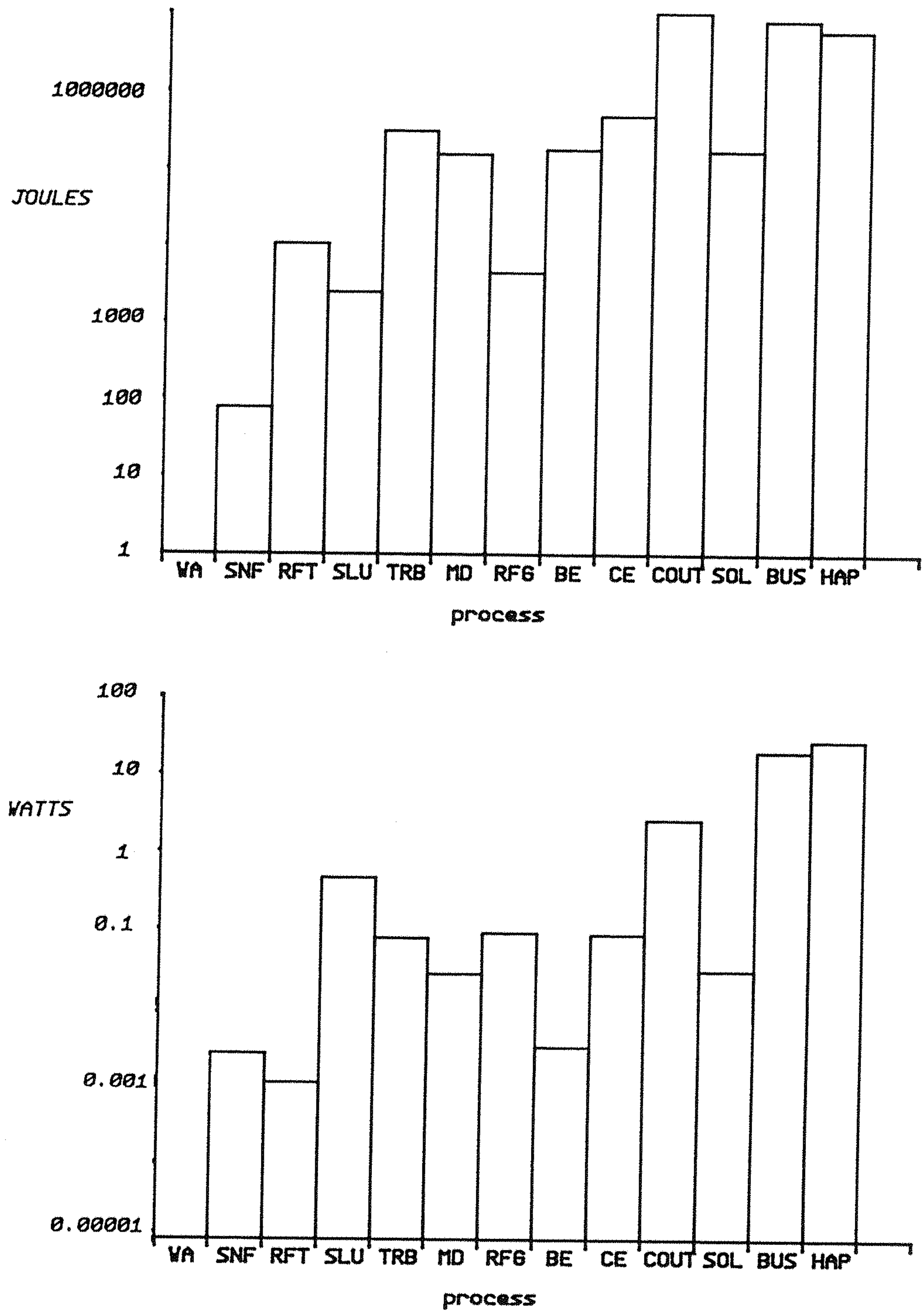
Table 9.1 and Figure 9.4 provides the work and power profiles for the sediment transfer processes measured in the Bas Arolla proglacial fluvial sediment system. However following Caine (1976), who rightly cautions the comparison of data of this kind without considering the assumptions involved, each process calculation is reviewed.

Work is calculated from two general equations based on Caine (1976):

$$\text{work} = m \times g\ (h_1 - h_2) \qquad (9.1)$$

$$\text{work} = m \times g\ (d \times \sin s^{o}) \qquad (9.2)$$

where m = mass, g = acceleration due to gravity, h =



KEY

WA = WASH
SNF = SUPRA-NIVAL FLOW
RFT = ROCKFALL (TRIBUTARY)
SLU = SLUSHFLOW
TRB = TRIBUTARY STREAMFLOW
MD = MORAINE DEPOSITION
RFG = ROCKFALL (GLACIER)
BE = BANK EROSION
CE = CHANNEL EROSION
COUT = CLASTIC SEDIMENT OUTPUT
SOL = SOLUTE OUTPUT
BUS = OUTBURST FLOOD
HAP = HAUT AROLLA PULSES

Figure 9.4 Bas Arolla sediment budget processes expressed in terms of Joules (work) and Watts (power), (May-July, 1987).

elevation ($h_1 - h_2$ is the change in elevation in the time interval between 1 and 2), d = distance and s = slope angle. Since all values in Table 9.1 are already expressed in mass and their calculation explained in the preceding chapters, the only term to be estimated is the elevation change during the interval between time h_1 and h_2. For the proglacial fluvial and slope processes studied here either equation 9.1 or 9.2 can be used. For main channel processes and solutes this is estimated as the difference in elevation between the catchment outlet and the centre-point of the bed of the Bas Arolla glacier basin since this is an approximation to the site of sediment entrainment. For the sediment pulses from the Haut Arolla sediment purges, the elevation difference between the purge outlet and the catchment outlet is taken. For tributaries, the entrainment source is taken as the mid-slope elevation, with h_2 defined as the edge of the valley train. Similarly, for channel erosion the mid-point source elevation is defined as the centre of the valley train and for bluff erosion this is taken as an average bank height. The difference between h_1 and h_2 for moraine transfers is the height of the glacier snout. For slope processes, run-out distances were observed and, based on slope angle, the elevation difference was calculated (equation 9.2). The only estimate that needed special treatment was slopewash, since the original data were expressed in mass discharged across a unit contour length, and a representative slope movement distance could not be estimated directly. Therefore the mass was converted to volume by dividing by a bulk density of 1.9 (Caine, 1976). This was then divided by the cross sectional area through which this discharge passed. For the cross section calculation, depth was approximated by assuming that only the surface layer, one particle-diameter thick, is moving, whose thickness is represented by the D_{50} of the surface sediment. By this means velocity was calculated and the distance moved by a particle calculated. It can be seen that this approach has strong conceptual links with the *tonne-metres vertical* concept of Rapp (1960).

Power was then estimated by dividing work by the time over which the geomorphic process operated. In this study the times over which the particular sediment transfer processes operated are given in Table 9.1 as duration or 'action time'. Action time being the time (in seconds) in which geomorphic work is done. For continuous processes this is given by the length of the monitoring period. For discontinuous processes it is derived from field observation. The action time for pulses is defined as the cumulative action times of each individual pulse. Slopewash action time is difficult to estimate but in this case it is approximated by the number of rain days, since these are the likely periods of maximum movement.

Based on these calculations Figure 9.4 presents the work and power estimates for the geomorphic processes operating in the Bas Arolla proglacial fluvial sediment budget. Given the log scale to these diagrams, one of their most striking attributes is the range in energy estimates over 13 orders of magnitude for both work and power. The variations between power and work are also interesting. For example, the most *powerful* events are the flushes, the outburst and Haut Arolla sediment pulses. Although short in duration, they perform as much work as the continuous discharge of sediment. Comparing bank erosion with the slushflow event, bank erosion although of low power expenditure is quasi-continuous and is therefore important in terms of total work, whereas the slush flow has a high energy expenditure rate but is so short-lived that it does little total work. Again, broad process-morphology groupings are evident. Channel processes are high in power and achieve a great deal of work, solutes tend to be relatively low in power, but still achieve a substantial amount of work: whilst the slope processes are highly varied in both power and work, reflecting the episodic nature of their activity. From this diagram and from the discussion of magnitude-frequency concepts, a pattern is emerging with regard to the energy structure of the proglacial fluvial sediment budget. As a working model, three process groups are distinguishable. These groups are: the channel group, which are highly powerful and achieve considerable work ; the valley train marginal group, which are much less powerful but their continuity of operation ensures that they are work effective; and the slope processes which tend to be of variable power but lack the magnitude and frequency to accomplish great work. Again slopewash stands out alone and in terms of sediment supply to the fluvial sediment budget is of negligible importance (though its spatial significance should be noted). This division into sub-processes seems to stand out when energy is plotted against magnitude (Figure 9.5). Again the same process groups emerge.

The relative position of solute transport is interesting, since it could be that as a process it should be treated separately from the other sediment transport processes because only one element of this complex system was measured. In addition, it is envisaged that solute transport will exhibit a greater degree of variation than sediment transport processes because solutes are generally transported through the system at under-capacity loads (i.e. the power expended is relatively low in comparison to the work done e.g. Rapp (1960)). Furthermore, in mountain environments the range in power and work for a particular sediment transport process, given variations in rocktype, will be less than the range in power and work of solutes. This is because the energetics of the chemical system have a bigger influence over solute processes than they would over the physics of sediment transport.

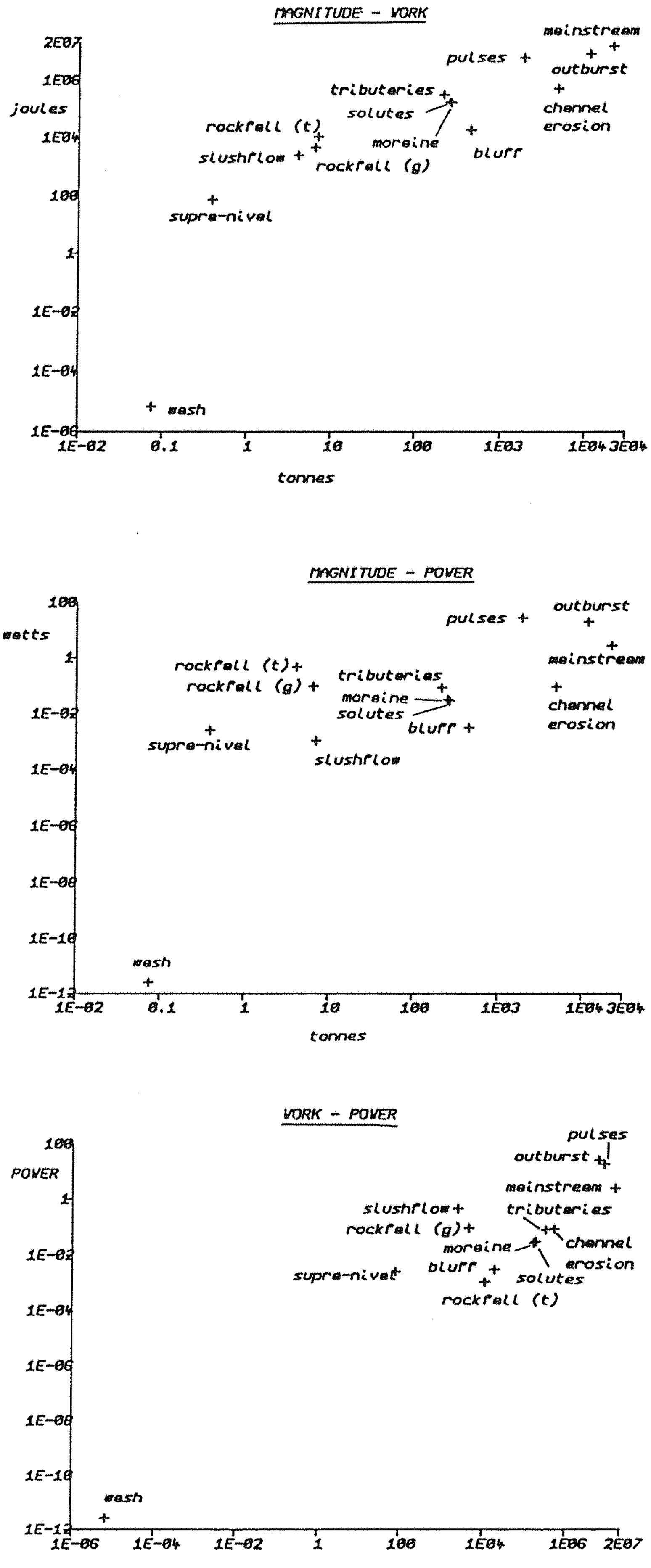


Figure 9.5 Energy relationships between Bas Aolla proglacial sediment budget components (May - July, 1987).

The final plot in Figure 9.5 compares work with power and in a sense can be considered as an efficiency diagram because it relates the translation of power into work. By fitting a regression to the data it should be possible to define the general stability of the environment being studied, because the slope on the regression would indicate the power needed to achieve a given amount of work and the intercept would possibly define a critical rate of energy expenditure at which processes were initiated. For example in a 'badlands' environment a small power is required to start erosion and once started it progresses rapidly. Therefore the regression of sediment transport processes in a work-power plot would have a high slope and a low intercept. The characteristics of similar regression equations could be used to define the general power trends for the particular types of environment.

Alternatively, in the same way as the sediment balance equation was used to define the sediment budget (Figure 9.2), the energy balance for any particular environment can be calculated. This gives a total figure for the energy balance of the proglacial fluvial sediment budget of 1.47×10^{7} Joules. The relative contributions from proglacial and glacial energy sources are interesting since proglacial sources expend 20.6% of the power but only perform 6.6% of the work. Particularly important is the outburst event which during its brief action time produced 53.9% of the work because it is seven times more powerful than average channel sediment transport processes.

With such sparse data for the energetics of sediment transport processes, comparisons between environments are difficult . Caine (1976) compared several mountain catchments and found that rates were similar. This comparison cannot be extended to the data presented here because in the Bas Arolla proglacial zone data were collected to represent a fluvial sediment budget, whereas in Caine (1976) the data were collected for a study of hillslope sediment transport processes. The only valid comparison relates to solute transport, and in this context Switzerland has solute transport rates an order of magnitude less than those of Caine (1976) for a small mountain catchment in the San Juan Mountains, Colorado, USA.

Although very useful for interpreting geomorphic processes, energetics must be applied with care since their extrapolation can lead to misunderstanding. An example relevant to the present study concerns the relationship between discharge and unit bed stream power during flood (Figure 9.6). Bull (1979) defines a threshold of critical power in streams whereby a critical value of 1.0 approximates the threshold condition at which, stream power = critical power. For the Bas Arolla proglacial stream critical power is defined as the critical condition for full bed mobility calculated using the Schoklitsch equation (Chapter 4) (Figure 9.6). Based on a range of discharges and average channel width, the unit bed stream power is calculated (Figure 9.6, O-A-B) which gives a linear increase of power with discharge. However this takes no account of the increase in channel width associated with proglacial flooding. If this is taken into account then power does not follow a linear trend but follows a curving trajectory (Figure 9.6, O-A-C). What this defines is a loss of stream power with increased channel width due to greater resistance. This has three implications: firstly, simple power-work relations must recognise the problems of thresholds and non-linearity in process-response; secondly, because of this effect there is an optimal range in discharge which will cause extensive erosion in the proglacial zone; and thirdly, the initial channel capacity before flood may influence the outcome of the flood event. Even this is a simplification of the problem because no account is taken of the sediment load carried by the flood or

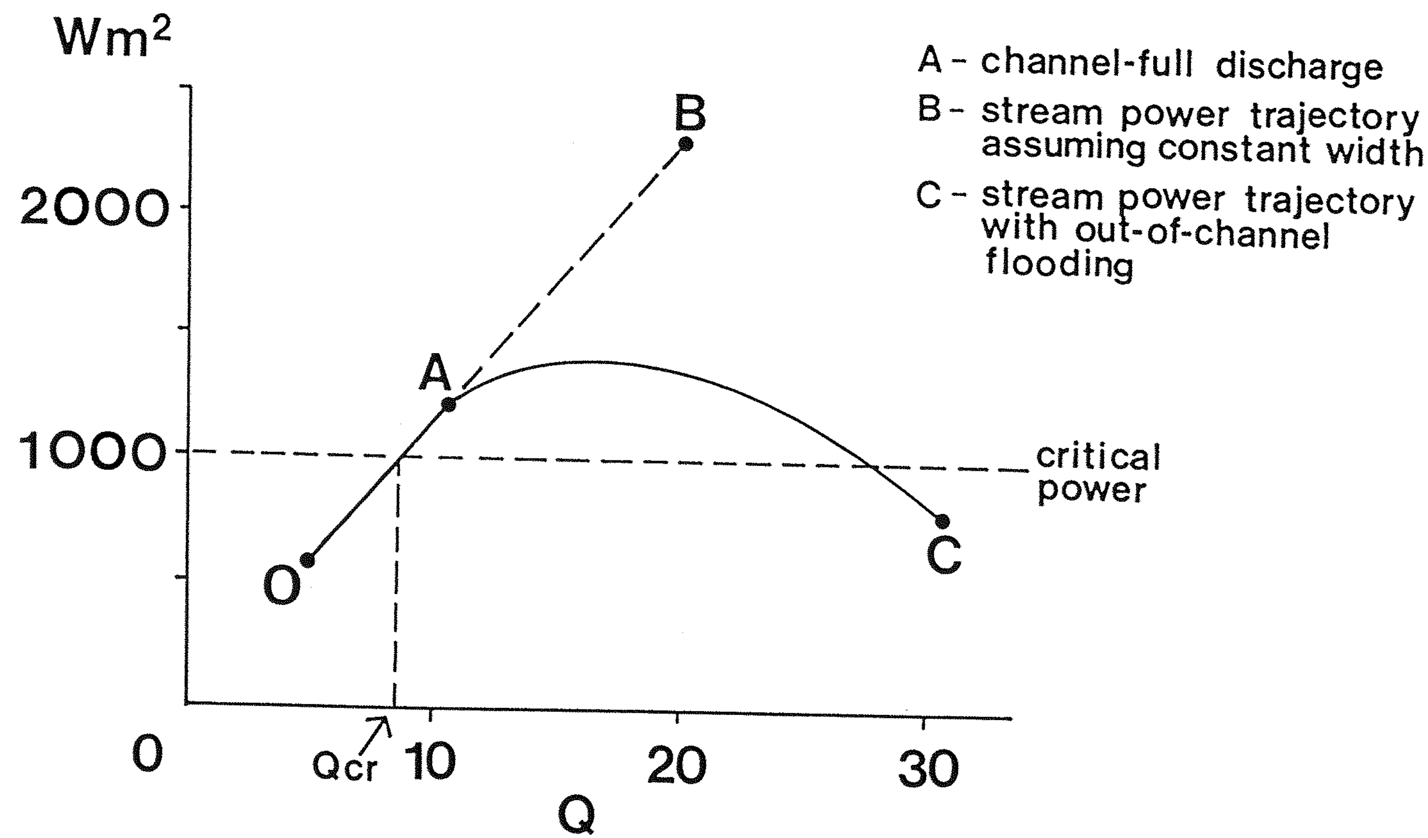


Figure 9.6 Relationship between discharge and unit bed stream power during flood. Values approximated for the Bas Glacier d'Arolla proglacial zone.

entrenchment of the stream channel during flood. This problem introduces the need for energetics research to consider the relationship between potential and kinetic energy in the natural environment.

9.4 Conclusions

Results from the sediment budget (Figure 9.1) show that proglacial sediment sources in the Bas Arolla fluvial sediment system for 1987 contribute approximately 23% of the sediment received at the catchment outlet. Of this an overwhelming proportion (95%) is derived from erosion of the valley train. This implies that tributaries and hillslope sediment sources contribute little to the overall budget. It is stressed that these conclusions only refer to the Bas Arolla proglacial zone and for the time period of monitoring. This rather narrow time window is partly constrained by the limitations of 'Ph.D. time' (Lewin, 1980) and partly by the short duration high intensity monitoring programme which is required to get a good budget estimate. This constraint may not be so limiting, given that in glacial environments the magnitude-frequency distribution of events is contracted over a relatively short time base. This means that even with a narrow monitoring window, a large spectrum of processes will be included in the budget (Table 9.1). Nevertheless, these estimates should be treated as preliminary since Walling (1978) suggests that 10 years of monitoring are required before the sediment transport system can be adequately characterised. One methodological problem which arises when a sediment budget is defined over a short timescale is the recognition that budget processes and their quantification are executed simultaneously, which may lead to a partial definition of the budget.

Expressing the budget processes in terms of work and power provides an interesting perspective on the sediment system (Caine, 1976). It appears that four basic process-morphologic sets can be distinguished in the Bas Arolla proglacial fluvial sediment system. These are: 1) A valley train - channel subset; (2) a valley train margin subset of quasi-continuous processes; (3) a sub-set of hillslope processes; and (4) slopewash. Contrasts in energy levels between the subsets may explain the geomorphic development of each landscape element e.g. the entrenched nature of the valley train may be in response to high levels of available energy? A hierarchy of energy levels may correspond with a hierarchy of sediment source areas?

Energetics provide a valuable means of deciphering sediment system interactions, and growing awareness of their potential (Gregory, 1987) may incite a revolution in physical geography of the type envisaged by Kuhn (1970) (Chapter 1). This trend towards sediment budgets and energy models is likely to resurrect the old prematurely abandoned systems models of the late 1960's and early 1970's e.g. Chorley and Kennedy (1971). The niche for this study lies in the need to illuminate 'black-box' sediment delivery models (Walling, 1983), and although this project has not provided the flood-lights for the whole system it is perhaps a torch which allows us to see one part in detail.

Chapter 10.

CONCLUSIONS

10.1 Introduction

This chapter summarises the results of work undertaken on the Bas Arolla, Tsidjiore Nouve and Haut Arolla catchments, Val'd Hérens, Switzerland in 1986 and the Bas Arolla catchment in 1987. For the 1987 field season all effort and resources were used to define a sediment budget for the fluvial sediment system of the Bas Arolla proglacial zone whereas in 1986 sediment yield studies were undertaken in all three basins. This was an attempt to move away from the 'black box' approach to monitoring glacierised basin sediment yield by applying a sediment budget framework to sediment transfer processes acting in the proglacial zone. This involved recognition and quantification of sediment transfer processes and storage sites, and the identification of linkages amongst transport processes (Dietrich et al., 1982).

Results summarised below provide the answers to the four research questions posed in the introduction to the project, namely:

1) What are the proportions of bedload, suspended load and solute load in the proglacial meltwater stream?

2) How is sediment load in the proglacial stream modified by contributions from sediment sources along the valley train and what are the proglacial fluvial sediment sources?

3) What are the consequences of sediment transport in modifying the morphology of the proglacial zone?

4) What are the major transport processes, storage elements and linkages in the proglacial fluvial sediment budget?

10.2 Summary of results

Field measurements (Chapter 2) were focused on four areas of the proglacial zone: at the outlet, where total yields were monitored; within the main proglacial channel; on the bluffs along the valley train margins; and on the valley side slopes and tributaries. Most of the measurement techniques were standard but due to the emphasis placed on suspended sediment measurement, laboratory, field and flume calibration experiments were carried out with turbidity meters (Appendix 1). Results from these experiments emphasised the need for thorough calibration of turbidity meters, careful siting of the instruments in the field and continued field maintenance involving cleaning of lenses, regular battery changing and shielding from ambient light if used at low sediment concentrations.

Sediment output from the three study basins (Chapter 3), Bas Arolla, Tsidjiore Nouve and Haut Arolla 1986 showed Bas Arolla to have the greatest sediment output, second was Haut Arolla and Tsidjiore Nouve had the lowest sediment yield. However, standardising these results for catchment area showed that Tsidjiore Nouve had the greatest relative yield at 4174 $t.km^2$, Bas Arolla was next highest with 4093 t.km2 and Haut Arolla had by far the lowest yield of only 2422 $t.km^2$. Of the total load from the Bas Arolla basin 58-64% was transported as bedload whilst at Tsidjiore Nouve only 36-51% was bedload (1986-1987 estimates). No load component estimates are available for Haut Arolla. Solute yields at Bas Arolla were less than 1% of the total load. Compared to regional estimates of sediment transport in other glacierised basins in the Val'd Hérens, the three study basins account for 86% of the measured sediment output from the region. These results accord with review of glacierised and mountainous catchment studies, which suggests that globally, glacierised basins have high sediment and solute yields with clastic sediment accounting for the major part of the load.

Paired suspended sediment monitoring stations at proximal and distal ends of the proglacial zone in 1987 showed an increase in the suspended sediment load over this area of 10% (580 tonnes) between the end of May and the end of July (Chapter 4). Tributary inputs following rainstorms accounted for a small proportion of this load but the major part was derived from channel sources. Bedload was not monitored using a paired station approach but measurements at a single section in the proglacial reach showed that the movement of bedload was temporally and spatially highly variable. Bedload moved in threads in the centre of the channel and from pool to pool along the long profile. For the Bas Arolla proglacial stream, two-phase bedload transport models (e.g. Jackson and Beschta, 1982) should be extended to three-phases (Figure 4.12) to account for the hierarchy of bed structures associated with the great diversity of sediment sizes characteristic of mountain streams. Bedload transport cannot be satisfactorily predicted using the Schoklitsch equation. However, the equation can be used for describing the very steep initial transport phase associated with large scale bedload movement. These findings confirm those of Duijsings (1985) who suggests that in small basins sediment delivery is usually close to 100% but can nevertheless vary markedly over short-timescales due to bed storage.

Sediment contributions from tributary streams were highly variable and were generally more important during rainstorms early in the ablation season (Chapter 5). Preliminary estimates of suspended sediment output ranged between 103.6-319.2 tonnes with 99% contributed from one tributary, and 52% transported in one day (June 15th). Estimates of bedload transport also varied greatly and, depending on the method of calculation, could be 3.32 to 8.55 tonnes (based on trap yields) or 17.1 to 50.4 tonnes (estimated as loss from tributary storage). Sediment transport in tributaries could not be easily partitioned into bedload and suspended load transport components since in these very steep shallow channels at high flows small pebbles were carried in suspension. The estimated mean total load from tributaries was 220.15 tonnes (Error +/- 71%) (Chapter 5).

Direct inputs of sediment to the valley train from the side slopes in 1987 came from streamflow over snow, slushflow and rockfall. Slope wash in the vicinity of tributary channels only contributed minor amounts of sediment (Chapter 5).

Monitoring of Bas Arolla stream channel change in 1986 and 1987 (Chapter 6) showed that changes were an order of magnitude greater during flood than normal flows (Figure 6.5). The Bas Arolla stream showed a cyclic channel response to flood. This involved a switching of habit from a single thread, low sinuousity stream to a wider channel with 'bulbs' or 'nodes' of braiding. Following flood, the channel pattern rationalised by a reduction in the extent of braiding so that a single thread channel was re-established by the end of the ablation season. Comparison between July floods in 1986 and 1987 showed that the 1986 flood was characterized by a laterally shifting thalweg and relatively low sediment yield. In 1987 the main characteristic of the flood was thalweg incision which lead to higher sediment yields. The effects of these floods left little imprint on the channel, since following flood the braided channel had a single dominant thread of flow which soon lead to a single thread channel once more. Although spatial patterns of channel response were evident there was considerable variation in channel adjustment. This episodic nature of channel adjustment and sediment transport fits the episodic model of sediment yields from mountain basins of Trimble (1988).

Erosion of bluffs along the margin of the valley train (chapter 7) constituted an important source of sediment (452.2 tonnes). Bluff erosion rates were 4 to 25 times greater during flood than in periods following flood events. This is because during a flood sediment is contributed by buttress failure, but following the flood erosion processes dominate. Early in the ablation season bluff erosion is also accelerated as banks emerge from under the winter snow-cover.

Sediment flushing events in proglacial streams cover a large spectrum from low frequency high magnitude meltwater floods to frequent small-scale pulses (Chapter 8). Three meltwater flood occurred during the period of study; in July 1986 and in July and August 1987. The 1986 flood was concentrated in the valley train and was geomorphologically the least effective flood. The July 1987 flood was a more complex event involving an overflow of water over the snout of the Bas Glacier d'Arolla. This event transported a considerable amount of sediment and had an influence over the entire valley train and lower part of the glacier. The August, 1987 flood was unusual in the sense that its origin was partly artificial, with the flood being generated from the release of stored meltwater from the hydro-electric system. Because the flood release was from the Haut Arolla basin the flood had the greatest spatial influence, affecting glacier, valley train and slopes. Recovery following the July 1986 flood was gauged by assessing change in proglacial channel characteristics. Following this event recovery in the channel system seemed to be complete by the end of the ablation season.

Frequent low magnitude pulses in suspended sediment were usefully investigated using small-scale field experiments (Chapter 8). Based on these results it is suggested that large-scale sediment pulses are unlikely to originate from bank collapse. Correspondence between Haut Arolla sediment trap purging and and pulses in suspended sediment recorded in the Bas Arolla proglacial stream suggested a causal link. Travel times of these pulses through the glacial drainage network of the Bas Glacier d'Arolla were very rapid, which implies that the internal drainage must be very efficient and therefore likely to consist of a large conduit system. Both meltwater floods and pulses offer a valuable insight into the glacio-hydrological network of the Bas Glacier d'Arolla and suggest intimate links between glacial and proglacial sediment and water systems.

Results from the Bas Arolla proglacial fluvial sediment budget 1987 (Figure 9.1) showed that proglacial sediment sources contributed 23% of the sediment yielded at the catchment outlet. Of this an overwhelming proportion (95%) was derived from the valley train. This implies that tributaries and hillslopes contribute little to the overall budget. The vast majority of this valley train sediment was transported during the short period of meltwater flooding in July 1987.

Expressing the sediment budget processes in terms of work and power (Table 9.1) suggested that four basic process-morphologic sub-sets could be distinguished in the Bas Arolla proglacial fluvial sediment system. These were: (1) a valley train channel sub-set; (2) a valley train margin sub-set; (3) a subset of hillslope processes; and (4) slopewash. Contrasts in energy levels between the subsets may explain the geomorphic development of each landscape element.

However, most field projects should be viewed as *stepping stones* because the results of a project will generate as many questions as are answered.

10.3 Further work - *What next ?*

The sediment budget proposed here is inevitably specific to the Bas Arolla proglacial fluvial sediment system but the implications are more widely applicable to the Alpine proglacial zone as a whole. From this project several further research objectives can be identified. These can be divided into sediment yield-budget considerations and individual objectives for research on particular material transfer processes.

10.3.1 Sediment yield-budget considerations

1) Continuous monitoring of glacierised basin sediment yield should be continued and where possible new programmes initiated so that long-term variability in yield can be estimated. Measurements should ideally be standardised so that sediment output results are comparable between regions.

2) Sediment budgets need to be constructed in other proglacial zones to assess the inter-basin variability in sediment supply characteristics. An energy approach may be useful in this context, since results would be directly comparable and ranking of the importance of sediment transfer processes may be used as a 'geomorphological signature' for the particular sediment system.

3) In order to control for between-site variance, the level of within-site variability should be addressed by continuing sediment budget monitoring at single sites over many years.

4) The effect of large-scale slope sediment inputs (i.e. slope failure or en-masse debris transport) was not quantified during the 1987 budget period. These processes need to be quantified if the budget temporal boundaries are to be widened.

5) Because the 1987 budget was only constructed for the period from the end of May until the end of July, a full ablation season budget from May to September/October needs to be constructed. A full year budget would be a further step, since many slope sediment yields take place before the commencement of the main melt season.

6) Proglacial sediment budgets need to be combined with glacial sediment budgets in order to specify the linkages between the two. In this respect pioneering studies of tracing sediment transport through glaciers are urgently required. This project has very briefly touched on this area by making inferences about glacial hydrology based on sediment output characteristics during meltwater floods and pulse events. This is essentially a 'bottom-up' approach: what is need is a 'tracing-down' approach. This would involve the tracing of sediment from source through the glacial system.

7) The budget is 'spatially limited' because it was only constructed for the zone immediately in front of the glacier. Therefore, this sediment budget represents an extreme proximal condition. With greater distance from the glacier, the pattern of sediment delivery may radically alter. Therefore a study determining sediment contributions and storage in a down valley direction is important. Down valley changes in sediment supply may represent the spatial manifestation of the Church and Ryder (1972) paraglacial model for sediment delivery following deglaciation.

10.3.2 Objectives for material transfer process studies

1) Bedload transport in mountain streams is poorly understood. Further field measurements are required to account for the inherent variability in transport rates and to test the three-phase bedload model proposed in Chapter 4. Proglacial streams, although complicated in structure, offer a superb geomorphic field laboratory for studying bedload transport due to the large frequent variations in discharge, abundant sediment supplies and the dynamic nature of bedload transport. It would be useful in future studies, which focus on bedload transport past a single site, to erect a portable of sampling platform across the full width of the stream channel. This would considerably aid data collection and greatly reduce bed disturbance during sampling. Poor correspondence between bedload transport equations and observed transport rates in mountain streams will be better explained as new data become available and better measurement of flow characteristics are made (e.g. Wiberg and Smith, 1987).

2) The division between bedload and suspended load is never clear but in proglacial streams this is even more diffuse. This factor as not been sufficiently quantified to determine the size range over which mutual exchanges of particles occur. Nevertheless, in steep, shallow tributaries the two ranges may almost completely overlap (Chapter 5). Sampling should be designed to account for this effect.

3) Channel change in the Bas Arolla proglacial stream is highly episodic and spatially highly variable. This is thought to be a common characteristic of most steep, coarse bed mountain streams and as such the thresholds for change need to be identified. This problem is intimately associated with the movement of bedload in mountain streams.

4) The suggested sequence of bluff erosion in Figure 7.9 needs to be tested with a full season's data of bank retreat rates at several sites in the proglacial zone.

5) Investigations into the causes and nature of suspended sediment pulses should be a priority. The examples discussed here were artificially generated pulses but can be used effectively to provide an insight into the 'plumbing' of the glacier. Natural pulses also occur and if processes of generation were better understood this information could be used to infer detail about the internal environment of the glacier.

6) Studies examining the interaction between glacial and proglacial sediment systems in Alpine environments are virtually non existent. It is vital to integrate measurements from both areas if the causes of meltwater flooding and the generation of sediment pulses are to be determined. It appears that the ice margin is a boundary over which very few researchers cross!

7) There is also merit in extending the power/work, frequency-magnitude studies to other years i.e. are the rankings of power and work dependent on the peculiarities of an individual season or are the process rankings similar, only varying in total magnitude? Application of this approach may need revision given the episodic nature of some of the sediment transfer processes e.g. sediment loss from the valley train during flood accounts for virtually all the erosion of the proglacial channel.

10.3.3 A step in the right direction?

There is clearly a compromise between attempting to understand the detail of one site or system and gaining experience from several areas in order to appreciate the wider variations in a particular system. It could be argued that most Geomorphologists have been 'fidgeting' in the environment setting up capricious field research programmes over timescales which are best characterized as 'here today gone tomorrow'. This provides few long-term data series and very little spatially concentrated research effort.

First and foremost the rubicon must now be to cross the hurdle of designing and implementing research programmes which have long-term aims at particular sites. Only when the internal variability of a particular system is suitably characterised can other sites be considered. In mountainous catchments examples of this kind of work are few. However, the work of Rapp (1960) in Karkevagge, Northern Scandanavia; Østrem, Hooke and fellow researchers working on the glaciers of Southwest Norway (e.g. Hooke et al., 1985); Collins on the Gornergletscher; and research in the Arolla Valley, Valais, Switzerland (Gurnell and Clark, 1987) are notable exceptions. The shift from single site monitoring to regional analysis is not without drawback since the change in scale is associated with the jettisoning of detail, and without investing equal research effort at all sites, comparisons between studies will be difficult.

Research in the Department of Geography, University of Southampton has developed a site specific research programme in the Arolla Valley, Val d'Hérens, Switzerland. The Tsidjiore Nouve glacier in particular has been the focus of much of this work (Gurnell and Clark, 1987). The combination of research on glacial sediment transport-moraine formation (Small et al., 1979; Small et al., 1984); proglacial streamflow series (Gurnell and Fenn, 1984; Gurnell and Fenn, 1985) and more recently sediment yield characteristics (Gurnell, Warburton and Clark, 1988) provide a detailed assessment of the sediment system for Tsidjiore Nouve glacier. The Bas Arolla glacier, in the same valley has also been previously studied (Fenn, 1983). This study filled a gap in the existing research from the Arolla Valley by examining sediment transfer processes in the proglacial zone of the Bas Arolla glacier.

It is time to recognise that most of the existing glacial sediment system studies are effigies and if research is to progress conformity in measurement techniques and research goals needs to be established.

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APPENDIX 1

Calibration of Suspended Sediment Monitors

The purpose of this short appendix is to outline the principle of turbidity measurement and how turbidity can be related to suspended sediment transport. In order to accomplish this aim four sections are considered: section 1 provides a background to the measurement of turbidity and the type of equipment used; section 2 describes a calibration experiment based on flume measurements; guidelines for the field siting of turbidity meters is given in section 3; with a summary of the turbidity - suspended sediment concentration relationships given in section 4.

1. Measurement of suspended sediment concentration

Proglacial meltwater streams are characterised by high suspended loads which show sudden marked variations in concentration (Gurnell, 1987b). For the streams studied here, concentrations generally vary between 500-3000 mg l^{-1} but may be as high as 10000 to 15000 mg l^{-1}. Therefore measurements need to take account of a wide range of concentrations at a resolution which will characterise rapid fluctuations in. Under these conditions the turbidity meter, once calibrated, can provide a reliable, near-continuous record of suspended sediment concentration variations (Truhlar, 1978).

Turbidity is an expression of the optical property of a sample which causes light to be scattered and absorbed, rather than being transmitted in a straight line through a sample. Therefore a turbidity meter measures the intensity of a beam of light passing through a turbid suspension, the source and measuring device (a photocell) being in the same straight-line. This method is known as optical extinction, or attenuation, and is measured by an instrument called a absorptiometer. Alternatively, the intensity of scattered light can be measured at an angle to the incident beam and this is done using a nephelometer. Light received by the photocell is converted to a current and displayed on a meter; the reading on the meter being directly proportional to the intensity of the light falling on the cell. Therefore the maximum intensity of transmitted light occurs for zero suspended sediment concentrations and decreases as the concentration of the suspension increases. Meters are usually fitted with an inverse logarithmic scale which measures the ratio of the light passing through clearwater to that transmitted through suspension. In this case the readings (referred to as the optical densities of the suspension) are almost directly proportional to the concentration of the suspension (up to at least 5000 mg l^{-1}) (Fish, 1983).

Three Partech portable suspended solids monitors (Type 7000, 3RP) were used in this study. This instrument consists of a monitor connected to a sensor linked to an external Rustrack chart recorder. The instrument requires a 12 volt power supply which can be provided from an internal or external battery pack, which will operate for a minimum of 22 hours. A low battery indicator provides a check on the battery level which suggests when readings become unreliable. All sensors were of a single-gap type (SDM-10) and consisted of a light source (2.2 watt bulb) and a detector separated by a gap of 6 mm. This is in contrast to a twin-gap sensor which has a light source and two detectors separated by gaps of 6 and 13 mm, which can compensate for 'fouling' of the optical windows, colour variations in the water and ambient light, since the difference between the two photocells is measured, but directional sunlight may still be a problem. The SDM-10 sensor has a nominal linear range of up to 5000 mg l^{-1} but can be used for concentrations greater than 25000 mg l^{-1} if a non-linear calibration curve is acceptable. The monitor is set by means of a zero potentiometer , set for clearwater, and a span potentiometer set for the maximum anticipated suspended sediment concentration. Three scales allow three separate scale ranges to be set. Data are recorded every two seconds on a Rustrack chart recorder which, in the case of the three monitors used in this study, had a chart speed of approximately 25 mm h^{-1}, although this varied. Chart life is approximately one month and the operational range of the instrument is between -5 and 60 ^{o}C .

Calibration of the instruments was carried out for two principal reasons:

1) To relate the monitor reading to the true concentration of suspended sediment.

2) To assess the stability of the monitor.

In studying suspended sediments in proglacial streams standard calibration procedures, using formazin turbidity standards, cannot be reliably used, since formazin is a mixture of hexamine and hydrazinium sulphate and bears little resemblance to proglacial suspended sediment properties. As Hach (1971) points out " about as far as one can go in relating turbidity to the quantity of suspended matter is to restrict comparisons to only cases where a succession of turbidity measurements are made with the same, or same type of turbidimeter, on samples containing the same turbidity material". Indeed calibration can only be considered valid for data from a particular stream (Richards, 1982) since no universal relationship exists between turbidity and suspended sediment concentration in natural streams (van Rijn and Schaafsmann, 1986). In fact, the relationship between turbidity and the suspended material is a function of particle shape, particle light absorption, particle size distribution, particle specific gravity, sediment colour and refractive index as well as sediment concentration (Thorn and Burt, 1975). Therefore calibration using 'natural' , site specific sediment concentrations is the crux of satisfactory and meaningful interpretation of the resulting data.

The calibration procedure used here is essentially the same as that outlined by Fish (1983). This involves setting of the zero potentiometer for clearwater and the span potentiometer at 100% for concentrations of 5000 mg l^{-1}, 10000 mg l^{-1} and 15000 mg l^{-1} on the respective scale ranges 1, 2 and 3. This was done by mixing a known weight of sediment (collected from the proglacial stream of the Bas Glacier d'Arolla and pre-seized retaining the less than 63 micron fraction) with a known volume of water. In order to provide more accurate calibrations of these approximate ranges a series of bottle samples (volume approximately half a litre) were collected in the field using polypropelene bottles, a USDH-48 sampler and a Manning automatic liquid sampler. At the time of sampling the time, volume of sample, turbidity value and range were recorded.

These samples were hand filtered, usually within 48 hours of collection, at approximately 40-45 cm mercury vacuum through preweighed Whatman 40 cellulose filters which retain sediment greater than 8 microns. Filters were then folded whilst moist and sealed in polythene bags to await drying and re-weighing in the laboratory.

2. Calibration Experiment

Since calibration of the turbidity meters using natural sediment concentrations was of primary importance a small-scale experiment was carried out to investigate the calibration of monitors under controlled conditions. The principal objectives were: 1) To compare calibrations using 'natural sediment' and 'formazin standard' concentrations. 2) To investigate the effects of ambient and directional light on the turbidity meter response. As infra-red turbidity meters should be unaffected by this effect a meter of this type was run concurrently in order to determine the relative merits of the two instruments. 3) To investigate the process of lens fouling. The experiment was carried out in conjunction with the Department of Civil Engineering as a part of a project investigating sediment dynamics in an estuarine environment, therefore the sediment used was taken from this source. Compared with the type of suspended material found in the Swiss field samples, the estuarine material is finer, dominantly silt with a relatively high organic content (10%).

The experiment was carried out in a recirculatory flume (length 9.5 m, width 0.61 m) at the University of Southampton's Chilworth Research Centre. Circulation was maintained by two pumps, which give a maximum flow velocity 0.254 $m^3 s^{-1}$ for a water depth of 0.38 m. Longitudinal and cross-sectional velocity profiles showed this to be relatively constant throughout the flume. Water temperature varied little, the maximum being 20 ^{o}C the minimum 14 ^{o}C. Evaporation losses were corrected by adding fresh water daily. This had little discernible effect on the turbidity values. Three sensors, the same ones as used in the field measurements, were mounted at 7 m along the flume, two at 20 cm in from opposite sides of the flume at 18 cm above the bed; and the third 20 cm in from the side and at 38 cm above the bed. The sensors were linked to Rustrack recorders and Squirrel data loggers, which recorded variations to the nearest half a percent on the monitor scale. A USDH-48 sampler was used to collect a 9 grid-point sediment profile across the flume with a Manning pumping sampler taking samples every 4 hours. These samples were used in a gravimetric determination of the flume sediment concentration. Samples were filtered, in a two-phase process, through pre-weighed Whatman 42 filters (2.5 microns retention) then through pre-weighed Whatman GFA filters (1.6 microns retention).

These papers were oven dried at 105 ^{o}C and reweighed.

The experiment was divided into two stages: Stage 1 concentrations of 100 - 300 mg l^{-1} and Stage 2 concentrations 300 - 1200 mg l^{-1}, with concentrations being increased in step increments (Figure A.1). Calibration of instruments, using formazin standards, was carried prior to Stage 1, between the stages and at the end of the experiment (Figure A.2).

Results for the experiment were consistent for the three sensors with very little difference with depth, therefore, results from one sensor only are presented. The principal findings were :

1) Although the calibration between turbidity (%) and formazin calibration appeared to be linear, drift over time was evident (Figure A.2, A to C). If not corrected for, this would lead to an overestimation of the actual concentration. The same problem was noted by van Rijn and Schaafsmann (1986) who also used Partech meters.

2) The formazin calibration consistently over-estimated the sediment concentration determined by USDH-48 sampler mean sediment profile samples or Manning pump samples averaged over time (Figure A.3). The error could be as great as 49% at concentrations greater than 1000 mg l^{-1}. The USDH-48 9 point profile averages were always smaller than the manning pumping sampled, time-averaged concentrations. This is thought to be due to the temporary re-entrainment of sediment particles during the purging cycle of the sampler, therefore leading to a slightly higher concentration.

3) Variations in ambient light also produce variations in the turbidity record which are unrelated to variations in sediment concentration. A comparison of an 'artificially generated' day/night shading curve and an actual trace from the Bas Arolla proglacial stream

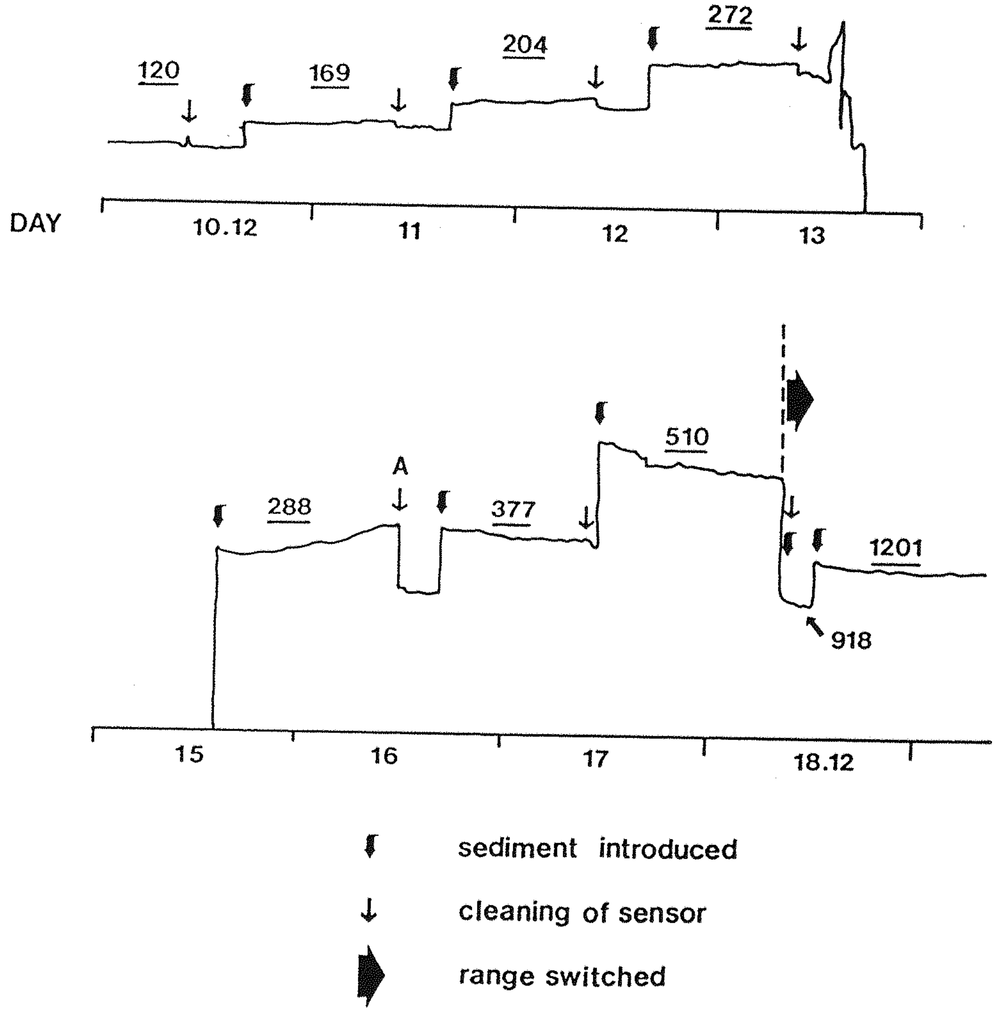
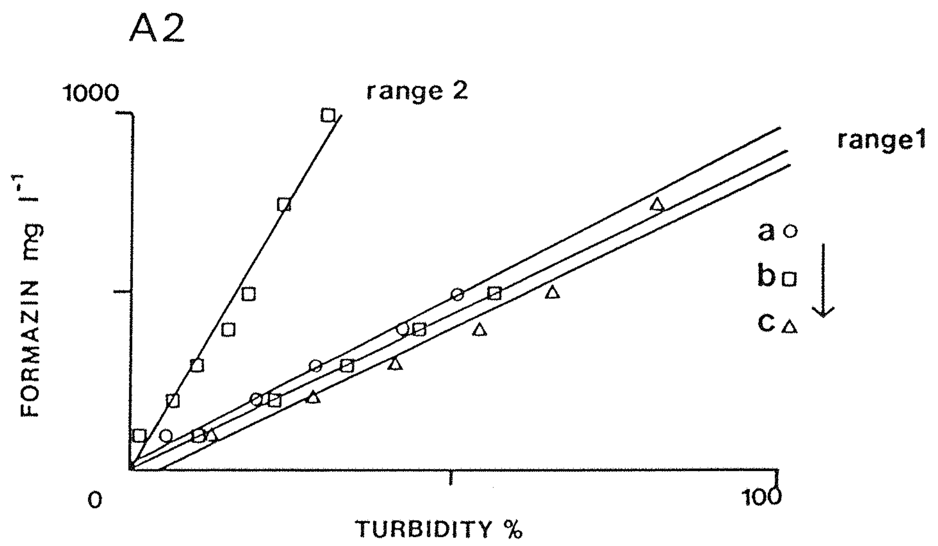


Figure A1 Sediment concentration changes and cleaning during the December 1987 flume study. Underlined values = mean suspended sediment concentrations determined from a 9-point sampling grid (sampled with a USDH-48 sampler). Timescale (x) is in December days.



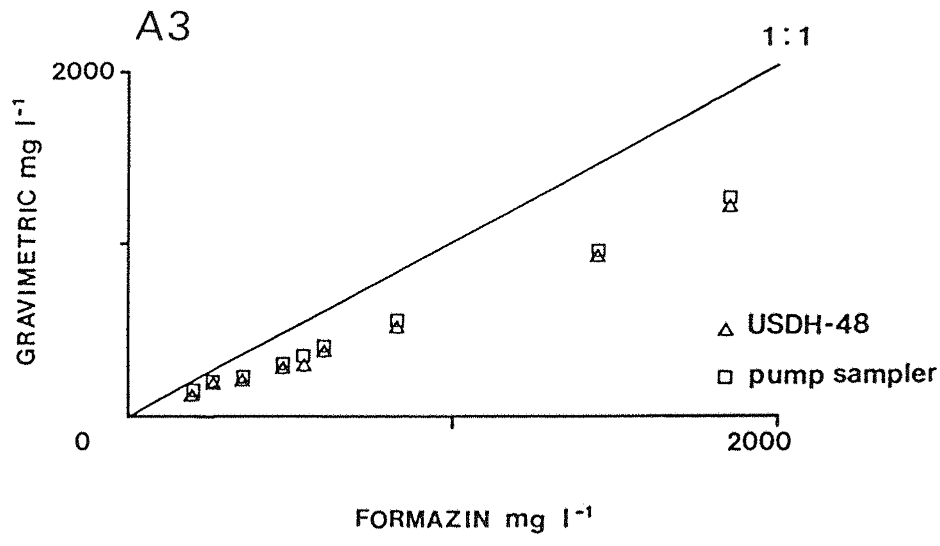


Figure A2 Calibration of Partech turbidity monitor using Formazin standards. Recalibration over time shows drift in the monitor. Calibrations are linear for both ranges 1 and 2.

Figure A3 Comparison of formazin concentrations and concentrations determined by filtering and gravimetric methods (see text). Sampling involved use of a USDH-40 sampler and a pump sampler.

(Figure A.4) clearly demonstrates this effect. The characteristic 'hump-backed' profile, of the kind shown in Figure A.4, is indicative of this effect. However this can be corrected for and it is only apparent at concentrations less than approximately 300 - 400 mg l^{-1}. This is supported by evidence from a flume test, since no response in turbidity was recorded when a 100 watt lamp was shone directly at the sensor submerged in water with a suspended sediment concentration of 377 mg l^{-1}.

4) Deviations in the recorded turbidity value were also caused by lens fouling. The majority of lens fouling caused only minor deviations 2-4 % for the flume and 0-1 % for the proglacial stream (Figure A.5). However, occasionally marked deviations can occur (see for example point A on Figure A.1). The exact mechanism of lens fouling is unknown but it is probably related to the growth of micro-organisms (algae) over the lens leading to partial extinction of the beam. In proglacial streams this effect seems to be less than in the flume probably due to the colder conditions (0-1 ^{o}C), low organics and coarser sediments which will, to some extent, abrade the lens clean. Estuarine silt with it's higher clay content may adhere to the lens more freely. The large discrepancy of 20 % (before and after cleaning) at the start of the second flume run is an anomaly however which may have resulted from exposure/disturbance of the lens during cleaning of the flume.

3. Field Siting of Turbidity Meters

Some basic guidelines should be observed when siting the turbidity instrument in stream channels:

1) The sensor should be fixed so there is free-flow of water between the lenses.

2) If a light screen is used this should not inhibit the free flow of water.

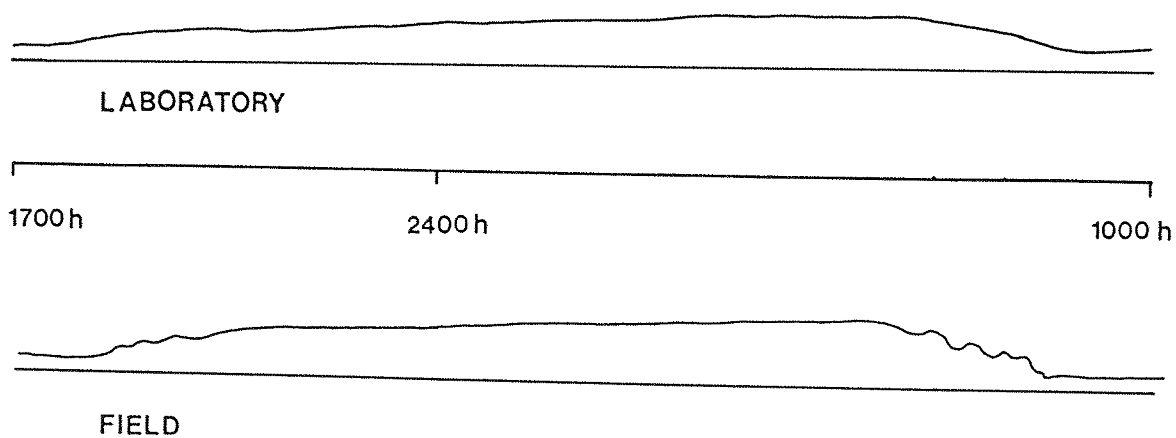


Figure A4 Comparison of field and Laboratory generated turbidity curves showing night/day shading cycles.

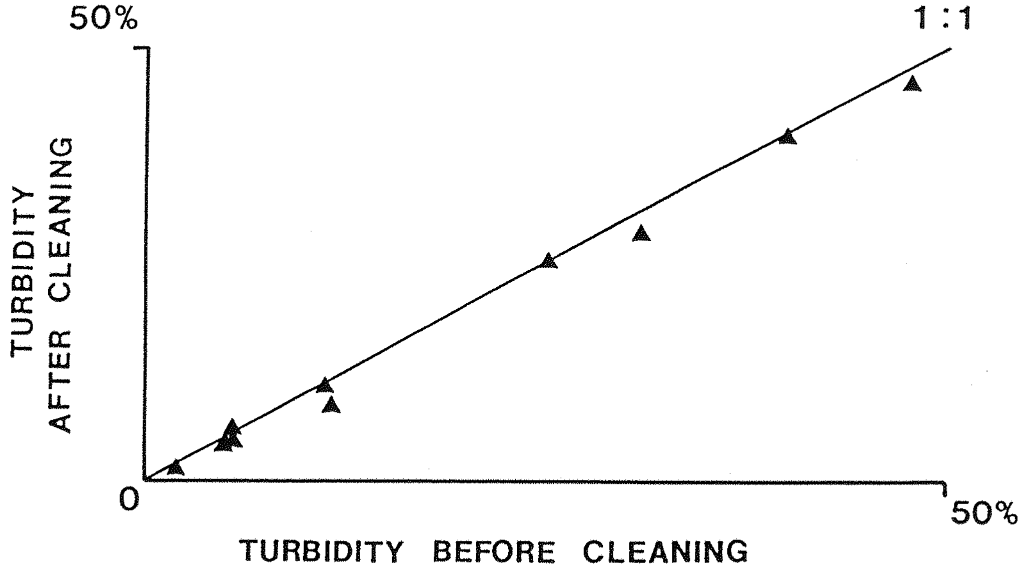


Figure A5 Turbidity meter readings (%) before and after cleaning of the sensor head. Cleaning in the field lasted approximately 4 minutes between removal from the stream and remounting back in the flow.

3) The probe should be positioned with lenses vertical to prevent sedimentation on the lens surfaces.

4) The sensor should be positioned in an area which is free from air bubbles and excessive turbulence.

5) The site should be representative of the stream cross-section

6) The site should be easily accessible for regular cleaning.

7) The sensor head should be mounted at an optimal height to account for fluctuations in stream stage and bed elevation.

Difficulties are encountered when using turbidity meters in proglacial streams because of the high sediment load and dynamism of the channel structure. For example it is not feasible to use a light-shield around the sensor since sediment accumulates on top or inside the shield. This is potentially difficult to resolve because at low concentrations ambient light effects the turbidity (see previous section).

4. Turbidity/Suspended Sediment Concentration Relationships

In addition to the problems of calibration discussed in the flume study turbidity records were also affected by operational problems in the field, including electrical supply problems, chart jams, chart pen sticking, inappropriate scaling and insufficient calibration samples. These problems are discussed in detail in Chapter 3. However using the siting guidelines outlined above most of these problems can be overcome and generally good relationships were established between turbidity and suspended sediment concentration (Table A.1).

Table A.1 summarises the main calibration curves used at the three sites during 1986 and 1987. In all but one of

Table A.1 Regression equations for turbidity - Suspended sediment concentration relationships

EQUATION	R^2	SCALE	RANGE mg/l
Bas Arolla 1986			
SSC = 1.74 x TURB + 71.59	0.97	1	0 - 246
SSC = 3.89 x TURB - 14.30	0.94	1	0 - 375
SSC = 27.56 x TURB - 292.10	0.95	1	0 - 2464
SSC = 5.87 x TURB + 15.84	0.98	2	0 - 603
SSC = 43.35 x TURB - 279.20	0.97	2	0 - 4043
SSC = 14.6 x TURB + 139.84	0.99	3	0 - 1600
SSC = 21.38 x TURB + 70.12	0.89	3	0 - 2208
SSC = 64.90 x TURB - 198.32	0.90	3	0 - 6292
Tsidjiore Nouve 1986			
SSC = 5.95 x TURB - 166.19	0.81	2	0 - 429
SSC = 15.21 x TURB - 312.73	0.92	2	0 - 1208
SSC = 32.87 x TURB - 593.93	0.90	2	0 - 2693
SSC = 29.04 x TURB - 244.24	0.93	3	0 - 2660
SSC = 82.31 x TURB - 807.60	0.94	3	0 - 7423
SSC = 71.46 x TURB - 372.70	0.95	3	0 - 6773
Haut Arolla 1986			
SSC = 21.78 x TURB - 33.22	0.79	1	0 - 2145
SSC = 27.60 x TURB + 22.74	0.85	2	0 - 2783
SSC = 62.15 x TURB + 234.92	0.81	3	0 - 6450
Bas Arolla 1987			
Lower station			
SSC = 4.23 x TURB + 26.14	0.86	1	0 - 399
SSC = 30.08 x TURB + 22.48	0.67	2a	0 - 1030
SSC = 201.77 x TURB - 5577.46	0.88	2b	1030 - 14600
SSC = 450.88 x TURB - 3469.20	0.95	3	0 - 41619
Upper station			
SSC = 26.45 x TURB + 157.73	0.74	2	0 - 2802
SSC = 169.47 x TURB - 528.16	0.76	3	0 - 16419

the calibration curves simple linear regression provided a good-fit to the data. The only exception was for the Bas Arolla lower site 1987 where marked non-linearity was evident in the scatter of the calibration data. Two procedures were used to fit a calibration curve to the data, the first, log-linear regression, provided large overestimates of suspended sediment concentrations at higher turbidity values. The second method involved the fitting of two separate linear regression equations to two sub-populations using an iterative procedure. The optimum combination of the two regression equations was determined when the combined error sum of squares was minimised. This procedure is akin to regression using a 'full-model' dummy variable analysis (Silk, 1976) where no constraints are placed on either slope or intercept terms. This procedure is justified since the aim is to provide the best-fit to the observed data i.e. the best prediction of suspended sediment concentration (Y) from turbidity (X). No strict a priori relationship is assumed except that turbidity is positively related to suspended sediment concentration. However, given the range over which the meter is adjusted it is likely some form of curve or dog-leg relationship would best-fit the data since at higher concentrations the sand fraction becomes an important element of the suspended load and under these conditions turbidity is no longer directly proportional to the concentration in suspension (Fish, 1983). Which explains the need for differences in slope in the relationship.

The replication of some ranges in the calibration curves is due to initial inappropriate setting of the span (i.e. the maximum concentration setting was too low). In these cases ranges were recalibrated and new relationships set-up e.g. Bas Arolla 1986.