SEDIMENT SORTING IN THE GRAVEL-SAND TRANSITION ALONG RIVERS: A FIELD AND MODELLING INVESTIGATION

Daniel John Bloomer

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Department of Geography,

University of Sheffield

Abstract

Gradual downstream fining in gravel bed rivers is often followed by a spatially rapid switch to a sand bed. These gravel-sand transitions (GSTs) can occur in rock types where abrasion rates are low. A common factor is declining shear stress towards base level causing size selective sorting. This study investigates the characteristics of visually abrupt GSTs in two streams of different size and morphology (Allt Dubhaig, Scotland and Vedder River, British Columbia, Canada).

A one-dimensional numerical model of width-averaged size selective gravel sorting is enhanced to simulate gravel-sand mixtures. The updated model fails to generate a GST unless an abrupt break of slope is specified at the start of a run, although only one of the fieldsites exhibits this feature today. This finding suggests that additional processes are crucial to initiate a GST.

A qualitative method of assessing bed surface facies is developed. This is shown to be quantitatively accurate in predicting the bed surface sand range and, when combined with bulk bed grain size distributions (GSDs), indicates a that threshold exists for gravel bed sand content. Above this threshold the channel bed facies switches from gravel framework to sand matrix causing a non-linear relationship between bulk and areal sand content.

Laterally-distributed sampling shows alternation in width-averaged GSD along Vedder River above the GST, with gravel bar samples having higher D_{50} and lower sand proportion than those between bars. The channel bed exhibits a sandier GSD above the GST than would be indicated by inspection. The drop in D_{50} and increase in sand proportion across the GST is of similar magnitude to that associated with bars upstream although the change in grain size is extremely abrupt in surface appearance. Beyond the last gravel bar there is a much greater lateral variability in facies than either upstream or downstream. Point sampling of GSDs, which tends to be done on bars, may be inadequate to characterise the GST or positively misleading.

Evidence from subsurface probing investigations and bed surface sedimentology indicates a slowly prograding gravel front. The position of the front is dependent on near-bed hydraulics. A fine-gravel tracer experiment shows that the transport of these sizes in the GST reach is size selective, although this is not the case in the distal gravel reach.

Field characterisation indicates that the crucial processes missing from the model include: the overwhelming of a gravel framework bed by sand, as the threshold for sand storage is approached, leading to an increased availability of sand on the bed surface; and the lateral sorting of sediment into patches of different ambient grain size, further increasing the availability of the fine fraction.

Contents

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Chapter 1. Introduction

1.1 Background

Rivers provide a means of transporting mass and energy through a channel in the form of water and sediment. The characteristics of the sediments present on the bed of the river can vary along the channel. The traditional image of rivers in upland regions is as steep, rapidly flowing streams with coarse boulders on the bed. In lowland areas they are often viewed as more slowly flowing watercourses with shallow gradients and fine sands on the bed as they approach base level. Lateral inputs of coarse or fine sediment, the wearing down of grains and changes in channel slope can all alter the grain size characteristics of the sediments that are entrained from or deposited to the bed. The rate of change of dominant bed grain size is therefore not constant along a stream.

The size of sediments present on a river bed affects the channel gradient and flow characteristics by influencing the amount of sediment removed from the bed and drag exerted on the flow. These factors control the depth of flow and therefore influence the flood risk exerted by a river. The particular range of grain sizes present on a channel bed, together with the discharge characteristics and nutrient content of the water, control the ecological assemblage present in the stream. It is therefore important to understand the main factors controlling the change in bed grain size along a river channel.

The present research aims to elucidate the most important forms and processes occurring in contemporary streams as their channel beds change from having gravel (> 2 mm diameter) to sand (< 2 mm diameter) as the dominant sediment size. This change in bed sedimentology, the Gravel-Sand Transition (GST), can occur over a relatively short distance compared to the rate of fining of the gravels upstream, without an input of a large volume of fine sediment from a lateral source.

1.2 Downstream fining

The general reduction in bed sediment grain size with distance downstream is termed downstream fining. The investigation of this phenomenon in river gravels has a long history in the fluvial literature, hence the processes involved are reasonably understood (see detailed studies by Ferguson *et al*, 1998; Seal *et al*, 1998). The causes of the overall reduction in bed grain size include lateral inputs of fine sediment from a variety of sources, the wearing of grains in motion and those over which they are transported (abrasion), size-selective sorting (through entrainment, transport and deposition), weathering of exposed grains on the bed surface and landscape history. The three main processes operating to cause a reduction in bed grain size in a contemporary stream are lateral inputs of sediment, abrasion and selective sorting.

1.2.1 Lateral inputs

Lateral inputs of sediment that is finer than that of the channel can come from many different sources. Rice (1998;1999) noted that there was an association between lateral inputs of sediment and discontinuities in the trend of grain size reduction in the main channel. Types of lateral input include tributary confluences (Church and Kellerhals, 1978; Knighton, 1980; Ichim and Radoane, 1990; Brewer and Lewin, 1993; Rice, 1998;1999), tributary fan contacts (Dawson, 1988), outcrops of non-alluvial sediments (Werritty, 1992; Rice, 1999) and mass movements from slopes into the channel (Brierley and Hickin, 1985). A discontinuity of the fining trend can only be caused when a sufficient volume of sedimentologically distinct grains is supplied to the channel. These lateral sources are recognised to generally cause a coarsening of the bed sediment texture (Rice, 1998). However, decreases in bed grain size have been noted (Andrews, 1979; Knighton, 1989; Sambrook Smith and Ferguson, 1995) although these cases are relatively rare. Rhoads (1989) attributes this to the fact that the input of fine sediment must be sufficiently voluminous to alter the competence and capacity of the main channel if the finer sediment is to remain on the bed surface. If only a little fine sediment is added this would be removed from the bed due to the channel's competence to entrain finer sediments than those already dominating the bed (Rice, 1999, pers comm).

1.2.2 Abrasion

Abrasion is a term which describes a wide range of erosion processes at the grain scale, including breakage, grinding and chipping, which mechanically reduce the size of a given clast (Kuenen, 1956). These processes occur to particles not only during transport but also while they are stationary on the bed surface as other grains are transported over them (Brewer *et al*, 1992). The combined effects of lift and drag can also cause particles to vibrate on the bed surface leading to further wear and size reduction (Schumm and Stevens, 1973). Parker (1991a,b) attempted to model abrasion theoretically by concentrating on the collisions between bedload in transport and particles on the bed surface. However, the size of sediment supplied to a stream, its lithology and the degree to which it was weathered prior to its movement by the stream all influence the rate of abrasion (Bradley, 1970; Wolcott, 1988; Werritty, 1992; Kodama, 1994a; Jones and Humphrey, 1997). The range of sediment sizes present in a channel can also affect the rate of abrasion (Kodama, 1994a) indicating that this method of size reduction may be most important in the upper reaches of a channel where the range of grain sizes is larger. In this region, grains supplied to the channel may be heterogeneous, angular or fractured and therefore prone to higher abrasion rates than those that have been transported some distance by the channel (Adams, 1979).

1.2.3 Selective sorting

Selective, or hydraulic, sorting occurs through selective entrainment, differential transport and selective deposition of different sized grains. This sorting occurs by larger particles moving shorter distances, or less often, relative to smaller particles, because higher near-bed flow velocities are required to entrain them. Parker *et al* (1982), Andrews (1983), and others, indicated instead that all sizes in a sediment mixture may be entrained at the same flow strength and the sediments would therefore exhibit equal mobility regardless of their size. This was thought to be achieved through a combination of the hiding of finer grains in interstices in the bed and the protrusion of coarser sediments into the flow. Bradley *et al* (1972), however, after laboratory tests, found that only 10% of the downstream fining exhibited by the gravels found in an Alaskan stream could be caused by abrasion, with the remaining

90% attributed to selective sorting. This is an extremely high proportion that is unlikely to be accounted for if all sizes were entrained at the same discharge. Brewer and Lewin (1993), also during laboratory tests, agreed that abrasion processes were unlikely to account for the majority of the reduction in mean grain size with distance witnessed in many rivers. Clearly, therefore, in some streams the bed sediments do not follow the hypothesis of equal mobility. Hoey and Ferguson (1994) found that rapid downstream fining could be produced by a sediment mixture exhibiting only a slightly size-selective tendency if the river long profile was strongly concave. Even with the assumption of equal mobility, Paola and Seal (1995) proposed that downstream fining could occur through the lateral sorting of finer and coarser sediment mixtures across the channel width. In this case sediments from a finer patch could be entrained at a lower flow than those in a coarser patch while still conforming to equal mobility within each sedimentary patch.

1.3 The Gravel-Sand Transition (GST)

In gravel-bed rivers the gradual downstream reduction in grain size is often followed by a switch to a sand bed over a relatively short distance compared to the rate of fining upstream. Sambrook Smith and Ferguson (1995) termed this phenomenon a Gravel-Sand Transition (GST) and this terminology will also be employed for the present research. Parker and Cui (1998) and Cui and Parker (1998) referred to this spatially rapid grain size switch as a gravel front, and Howard (1980;1987) called it a threshold between gravel and sand bed channel types. It is unclear whether the formation of a GST is simply the extension of one or more of the processes causing gradual gravel downstream fining or whether other controlling factors are involved. This uncertainty is due to the fact that although some studies have noted the existence of accelerated downstream fining with distance between a gravel and a sand bed along a given reach, (for example Yatsu, 1957; Knighton, 1980; Shaw and Kellerhals, 1982; Dawson, 1988; Ichim and Radoane, 1990; Ferguson and Ashworth, 1991), detailed discussion about the GSTs themselves rarely occurs. Only a handful of sources attempt to elucidate the forms and processes responsible for initiating a GST.

The first detailed collection of the general features of GSTs was presented by Sambrook Smith and Ferguson (1995). Three possible causes were postulated: local base level control, abrasion or breakdown of fine gravel, and an excess supply of sand. It was noted that transitions often involved a change in bed surface sedimentology from unimodal gravel, through a bimodal gravel-sand mixture, to dominantly sand-sized sediments. Also common in the region of many GSTs was a sharp reduction in bed slope in the downstream direction. This break of slope and the sedimentological changes were thought to be indicators of a natural feature in fluvial systems that is geographically widespread.

Sambrook Smith and Ferguson (1995) attributed the rapid reduction in channel slope associated with some GSTs to the river approaching a base level, such as a dam, lake, debris fan or main channel. These features induce deposition by forcing a reduction in bed slope and therefore sediment flux. This results in a reduction of the dominant bed grain size downstream by selective deposition of coarser sediments from bedload and suspension (Sambrook Smith and Ferguson, 1996; Dade and Friend, 1998). It was felt that if only sand sizes were mobile these would clog the pores in a gravel bed creating a sand dominated bed downstream in only a short distance (Sambrook Smith and Ferguson, 1995). The effect of the infiltration of fine sediments into gravel beds has been noted by other authors (for example Beschta and Jackson, 1979; Carling, 1984; Peloutier *et al*, 1997). The infiltration rates and variations in grain size distribution of fines deposited onto a gravel bed are complex, related to the supply of sediment, the transport mechanism, local hydraulics, the dimensions of the interstices in the gravel matrix, gravel bed dynamics during flood events, and the reach morphology (Frostick *et al*, 1984; Reid and Frostick, 1985; Sear, 1993). Although these factors are recognised separately, a holistic view of the relative importance of the factors is still to be satisfactorily evolved. Pickup (1984) suggested that rising sea level could cause a break of slope in a channel by effectively creating a backwater zone stretching up the river valley. In this situation, gravel sediments would be left as a lag deposit on the channel bed and only sand sizes would be mobile, again creating a GST.

Yatsu (1955;1957) argued that GSTs may be caused by the tendency for some fine gravel lithologies to be weathered to sands, on hillslopes or bars, or to be worn into sand-sized particles by abrasion. His work on large rivers in Japan suggested that this processes was the main cause of the switch to a dominantly sandy bed. This hypothesis was supported by Ichim and Radoane (1990), although neither study supplied any direct evidence that abrasion was the dominant process. Shaw and Kellerhals (1982) suggested that abrasion of fine gravels may be accelerated because these sizes can be preferentially transported over a smooth sand bed, thereby creating additional sand. The laboratory experiments of Kodama (1994a), however, indicated that abrasion processes may be most prevalent in large, high-energy rivers and are therefore unlikely to be the dominant control on grain size in lower-energy sand-bed channels, as proposed by Shaw and Kellerhals (1982).

Campbell (1970;1977) introduced the possibility that an excess supply of fine sediment could, in some cases, initiate a GST. In his work, on rivers which erode large sources of lateral sandy sediments, he found that the input of sand from this erosion could clog and bury a river's gravel bed. In these cases a break of slope was not found as the extra sediment load was traded off against its lower grain size. Pickup (1984), Higgins *et al* (1987) and Knighton (1991;1999) showed that transitions could also be caused by large volumes of sand-sized mine waste being deposited in a stream.

Whichever of the three causes is responsible for initiating a GST, Shaw and Kellerhals (1982), Sambrook Smith and Ferguson (1995) and Ferguson *et al* (1998) noted that the bed sediments upstream of the transition were often bimodal, with peaks in the medium gravel and sand sizes, and a relative dearth of the intervening material, creating a grain size gap. They suggested that an understanding of the causes of bimodality would lead to a better elucidation of the processes responsible for creating a GST. Three mechanisms were proposed for the creation of a bimodal sediment mixture: preferential entrainment of the grain size gap sediments between the modes, preferential breakdown of grain size gap material, and the influence of sediment supply on stream bed sedimentology. Once bimodal sediments are present on the bed surface they may be organised into distinct gravel and sand zones (Iseya and Ikeda, 1987; Ikeda and Iseya, 1988; Kuhnle and Southard, 1988; Ferguson *et al*, 1989; Paola and Seal, 1995; Wilcock, 1998). In these cases sand can be entrained from the finer patches at a lower flow than would be required if all sediments present were well mixed, thereby increasing the sand flux downstream. Bed slope may be reduced in association with the decreased bed roughness, rendering the gravel patches immobile and increasing the sharpness of the GST (Sambrook Smith and Ferguson, 1995; 1996).

1.4 Numerical modelling of the GST

There have been several attempts to simulate the GST through numerical modelling by different research groups. Hoey *et al* (unpublished) employed the model of Hoey and Ferguson (1994). This one-dimensional sediment routing model (SEDROUT) was shown to produce reasonable simulations of the downstream fining of gravels. The predicted bed grain sizes for the distal part of the simulated stream, however, were finer than those observed in the prototype. The model requires development to include sand sizes to allow the simulation of a GST (Hoey *et al*, unpublished). There is, however, no published work regarding the model's application to gravel-sand mixtures to date.

Robinson and Slingerland (1998) employed the one-dimensional MIDAS model of Van Niekerk *et al* (1992) to test the sensitivity of downstream fining to a number of variables while investigating facies belt development in ancient fluvial systems. The study attempted to take the changing bed sedimentology and channel morphology associated with a GST into account when modelling the system but the transition was not mentioned explicitly in the discussion of the simulation results.

Parker and Cui (1998) and Cui and Parker (1998) outlined the development of a model to specifically simulate a non-migrating GST. The model assumes that two processes could cause a stationary GST: abrasion of gravel, or basin subsidence upstream of a base level. As a result of the first assumption, once a gravel grain was reduced to a particular size it spontaneously broke down into sand, in effect forcing a GST. These papers assumed GSTs were transient features unless there is a specific mechanism that arrests their progradation towards a base level.

Gasparini *et al* (1999) simulated downstream fining through selective transport for an entire channel network in a river basin. Using the GOLEM model of Tucker and Slingerland (1997) downstream fining emerged as a natural dynamic adjustment to the variables simulated even under conditions of uniform grain size distribution in the sediment flux. The simulated transition from a gravel to a sand bed, however, did not include the sharp reduction in bed grain size witnessed in many field situations. The transition only occurred in simulations where the channel bed surface was eroded and a sand-dominated subsurface sediment had been specified at the start of the run.

1.5 Thesis aims and scope

1.5.1 Aims and objectives

The overall aim of this thesis is to elucidate the forms and processes that occur in the field when a river switches from a gravel-dominated to a sand-dominated bed. This aim will be tackled through three supplementary objectives. Firstly, the characteristics of contemporary GSTs are characterised in detail. A second objective is to outline how these transitions change over timescales of the order 10^0 to 10^2 years with a view to assessing potential causal mechanisms. The third and final objective is to investigate whether a GST can be simulated through selective sorting alone by a one-dimensional numerical model.

1.5.2 Scope

Several factors which may be important in causing downstream fining and a GST in some rivers are not considered in detail in the current research. These include: network analysis and the importance of lateral inputs, abrasion in the GST zone, and detailed hydraulic investigations on gravel and sand beds. Although these aspects will be discussed, original research will not take place. It is also important to note that the current research is not aiming to develop an accurate numerical model of GST formation and evolution, rather the research is attempting to use a limited one-dimensional selective sorting model as a tool to elucidate the important processes. If the model fails to simulate a GST then field investigations can be used to outline which forms and processes are missing from the model and therefore indicate the importance of these in creating a GST.

1.5.3 Model considerations

An awareness of the physical basis of any numerical model employed in research is critical as this will lead to an understanding of the inherent limitations of the model related to assumptions made during its development (Lane, 1998). Numerical models are a simplification of reality and for this reason they cannot be used on their own to reliably predict relationships between isolated parts of the natural system to which they are being applied. A numerical model may also be limited by an overall lack of understanding of the system to which it is being applied. Because of this lack of understanding it is often difficult to assign predictive inadequacies to a particular assumption or part of the model structure. Even if the model predicts successfully the outcome of the processes which it is simulating it cannot be taken for granted that the model will hold beyond a specific situation to which it has been applied (Lane, 1998) and the model may be making accurate predictions for the wrong reasons. These facts must be kept in mind when applying numerical models to the natural environment, making it clear that caution is required when discussing their predictions. In some circumstances, for example for complex or poorly-understood systems, a wiser use of numerical models may be to further our conceptual interpretation of a natural system rather than as predictive tools to simulate a number of specific processes. For these reasons a two-pronged approach is undertaken for the current research using computer modelling to supplement field investigations of the systems involved to give a broader understanding of GST processes.

To carry out the numerical simulations the SEDROUT model of Hoey and Ferguson (1994) will be employed. Although some model development is required the structure of SEDROUT makes it the most appropriate choice of the models available. SEDROUT simulates one channel (rather than the network of channels in the GOLEM model), does not simulate abrasion processes (unlike the model of Parker and Cui (1998) and Cui and Parker (1998)), and is easily available. SEDROUT, therefore, does not simulate lateral inputs of water or sediment, which would confuse model interpretation. These facts will simplify the modelling undertaken and ease interpretation. The model will be run for small rivers over a relatively short timescale so that basin tectonics can be neglected.

1.6 Thesis Outline

The thesis is structured in four sections. The first section (Chapter 2) reviews the forms and processes thought to be important for downstream fining and GST development. A more detailed series of aims is presented at the end of this chapter, together with a fuller plan of Chapters 3 to 8.

The second section (Chapters 3, 4 and 5) concentrate on the two fieldsites chosen for further study (Allt Dubhaig, Scotland and Vedder River, British Columbia, Canada) to elucidate the forms, processes and changes over time in GST zones. Chapter 3 outlines previously published research regarding the fieldsites. The methodology, results of investigations and preliminary interpretation of the information collected for this thesis is presented in Chapters 4 (contemporary GSTs) and 5 (GST evolution and channel change).

A third section details the numerical modelling aspect of the research. Chapter 6 will examine the physically-based numerical models that have the capacity to simulate downstream fining and GSTs and identify the model best suited to the current research. Chapter 7 describes the enhancement of the chosen model and details a sensitivity analysis of the modified version. Chapter 8 outlines the attempts to generate a GST through a series of simulations using the enhanced model.

The final section includes a chapter interpreting the results of the current research (Chapter 9). Here the field and model aspects are drawn together to elucidate the important forms and processes present in GST zones. The wider implications of the thesis for further studies in rivers with both gravel and sand sediments present on their beds are also outlined. The final, concluding, Chapter 10 summarises the main findings of the research.

Chapter 2. Sediment sorting in gravel and gravel-sand bed rivers

This chapter will provide a broader context into which the ideas and objectives concerning research about the Gravel-Sand Transition, discussed in the previous chapter, can be placed. The aim of this chapter is therefore to review the current knowledge regarding forms and processes occurring in gravel and gravel-sand bed rivers. This is tackled in three stages: firstly, the processes which are likely to be important in governing the local form, flow and sediment transport at a given distance downstream in a natural gravel-bed channel are outlined; secondly, these processes are generalised to account for changes that may occur down a gravel-bed river; and finally the additional processes associated with an increasing proportion of surficial sand in a gravel-bed river are discussed.

To fulfil the objectives listed above the discussion in this chapter is presented around a series of flow diagrams. These diagrams show the linkages between different forms and processes which can cause substantial feedback in the fluvial system. As each diagram is discussed it forms the structure of an investigation into the areas of research that are crucial for a clearer understanding of the processes involved in GST initiation and evolution. Aspects of the fluvial system that are of particular relevance to the development and evolution of a GST will be highlighted. A literature review allied to the discussion of the diagrams will indicate which of these aspects have not yet been fully investigated. The questions that the current research aims to address and a detailed thesis plan follow at the end of the chapter, building on that presented in Chapter 1, focusing on where the specific results can be found.

2.1 Form, flow and sediment transport in a gravel-bed river at the local scale

The flow diagram below (Figure 2.1) shows the cause and effect relationships operating at a given distance downstream in gravel-bed rivers. These processes have received extensive analysis in the literature and a short review of the points pertinent to the current research is prudent. A discussion of the various aspects follows a brief explanation of the diagram.

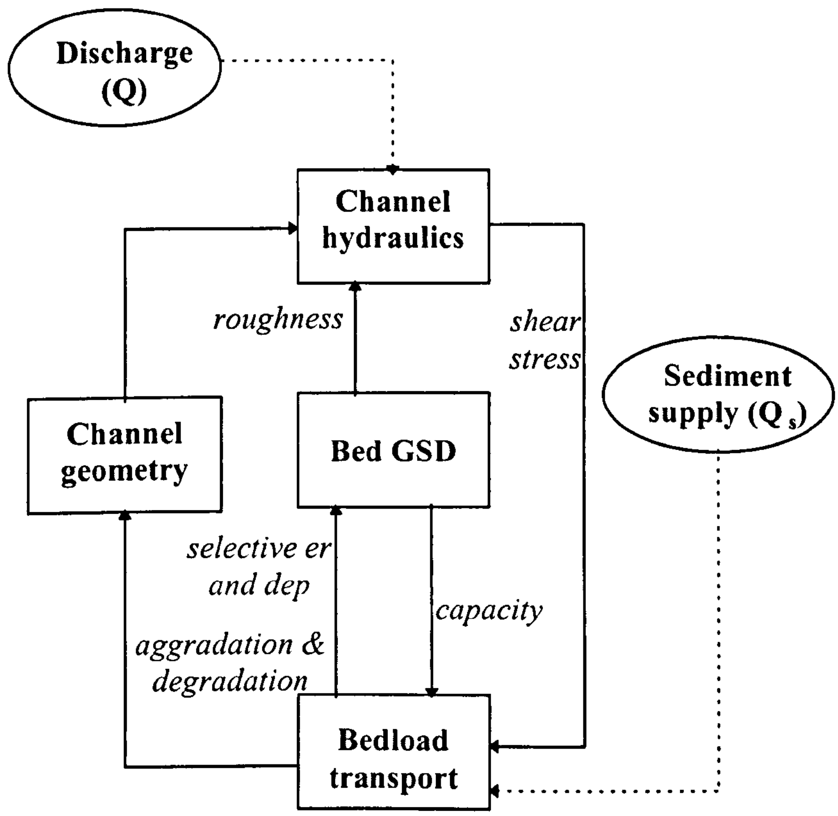


Figure 2.1: Conceptual model of river form and process interactions at the LOCAL scale (after Ashworth and Ferguson, 1986). Solid lines and rectangles indicate the interactions that are occurring locally in the stream. Dashed lines and ellipses indicate supply from upstream. Selective er and dep is selective erosion and deposition, bed GSD is bed grain size distribution.

Ashworth and Ferguson (1986) suggest that Figure 2.1 is best tackled from the top left where unsteady **discharge (Q)** enters a given point in a river which has a non-uniform channel and a rough bed. This can cause a complicated spatial and temporal pattern of water velocity **(channel hydraulics)**. The vertical velocity gradient through the water column at a given point determines the rate of change of the near-bed velocity **(shear stress)** creating a drag force acting on the bed at this point. **Sediment supply (Q_s)** of bedload from upstream is a source of material that can be transported through or deposited in the local reach of interest. The shear stress influences the size and amount of bed material that can be entrained and transported **(bedload transport)**. The balance between sediment supply and transport defines whether **aggradation or degradation** occur and the amount of sediment entrained and deposited locally over time accounts for the mass flux of sediment. The material in transport can act to either maintain or alter the shape and pattern of the channel **(channel geometry)**, through scour, fill or lateral migration. Sediment transport is also affected by the specific bed grain size distribution **(bed GSD)** at this point, together with the way in which the sediments are sorted on the bed surface. The bed GSD is also a control on the availability of sediment for entrainment by the flow. For a given discharge a higher flow velocity can occur either because of a narrower or steeper channel or because the bed is smoother with finer grains and poorly defined bedforms **(roughness)**. **Capacity** is the amount of sediment that the river can transport, entrained from the local channel bed and from upstream. These processes can lead to a change in the bed GSD if they are not in equilibrium. Bed configuration controls flow, which controls transport capacity; if capacity does not equal supply, then the bed will either aggrade or degrade, and it may also coarsen or fine, through **selective erosion and deposition**. The bed GSD and channel geometry directly affect flow properties, and hence bedload transport, through altering the surface roughness of the channel. Grain sizes and shapes, microtopographic features (for examples clusters and imbrication), and large-scale bed undulations (pool-riffle or step-pool sequences) can all cause a change in the surface roughness near the bed of a channel. Changing the bed GSD through selective erosion and deposition may also alter the channel geometry. Erosion of the bed may lead to the preferential entrainment of the finer fractions and a coarsening of the bed. Aggradation can be caused by preferentially depositing the coarser fractions of the load.

2.1.1 Flow and channel hydraulics

Discharge (Q)

The morphology of a gravel-bed river channel is largely conditioned by fluid flow and its interaction with erodible material at the channel boundaries. The variation in fluid flow is dependent on the temporal and spatial pattern of discharge passing through the channel network. The discharge of a reach will influence the width, depth and velocity of flow. When a river is in flood, and at bankfull discharge, any additional discharge is unlikely to lead to much extra sediment entrained and transported, because there is little increase in channel depth. This is due to the flow spreading out over the floodplain. There may also be considerable energy losses through vortices created at the bank tops if these are sharp (Ackers, 1992), although if the water edge were on a point bar less energy would be lost. Various authors have argued that there is evidence from bedload transport sampling to suggest a close relationship exists between the bankfull discharge and the most geomorphologically effective flow (see Andrews, 1980; Andrews and Nankervis, 1995; Batalla and Sala, 1995). For much of the modelling work carried out for the current research (presented in Chapters 7 and 8) a near bankfull discharge was employed during the runs for the reasons stated above.

Channel hydraulics

The in-channel hydraulics of a stream at a given point are influenced by the water depth, slope and roughness of the bed, and also channel geometry. These factors, in turn, influence sediment transport. As discharge fluctuates at a cross-section, the flow variables will also change, causing feedback into other parts of Figure 2.1. At the local scale variation of flow can only be accounted for by changing discharge or conditions in the region of interest. Spatial differences in hydraulics are discussed in section 2.2.1 below. By far the most common type of flow in natural streams is unsteady non-uniform flow, where the depth of water changes from place to place and over time.

Discharge (Q) varies over time at a point (and for a given point in time Q can also vary spatially). Since the following is true:

$$Q = wdv \qquad \text{Equation 2.1}$$

where w and d are the width (m) and mean depth (m) of the wetted boundary of the channel respectively and v is the cross sectionally-averaged flow velocity (m s^{-1}), an increase in Q is always accompanied by an increase in one or more of w, d or v.

Shear stress

The boundary shear stress (τ) is defined as the drag force per unit area acting on the bed and banks of the channel (or the wetted perimeter) in the direction of flow. For uniform steady flow the following relationship holds:

$$\tau = \rho g R S \qquad \text{Equation 2.2}$$

where ρ is the density of water (kg m^{-3}), g is the acceleration due to gravity (m s^{-2}), R is the hydraulic radius (width*depth / wetted perimeter, in m) and S is the water surface slope. Calculating boundary shear stress in this way gives a value averaged over the wetted perimeter.

There are two scales at which to consider shear stress: width-averaged (as in Equation 2.2 above) and locally related to the velocity profile. Flows with non-uniform depths or unsteady discharges can experience local variations in the wetted perimeter-averaged boundary shear stress that would not be predicted using the equation above. Local shear stress depends on the vertical velocity gradient through the near bed water column and this is usually greater towards the centre of the channel where the flow is deeper and faster. In meander bends the maximum shear stress tends to be offset from the centre of the channel, towards the outer bend. One-dimensional numerical models of processes occurring in gravel-bed rivers, (outlined in Chapter 6), do not take these local variations in shear stress into account, potentially undermining their accuracy. Change in shear stress over time, at a point, is caused through a change in depth, and in some cases through a change in channel slope, for example when pools and riffles become drowned out (Thompson *et al*, 1996). The stream beds discussed in the present research consist of cohesionless grains. As the discharge, flow velocity and therefore shear stress over the surface of these grains increases the forces acting to move these particles may exceed the forces resisting motion, initiating sediment transport.

Roughness

The flow velocity is conventionally held to depend on three factors which control the downstream movement of water under gravity: the flow depth or hydraulic radius, slope and flow resistance. There are various controls on the flow resistance, or roughness, in natural channels such as: the grain size, sorting and geometric properties of the bed and bank material; bridge pillars, sewerage outlets and other obstacles; bedforms of various dimensions and shape; vegetation on the bed and banks (particularly important at low flow); large amounts of sediment in transport (Bergeron and Caronneau, 1999); meanders as the line of fastest flow moves towards the outside of the bend; and hydraulic jumps where a rapid change in depth from shallow, fast supercritical flow to deeper, slower subcritical flow creates turbulence and dissipates energy.

An analysis of Equation 2.1 shows that the flow depth in a channel for a given discharge is therefore also influenced by the degree of roughness of the boundaries since a reduction in flow velocity may result in an increase in depth. Several equations exist for this calculation but most are of the general form:

$$V = \text{function of } (d \text{ [or } R\text{]}, S, \text{roughness}) \qquad \text{Equation 2.3}$$

where V is the cross-sectional mean velocity (m s^{-1}), d is flow depth (m), R is the hydraulic radius (m), and S is slope. In the case of Equation 2.3 the roughness factor would be defined inversely as the value that gives the measured V, such that as roughness increases V decreases. The flow velocity can therefore be calculated using a function containing terms for the depth of flow, channel slope and a flow resistance, or roughness coefficient. The most commonly used flow equations following the form above are the Manning and Darcy-Weisbach functions. The development of flow resistance equations lies outside the scope of the present research. The numerical model used for the current research, however, (see Hoey and Ferguson, 1994), employs a modified version of the widely used Darcy-Weisbach function for the calculation of channel roughness. For the purposes of clarity and thoroughness, therefore, it is useful to define this equation here. The flow velocity, V, is calculated using the following function:

$$V = \left(\frac{8gRS}{f}\right)^{1/2} \qquad \text{Equation 2.4}$$

where f is a friction factor (the Darcy-Weisbach friction factor).

As $\tau = \rho gRS$ (see Equation 2.2 above):

$$f = \frac{8\tau}{\rho V^2} \qquad \text{Equation 2.5}$$

The shear velocity (U_*, m s^{-1}) is defined as:

$$U_* = \left(\frac{\tau}{\rho}\right)^{1/2} = (gRS)^{1/2} \qquad \text{Equation 2.6}$$

By combining equations 2.5 and 2.6 it can be shown that:

$$\frac{V}{U_*} = \left(\frac{8}{f}\right)^{1/2} \qquad \text{Equation 2.7}$$

From this equation it can clearly be seen that the Darcy-Weisbach friction factor is dimensionless, unlike the roughness factor used in the Manning function (n). For this reason it is used widely in the fields of fluid friction to assess the degree of roughness of channel boundaries, which defines the force exerted by the flow.

Using Equation 2.4 a calculation of how an increase in width-averaged discharge is allocated between increasing flow depth and flow velocity can be carried out. Generally, depth is lower for a given discharge when channel slope is higher, and velocity increases less fast than depth at a point as discharge increases.

As with shear stress, there are two scales at which to consider roughness: width-averaged as above and locally related to the velocity profile. In turbulent boundary layers the flow velocity usually increases with the log of height above the bed and this is commonly known as the log law. The maximum flow velocity will occur at the water surface if the log law extends that far. The numerical expression for the log law is shown below:

$$\frac{\mathrm{U}}{\mathrm{U}_*} = \frac{1}{\kappa} \ln\left(\frac{z}{z_0}\right) \qquad \text{Equation 2.8}$$

where U is the point velocity (m s^{-1}) at height z above the bed surface (m), κ is the von Karman constant (approximately 0.4), and z_0 is the roughness height (m), or the height above the bed surface at which the flow velocity is zero due to friction with the bed.

If the log profile holds throughout the water column, the depth-averaged mean flow velocity ($\overline{U}$) can be written as:

$$\overline{U} = \frac{U_*}{\kappa} \ln\left(\frac{d}{ez_0}\right) \qquad \text{Equation 2.9}$$

where e is the base of natural logs (approximately 2.718).

Nikuradse's experiments with roughened pipes suggest:

$$z_0 = \frac{\mathrm{k}_s}{30} \qquad \text{Equation 2.10}$$

where k_S is approximately equal to the D_{50} (median grain size) of the sand on the pipe walls. Contention, however, surrounds the relationship between k_S and the poorly sorted river gravel or gravel-sand mixtures that occur on the beds of many alluvial channels. For these sediments z_0 may not have such a clear relationship to bed D_{50} (k_S) since larger gravel grains will protrude further into the flow compared to sand grains. The form drag on these obstacles increases the overall skin resistance.

Combining Equations 2.9 and 2.10:

$$\frac{\overline{U}}{U^*} = \frac{1}{\kappa} \ln\left(\frac{30d}{ek_S}\right) \qquad \text{Equation 2.11}$$

An analysis of equations 2.7 and 2.11 shows that $(1/f)^{1/2}$ varies as the log of d/k_S, or relative smoothness.

The equation above can also be written:

$$\frac{\overline{U}}{U^*} = A + B \ln\left(\frac{d}{k_S}\right) \qquad \text{Equation 2.12}$$

where A and B are constants. This is often called the Keulegan (1938) equation, if it is assumed to apply to the whole cross-section and not to one vertical velocity profile.

As noted above there is much debate about how k_S relates to different parts of the GSD of a mixed-sized gravel bed. Bray (1982) discovered that in gravel-bed rivers k_S did not predict roughness when specified as the D_{50}, D_{65} or D_{90} of the bed material. These findings agree with those of other researchers (see Kamphuis, 1974; Burkham and Dawdy 1976; Charlton *et al*, 1978; Hey, 1979) who indicated that to achieve the most accurate calculations of flow velocity in gravel-bed rivers the bed grain size parameter must be multiplied by a constant. Failure to employ this technique can result in errors in the calculated velocity of up to 100% for low values of relative smoothness (Bray, 1982). In flume experiments with a fixed gravel-sand beds

Kamphuis (1974) found the best results for observed versus calculated flow velocity were achieved when k_S was set to equal $2D_{90}$. During investigations into flow resistance in gravel-bed rivers Charlton *et al* (1976) found this value to be $3.5D_{90}$ and Hey (1979) suggested a value of $3D_{84}$. The model of Hoey and Ferguson (1994) specified k_S as some multiple of D_{84} of the bed grain size.

2.1.2 Sediment transport

Sediment supply (Q_S)

Sediment passing a point is generally supplied from upstream, although may come from a lateral source for example bank collapse, mine waste or tributary input (Knighton, 1989; Rice, 1999). The amount and size of bedload entering will influence the transport and depositional processes occurring. If a large amount of sediment is supplied from bed load then it is likely that the transport capacity will be exceeded and deposition onto the bed will occur. This may, in turn, alter the bed GSD and therefore the roughness of the bed. The shape of the channel cross section may also be affected.

Bedload transport

The shear stress acting on the bed of the channel is important since it defines whether entrainment or deposition of a particular grain size occurs and therefore whether or not the grain is transported (Gomez, 1991). Shear stress also controls the rate of sediment transport. Bedload is defined here as sediment that is in contact with the bed for at least part of a transport event. Sediment can also be transported as suspended load where bed material is held in the water column for the majority of the event. Sediment transported in this way is discussed further in Section 2.3.2 below. It should be noted that the size division between bed and suspended load is likely to change over time as discharge varies (Andrews, 2000). An additional mode of sediment transport, that of wash load, which comprises finer sediments derived from hillslopes or upper banks, can also route sediment through the channel system. Sediment transported in this manner, however, is irrelevant to change in the bed GSD as it remains in the water column even when discharge is low, rather than being moved in a series of hops through the channel network when discharge and hence shear stress increase, as in the case of suspended or bedload. In most perennial streams suspended sediment and wash load make up the majority of sediment in transport although sediments transported by these mechanisms are not responsible for changing the channel boundary characteristics in gravel-bed rivers since gravel is rarely transported in this way.

Because of this, much consideration of the transport of sediment at the local scale in gravel-bed rivers concerns bed load. An investigation of all the literature related to bedload transport is an enormous undertaking. For this reason only those processes that may be important in the development of downstream fining and a GST will be discussed in this chapter.

Once shear stress exceeds a certain critical value particles on the bed surface will be entrained into the flow. The competence of a particular flow may be expressed as the largest particle that can be entrained from the stream bed. There are three forces acting on a particle resting on the bed of a channel: (1) the submerged weight of the particle due to gravity; (2) a downstream drag force; and (3) a lift force directed upwards. These can be drawn in vector form (shown in Figure 2.2 below) since they have both force and direction. The fluid force (4 in Figure 2.2) acting to move the grain is the resultant of the downstream drag and the upwards lift.

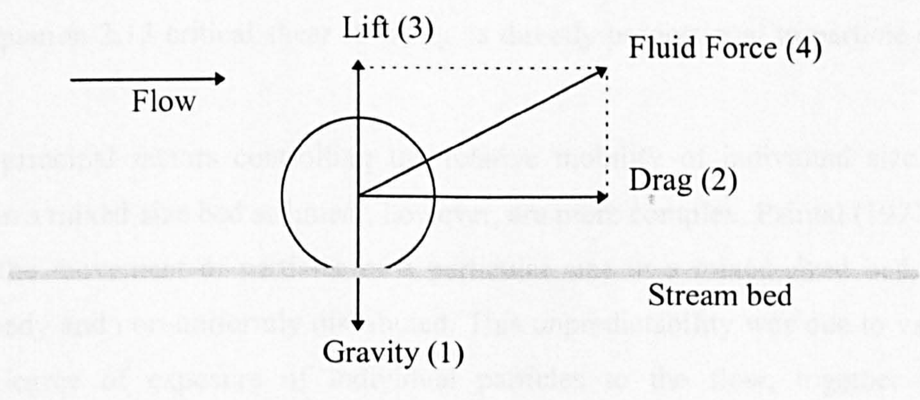


Figure 2.2: Forces acting on a loose grain on the bed of a river channel. Note that this is a simplified two-dimensional plot. The figure does not include scope for a grain to move laterally across a channel, for example down a point bar slip face.

Complicating factors may be introduced into the above diagram when the grain being considered is surrounded on the bed by other grains. These additional grains may act to increase the pivot angle that the grain must climb before it can be entrained (Fenton and Abbott, 1977; Reid *et al*, 1992). The shear stress required to entrain grain is often referred to as the critical shear stress. The forces on a particular grain will fluctuate over time, due to turbulent structures, and therefore have to be time-averaged. This time-averaging is empirically accounted for in the Shields calculation. Shields (1936) found that τ_c (the critical shear stress at the boundary which is responsible for grain movement) was directly proportional to grain-size (or diameter) if lift is neglected (or lift $\propto$ drag is assumed). Using grains of uniform size and shape, Shields calibrated experimentally a value of the proportionality function τ^*_c, and called this the dimensionless entrainment function. This was defined using the following function:

$$\tau^*_c = \frac{\tau_c}{(\rho_s - \rho)gD} \qquad \text{Equation 2.13}$$

where ρ_S is the sediment density (kg m^{-3}) and D is the particle diameter (m). Shields' results suggest that in a hydraulically rough channel, such as a natural gravel-bedded river, τ^*_c reaches a constant value of between 0.03 and 0.06, with 0.045 as an accepted good approximation (Komar, 1988) and therefore, as noted above, according to Equation 2.13 critical shear stress τ_c is directly proportional to particle size ($\tau_c \alpha$ D).

The principal factors controlling the relative mobility of individual size fractions within a mixed-size bed sediment, however, are more complex. Paintal (1971) showed that the movement of particles of a particular size in a mixed-sized bed was both unsteady and non-uniformly distributed. This unpredictability was due to variation in the degree of exposure of individual particles to the flow, together with bed structuring. Fenton and Abbott (1977) investigated entrainment of mixed-size sediments in a series of flume experiments. They discovered that the degree of exposure of a particle to the flow exerted a strong control on the likelihood that it would be entrained. If it is assumed that larger grains protrude further into the flow, by virtue of their size, than smaller ones, for particles between 0.3 and 4.2 times the median diameter of the subsurface sediment, the critical dimensionless shear stress varies almost inversely with grain size (Andrews, 1983). It should be noted that Andrews (1983) was investigating the maximum grain size transported by a given shear stress. Hiding and protrusion effects must therefore be taken into account when considering entrainment of different grain sizes in a mixed-size (or heterogeneous) bed, since the stability of a particle is influenced by its size within the bed GSD (Egiazaroff, 1965; Wiberg and Smith, 1987). This is often done using a function of the form shown below:

$$\frac{\tau_{ci}}{\tau_{c50}} = \left(\frac{D_i}{D_{50}}\right)^{x}$$ Equation 2.14

where D_i is the ith grain size (m) found in the bed, τ_{ci} is the critical shear stress required to entrain that grain size (N m^{-2}), D_{50} is the median bed grain size (m), τ_{c50} is the critical shear stress required to entrain that median grain size (N m^{-2}) and x is a hiding factor.

Equation 2.14 is the simplest and most widely used form of a number of hiding functions that have been suggested. The function indicates that the critical shear stress for the entrainment of a particular grain depends to some extent on its size in relation to the overall bed GSD, or relative grain size (Komar and Li, 1986; Li and Komar, 1986; Kirchner *et al*, 1990). In Equation 2.14 the exponent x lies in the range 0 to 1. If $x = 1$ is specified the simple Shields relationship will be followed with particle entrainment based on grain size. If $x = 0$ is specified then all grain sizes present in the bed will be entrained at the same shear stress and their relative proportions in the bedload will be defined by their proportions in the bed. This phenomenon is termed equal mobility.

The development of a coarse bed surface armour layer a few grains thick with a finer, more poorly sorted mixture beneath (Church *et al*, 1987; Tait *et al*, 1992) in a stream aids the development of equal mobility (Andrews and Parker, 1987; Sutherland, 1987; Parker and Sutherland, 1990). This type of armoured bed may form through interactions occurring at the grain scale as finer grains are removed leaving coarser grains which can only be entrained by the historical maximum flow (Dunkerley, 1990). Only once this coarse surface layer, from which fine sediments have been winnowed during low to medium flows, has been removed, can the finer subsurface sediments be entrained (Diplas, 1987). Dietrich *et al* (1989) suggested, following flume experiments, that a coarse bed armour develops to the extent that sediment supply is less than capacity without armour. Surface coarsening results from increased selective sorting and the winnowing of fines during flows that cannot entrain the framework. Lisle and Madej (1992) argued that a coarse armour layer would not develop in a channel which had high sediment supply through a continual replenishment of fine grains.

In streams where a coarse armoured layer is not present, or is poorly defined, for example ephemeral dryland rivers which experience only low-frequency high-magnitude flows, a full range of bed grain sizes is available to any flow through the system (Laronne and Reid, 1993; Laronne *et al*, 1994; Reid and Laronne, 1995). Bedload transport rates during flood are therefore substantially higher than those of perennial streams.

For most empirical evidence exponent x in Equation 2.14 is greater than 0. This implies that true equal mobility is rarely achieved in natural alluvial sediments and instead the selective entrainment, transport and deposition of bed sediments takes place, with varying degrees of size selectivity depending on a number of factors. These factors include the flow competence, the degree of sorting of the grains both downstream and vertically, and the packing and local sorting of the sediment framework on the bed surface. Ashworth and Ferguson (1989) and Komar and Shih (1992) indicated that size selective entrainment takes place, although not to the extent predicted by Shields (1936). Selective entrainment was also thought by these authors to decrease in importance as discharge increases. Wilcock (1992) showed using flume experiments that the value of x approached 0 as shear stress increases and equal mobility of all sizes may be achieved when shear stress exceeds twice that required for the initiation of grain entrainment from the bed.

Parker *et al* (1982) studied transport rates for different size fractions of bedload in Oak Creek, USA. A relationship was developed between shear stress exerted and transport for each size fraction which was then extrapolated back to a very small transport rate. It was assumed that the shear stress corresponding to this very small transport rate for each grain size was that critical for entrainment. Parker *et al* (1982) found that the exponent x was very near that which would give equal mobility for all sediment sizes. If near equal mobility prevails in a given stream then the main control on entrainment is bed surface D_{50} rather than the size of each individual fraction. Any change in the bed surface sedimentology can cause major changes in transport rates.

For rivers with mixed-sized sediments present on the bed Kirchner *et al* (1990) suggest that the critical shear stress for a particular size should encompass a probability distribution rather than a single value since grains of the same size may be present in both places where they are directly exposed to the flow and also hidden in the interstices between larger grains. This probability distribution will widen with decreasing grain size sorting and increasing bed roughness or D_{50}. The critical shear stress for a particular grain size may also vary by nearly an order of magnitude due to structuring of the bed surface (Church, 1978; Church *et al* 1998).

Some values of the exponent x, when considered as it appears in Equation 2.14, are: 0.128 (Andrews, 1983); 0.18 (Carling, 1983); 0.0 (Andrews and Erman, 1986); 0.36 (Komar, 1987); 0.35 (Ashworth and Ferguson, 1989); 0.0951 (Parker, 1990). Clearly there is considerable disagreement regarding the degree to which grain size controls mobility in a mixed-sized bed. The lack of reliable field data related to bedload entrainment during variable discharges over a wide range of bed GSDs limits the applicability of these hiding functions (Reid *et al*, 1997). Even so, the values for x quoted above indicate that the majority of streams with mixed-size sediments present on the bed tend to exhibit only a small degree of size selectivity at entrainment.

The entrainment and sorting processes that occur in gravel-sand mixtures are more complex than those acting in a purely gravel bed river as the channel bed sediments can have a very wide range of grain sizes and some sizes may be transported in suspension. These processes are discussed in Sections 2.3.1 and 2.3.2 below.

Equation 2.14 is often expressed using the dimensionless entrainment function of Shields which is derived, using Equation 2.13 as follows:

$$\tau^*_{ci} = \frac{\tau_{ci}}{(\rho_s - \rho)gD_i} \qquad \text{Equation 2.15}$$

where τ^*_{ci} is the critical entrainment function for the ith grain size

and:

$$\tau^*_{c50} = \frac{\tau_{c50}}{(\rho_s - \rho)gD_{50}}$$ Equation 2.16

where τ^*_{c50} is the critical entainment function for the bed D_{50}

therefore:

$$\frac{\tau^*_{ci}}{\tau^*_{c50}} = \left(\frac{\tau_{ci}}{\tau_{50}}\right)\left(\frac{D_{50}}{D_i}\right)$$ Equation 2.17

Combining Equations 2.14 and 2.17:

$$\frac{\tau^*_{ci}}{\tau^*_{c50}} = \left(\frac{D_i}{D_{50}}\right)^x\left(\frac{D_{50}}{D_i}\right) = \left(\frac{D_i}{D_{50}}\right)^x\left(\frac{D_i}{D_{50}}\right)^{-1} = \left(\frac{D_i}{D_{50}}\right)^{1-x}$$ Equation 2.18

Rearranging and combining Equations 2.14, 2.15 and 2.16 it can be seen that:

$$\tau_{ci} = \tau^*_{c50}\,(\rho_s - \rho)gD_{50}\left(\frac{D_i}{D_{50}}\right)^x$$ Equation 2.19

If τ^*_{c50} is assumed to equal 0.045 (Komar, 1988) then:

$$\tau_{ci} = 0.045(\rho_s - \rho)g(D_{50})^{1-x}(D_i)^x$$ Equation 2.20

From Equation 2.20 it can be seen that the factor causing size selective entrainment in rivers is the $(D_i)^x$ term. For downstream fining to occur, therefore, x must be greater than zero.

Capacity

The capacity of a stream is the total load of sediment that can be carried or its sediment discharge under varying flow stages. A distinction is often made between supply-limited and capacity-limited sediment transport (Knighton, 1998). If all the material that is supplied to a channel is wash-load it can be transported at almost any discharge. This type of transport is supply-limited, rather than being specified by the strength of the flow (Dietrich *et al*, 1989; Lisle and Madej, 1992). The transport of coarser material is more often limited by the flow strength, or capacity, making it intermittent. The primary relevance of this fact to the present research is that sorting of grain sizes in transport can only take place if the bedload capacity does not equal the supply for particular fractions. If the capacity of the stream equalled the supply of all sediment sizes then the bed GSD would not change with distance along the channel, precluding the formation of downstream fining and a GST. While there are many different bedload transport equations, those applicable to individual size fractions (q_i or the ith grain size) are mostly of the form:

$$q_i \propto F_i \text{ x function of } (\tau - \tau_{ci}) \qquad \text{Equation 2.21}$$

where q_i is the flux of a particular grain size in transport (kg m^{-1} s^{-1}) and F_i is the proportion of that that grain size present in the bed active layer from which sediment is entrained (kg).

Most sediment transport equations, whether based on excess shear stress (as equation 2.21 above), excess stream power, or some other flow parameter, involve a power law meaning that the transport rates of different sized sediments is non-linear. The widely used Meyer-Peter Muller (1948) equation, for example, raises the excess shear stress for each grain size by the power of 1.5. Bagnold's (1966) function which uses excess stream power to calculate a bedload transport rate also employs a 1.5 power law. The bedload transport function used for the modelling in the current research (Parker, 1990) calculates grain size specific bedload transport rates as a function of excess stress. Calculating transport rates on a grain size specific basis allows the development of downstream fining which could not occur if transport rates were calculated for a single size fraction to represent the entire bed GSD. The Parker (1990) bedload transport equation is described in detail in Chapter 7.

2.1.3 Channel change

Erosion and deposition

The balance between sediment supply and flow strength (or capacity, see above) determines whether erosion or deposition occurs. The processes previously outlined are concerned with the entrainment and transport of individual grains. If a large number of grains are entrained or deposited then the geometry of the channel may be significantly altered. Examples of these larger scale processes result from sediment transport variability in space and time (Hoey, 1992) and include scour and fill (see Andrews, 1979), and sediment pulses (Iseya and Ikeda, 1987) or sediment slugs (reviewed in Nicholas *et al*, 1995). Significant progress has been made in understanding which variables are important in governing whether the erosion or deposition of sediment takes place and how bedforms and different bed sediment mixes created by these processes are likely to evolve (Hoey, 1992).

Change in channel morphology can be investigated using the sediment continuity equation. This function assumes that any sediment entrained from or supplied to one part of a river will be deposited elsewhere in the stream. The equation is shown in its simplest form below:

$$\Delta S = I - O \qquad \text{Equation 2.22}$$

where ΔS is change in storage, and therefore bed elevation, I is sediment input, and O is sediment output, from a reach. If $I > O$ the bed aggrades, and if $I < O$ the bed degrades. For channels in equilibrium, $I = O$.

The sediment continuity equation can also be expressed in differential form, shown below:

$$\frac{dz}{dt} = \frac{-1}{1-\lambda}\frac{dq_T}{dx}$$ Equation 2.23

where z is the bed elevation (m), λ is the bed porosity, q_T is the total bedload transport rate ($m^3\ m^{-1}\ s^{-1}$) and x is distance in a streamwise direction (m). For the purposes of the current research this equation must be expressed for individual size fractions. The continuity equation of Parker and Sutherland (1990) can be used to take account of these different size fractions, as shown below:

$$(1-\lambda)\frac{dL_a F_i}{dt} = -\frac{d(q_T p_i)}{dx} + E_i\left(\frac{dq_T}{dx} + (1-\lambda)\frac{dL_a}{dt}\right)$$ Equation 2.24

where L_a is the thickness of the active layer (m) from which sediment is entrained or deposited, and p_i and E_i denote the proportions of the bed material of the ith size class in the bedload and exchange sizes respectively.

A complication is introduced into the equation because a proportion of sediment of a particular size may not be deposited to the active layer or may be transferred from the active layer to the subsurface layer, necessitating the inclusion of the E_i parameter. For this reason a bedload-bed exchange function (for example that of Hoey and Ferguson, 1994 or Toro-Escobar *et al*, 1996) is required. Further details of bedload-bed sediment exchange functions that are relevant to the current research are presented in Chapter 6.

Channel geometry

As noted above channel geometry, or morphology, is influenced by the amount of sediment eroded from or deposited to the bed. Scour and fill or the lateral migration of the channel at a particular point alters its form and this has a potential feedback effect on channel hydraulics and therefore onto many other forms and processes shown in Figure 2.1. Erosion from upstream of the channel bed will decrease the slope, therefore reducing the flow velocity and increasing water depth. The opposite will occur when a channel deposits sediment on the bed. The form of the channel at a point is an expression of the river energy and the resistance of the material present at the channel boundaries (Morisawa, 1985).

Bed grain size distribution (GSD)

The GSD of the bed material at the local scale will specify the amount of different sizes of sediment available for entrainment at a point. The size of the bed material has an impact on the roughness of the channel. This, in turn, creates a feedback which will influence entrainment at a point. As discussed above, complications are introduced when calculating roughness and sediment transport for beds with mixed-sized sediments, associated with different sorting processes occurring in the sediments, related to hiding and protrusion of different grain sizes.

2.2 Form, flow and sediment transport downstream in a gravel-bed river

The following section discusses how processes operating in a river channel alter in a streamwise direction. Many of the processes that vary in this direction have been discussed in the preceding section, although it should be noted that processes varying with distance downstream can create differences and complications that are not apparent when considering the processes operating locally. These processes therefore require further discussion. This discussion follows a brief explanation of how form, flow and sediment transport operate along a gravel-bed river following the structure of Figure 2.3 below.

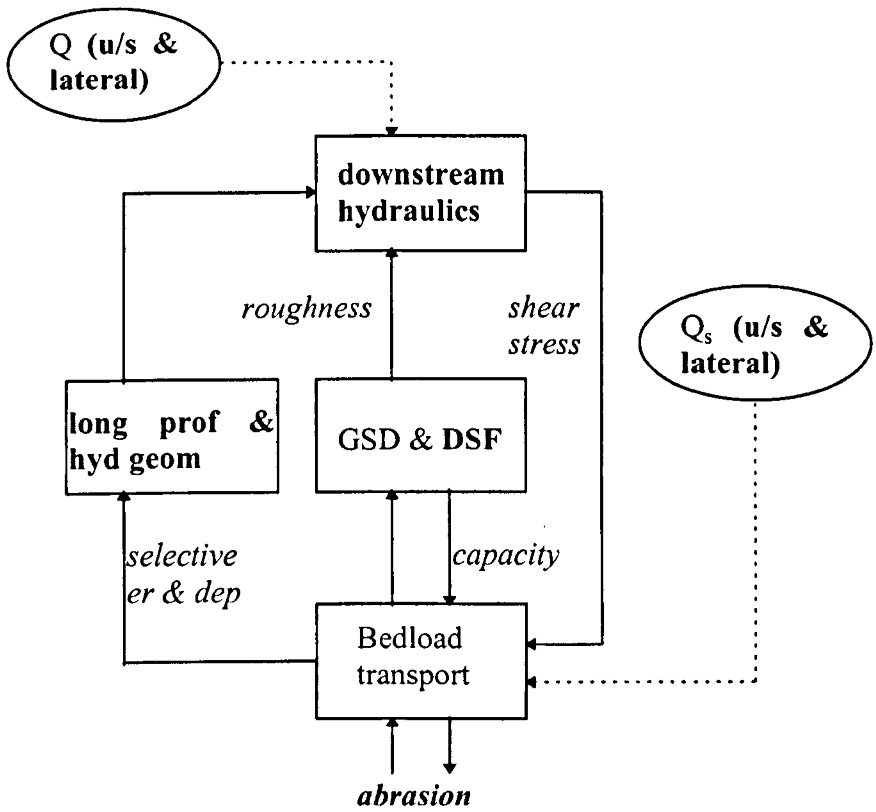


Figure 2.3: Forms and processes operating DOWNSTREAM in a gravel-bed river. The terms in **bold** indicate factors which become important when analysing forms and processes along a channel but do not require consideration when looking at a particular locality in a stream. Abbreviations as in Figure 2.1 apart from u/s is upstream, long prof is long profile, hyd geom is the hydraulic geometry and DSF is downstream fining.

Entering the Figure 2.3 from the top left, unsteady discharge **(Q)** flows down the river. The discharge varies depending on supply from upstream and tributary sources. If the slope decreases and discharge increases downstream the hydraulics acting on the channel bed will also change in a streamwise direction **(downstream hydraulics)**. This will cause variations in **shear stress** which will generally decrease downstream. Sediment supply **(Q_s)** from upstream has an impact on channel equilibrium, through the sediment continuity equation, and this is augmented by supply from lateral sources such as tributaries, mine waste or collapsed bank material. As the shear stress varies downstream so does the size and flux of material that the flow can move **(bedload**

transport). The supply of sediment on the channel bed and the shear stress along the river define the size of sediment entrained by the stream. This varies over space and time with finer sediment moved more often and further than coarser sediment (**selective erosion and deposition**). The slope of the channel **long profile** varies in a streamwise direction depending on discharge, sediment size and sediment sorting, among other factors. The **hydraulic geometry** of the channel also varies over space and time. The decrease in shear stress, combined with selective erosion and deposition often cause a gradual **downstream fining** of the bed **GSD** along the river. This tends to cause a decrease in **roughness** along the channel. As discharge increases downstream so does the **capacity** of the channel to carry load. As shear stress decreases, however, the flow is less likely to entrain and transport coarse grains as bedload. An additional factor to consider when analysing processes along a channel rather than locally is **abrasion** of sediment. This can occur as sediment is transported over a grain, or as the grain itself is in transport and therefore in contact with the bed and other bedload for a proportion of the time it is in motion. This process acts to reduce the size of the sediment.

The channel bed material, which influences flow resistance and sediment transport dynamics, varies both spatially and temporally. The general downstream reduction in bed slope, selective erosion, transport and deposition, and abrasion witnessed in gravel bed rivers can combine to cause an overall reduction in bed grain size along the channel. As noted in Chapter 1, this phenomenon is termed downstream fining. It should be noted that in many cases downstream fining can occur at a greater spatial rate than can be explained by abrasion alone (see, for example, Adams, 1979; Paola *et al*, 1992; Ferguson *et al*, 1996) and that the rate of downstream fining with distance increases in aggrading rivers (Shaw and Kellerhals, 1982). From this we can conclude that size selective sorting of bedload must be an important process in the downstream decrease in grain size exhibited by many gravel bed rivers. The basic assumption is that the coarser sediments are left in the upstream reaches of the river and the finer particles are preferentially winnowed out of the bed, and deposited further downstream (Ferguson *et al*, 1996).

A first attempt to quantify downstream fining was undertaken by Sternberg (1875) who proposed the empirically derived equation for bed grain diameter, D, shown below:

$$D = D_0 e^{-\beta x} \qquad \text{Equation 2.25}$$

where D_0 is mean particle diameter (m) at a reference section, β is the coefficient of particle size reduction and x is the distance downstream (m) from the reference section.

Further attempts have been made by various researchers to simulate downstream fining of river gravels numerically through the development of computer models. The key models (for example Parker, 1991a,b; Van Niekerk *et al*, 1992; Hoey and Ferguson, 1994; Cui *et al*, 1996) are based on similar assumptions and simplifications. In simple terms these state that: there is some relationship between the amount of sediment in transport and the specific hydraulics and sedimentology of the river concerned; initial conditions for the simulated reach are specified; a bedload transport function calculates fractional transport rates based on the specific hydraulic conditions; the degree to which the channel bed acts as a source or sink for particular grain sizes is evaluated; and the amount of aggradation or degradation along the entire reach is calculated. Further details of numerical downstream fining models can be found in Chapter 6.

2.2.1 Long profile and channel hydraulics

Downstream hydraulics and channel long profile

Rivers in temperate regions generally experience an increasing discharge downstream as tributary sources add to the flow in the main channel. This increase in discharge is usually accommodated by increasing the width to a greater extent than the depth, with little or no change in velocity (Leopold and Maddock, 1953; Hey and Thorne, 1986). The channel slope also tends to decrease downstream. These changes in slope and discharge have implications for shear stress, especially towards a base level. In this situation the channel slope decreases to zero. Shear stress, which is proportional to the depth-slope product, (see Equation 2.2 above), therefore also decreases. As the shear stress decreases the median bed grain size is reduced in a streamwise direction, as coarser sediment is deposited and finer sediment remains in transport, leading to the phenomenon of downstream fining. The bed roughness therefore also decreases in a downstream direction.

The change in form of a river channel in a streamwise direction is conditioned by a number of factors, for example discharge, sediment transport rate, sediment size, and sediment sorting, as well as geological structure and watershed evolution (Sinha and Parker, 1996). Shulits (1941) fitted an exponential decay relationship between slope (S) and distance downstream that approximates profiles to simple smooth mathematical functions:

$$S = S_0 e^{-\alpha x} \qquad \text{Equation 2.26}$$

where x is distance downstream (m) from a reference section whose slope is S_0, and α is the coefficient of slope reduction.

River long-profiles that are maintained over long timescales tend to be smooth and concave (Yatsu, 1955;1957) with the channel slope greatest towards the upstream end. Much past research has focused on fitting curves to various longitudinal profiles and as such provides only a limited explanation of the various processes involved (Sinha and Parker, 1996). The downstream decrease in slope can be attributed to the decrease in grain size of bed material through sorting and abrasion. Studies of downstream channel change can be grouped into three general categories (from Sinha and Parker, 1996):

1. The first approach isolates one particular variable for study, such as grain size variation downstream, or the effect of discharge (see Davis, 1899; Gilbert, 1914). In some cases, however, it is not possible to isolate one variable, for example where other variables are particularly complex or poorly understood. Hoey and Ferguson (1994) investigated a reach where discharge was constant downstream, but where grain sizes fine considerably, making their streamwise change the

dominant variable. Efforts have also been made to test Sternberg's (1875) law which states that the size of sediment on the bed of a river decreases exponentially with distance downstream (see, for example, Parker 1991a,b; Seal *et al*, 1997; Hoey and Bluck, 1999).

2. The second approach uses a process-response method, where both grain size and discharge have been considered together as the main controlling factors influencing variations in stream slope. Hack (1957) and Snow and Slingerland (1987) employed a statistical analysis of field data and a one-dimensional numerical model respectively to simulate long profiles for both equilibrium and non-equilibrium conditions. The studies determined how discharge, sediment transport rate and sediment size influenced the channel long profile. In the short term, the existing long profile controls shear stress and sediment transport, but if the transport capacity does not equal supply, then feedback occurs over time, via aggradation or degradation of the long profile, and the long profile will slowly evolve (as in the Hoey and Ferguson, 1994, model). This process would occur towards the left hand side of Figure 2.3.

3. A third group of researchers have concentrated on the likeness between physical and fluvial system modelling, rather than using grain size or discharge (as above) as the primary control on changing slope. Leopold and Langbein (1962) for example, in attempting to address their basic assumption that the hydraulic equations themselves were insufficient to determine river behaviour, used a random walk model to derive drainage control networks. Their findings showed that the drainage control networks produced exhibited some of the properties demonstrated by the streams studied by Horton (1945). Rinaldo (1999) employed fractal structures allied to digital mapping technology to reveal deep regularity in the forms of natural river networks.

Selective bedload sorting

The presence of a degree of size selective entrainment in many gravel bed streams, noted above, and changing channel characteristics downstream results in different sizes and volumes of sediment being entrained and transported. These changes define the size of sediment that is supplied to the downstream reaches of the channel. For selective sorting to be responsible for downstream fining the bedload transport rates must differ between size fractions. Selective entrainment implies that larger, more massive, particles remain less mobile than those which are smaller, and therefore require a higher shear stress to entrain them.

If an armour layer develops to the extent that all grain sizes become equally mobile then, apart from during exceptional floods, the only processes operating to cause downstream fining are the gradual wearing of grains, in transport and on the bed, and lateral input of fine sediment. It is probably true that in some situations these two factors can go a long way to explaining the downstream fining occurring in particular channels, but in relatively short rivers which have no appreciable lateral inputs of either water or sediment (such as Allt Dubhaig and Vedder River which are investigated for the current research), other processes must be occurring during entrainment and transport of bed material to generate the reduction in grain size.

2.2.2 Abrasion

As noted in Chapter 1, abrasion is a summary term covering mechanical actions such as grinding, breakage, impact and rubbing and these processes provide an alternative explanation to bedload sorting for the phenomenon of downstream fining. Various researchers (including Yatsu, 1955;1957; Adams, 1979; Shaw and Kellerhals, 1982; Kodama, 1992) have suggested that abrasion is the main cause of changing grain sizes along a channel. Sternberg (1875) and Davis (1902) originally suggested this hypothesis and it is supported by circumstantial evidence. Adams (1979) noted that angular pebbles are predominantly found near to their source and these become more rounded with distance travelled. Laboratory experiments have also shown that the abrasion of fluvial sediments occurs (Kodama, 1992;1994a,c). There is, however, a discrepancy between laboratory simulated abrasion rates and field downstream fining rates (Adams, 1980; Hoey and Ferguson, 1994). The laboratory tests indicate that a higher rate of abrasion occurs than that witnessed in the field. This discrepancy is based on the degree to which laboratory tests reflect the conditions in the field (Kodama, 1994c). Kodama (1992) also noted that particle lithology influenced both the size of the material supplied to the stream and the rate of reduction in size experienced by the particles. The differential rates of abrasion experienced by distinct lithologies provide further evidence that abrasion may play a role in the generation of downstream fining in some streams (Werritty, 1992).

As noted above, the term abrasion covers a number of distinct processes. Particles are chipped and may fracture both when in transport and in place when lift and drag can cause the particles to vibrate in their pockets in the bed (Schumm and Stevens, 1973). While in the bed the particles are also eroded by overpassing bedload (Brewer *et al*, 1992). The weathering of particles whilst in storage can increase their susceptibility to other abrasion processes. Jones and Humphrey (1997), using abrasion mill analysis in a laboratory, speculated that much of the abrasion experienced by a grain occurs soon after it is supplied to the stream. This initial high rate of abrasion was attributed to the presence of an easily eroded surface layer resulting from weathering. Once this weathered layer was removed the abrasion rate dropped off significantly. These results suggest that sediments present in a stream for a relatively long period of time will undergo only slow rates of abrasion. Also, if little new sediment is supplied to a stream the abrasion rate will decrease significantly with distance transported. These ideas supported those of Bradley (1970) who suggested that weathering of granitic grains could increase their abrasion rate by up to five times. Bradley also stated that biotite-bearing rocks (granite, gneiss and some aplite) were least durable and biotite-free rocks (pegmatite) were most durable. Quartz and chert were also highly resistant to abrasion.

It should be noted, however, that some researchers question the importance of abrasion in generating downstream fining (Brierley and Hickin, 1985). Bradley *et al* (1972) suggested that only 10% of the fining observed in the Knik River, Alaska, was caused by abrasion, with the rest generated by size selective sorting.

2.2.3 Lateral inputs of water and sediment

The grain size mix of sediments present on a river bed depends on the rate of sediment supply, of individual size fractions, to the river and the subsequent rate that the river transports these grains. If the magnitude of flow, rate of sediment supply or size of the particles is altered, the river channel will change its geometry and bed GSD towards a new configuration to allow the altered sediment load to be transported by the new flow capacity and competence (Mackin, 1948; Lane, 1955). A tributary joining the main stream in a given catchment will act to increase the discharge of the two streams downstream of their confluence. If bed slope remains constant this may lead to an increase in flow depth, and therefore shear stress, effectively increasing the maximum size and volume of sediment transported. This action will therefore act to reduce the rate of downstream fining with distance by transporting coarser sediment further along the channel.

In a similar way to additional sources of discharge, lateral sources of sediment can cause changes in the forms and processes developing and occurring along the channel by altering the GSD of the channel bed. This sediment may result from natural or anthropogenic activity, for example tributary channels (Campbell, 1970;1977; Bradley *et al* ,1972; Dawson, 1988; Pizzuto, 1992;1995; Rice and Church, 1998; Rice, 1999), mine waste (Knighton, 1989;1999) collapsed banks (Griffiths, 1979; Pizzuto, 1984) or valley sides (Schroeder, 1991).

Wolcott (1988) provided two contrasting data sets from lithologically distinct catchments which indicated that, rather than an in-channel process being responsible for bed GSD, the size of the material supplied to the channel may be the major control. This highlights the importance of these sources in defining the sedimentological characteristics of a stream. Where bimodal sediments were supplied to a channel the bed material remained bimodal. If the sediment input was unimodal the bed material also remained so. The streams analysed by Wolcott, however, were relatively short (2 and 10 km), and it was questionable whether any abrasion or sorting processes would have sufficient distance to modify the input material.

Brewer and Lewin (1993) investigated downstream trends and sediment characteristics in two rivers in Wales. Grain size generally decreased downstream although this trend was punctuated by 'jumps' below tributary inputs. These lateral sources of sediment also precluded the overall rounding of grains downstream by adding more angular material. Brewer and Lewin's field results indicated that hydraulic sorting rather than abrasion was the main process causing the downstream changes and these results were supported by laboratory evidence from an abrasion tank and tumbling barrel. The continued supply of angular material from tributaries and bank erosion complicated the downstream changes. These findings are supported further by Schroeder (1991), Rice and Church (1998) and Rice (1999) who argued that lateral inputs of coarse sediment into the main channel, from valley sides or lower order tributaries, disrupted the development of gradual downstream fining along a stream. Rice (1999) suggested that downstream fining was best developed between tributary inputs of coarse sediment which acted to reset the bed GSD by adding coarse sediments.

2.3 Form, flow and sediment transport in a gravel-sand bed river

This section discusses how the introduction of relatively large proportions of sand-sized sediment into a gravel-bed stream alters the forms and processes occurring locally and along a river. These processes require consideration for the mechanics of GST formation and evolution to be investigated. The following sections therefore concentrate on the new aspects of the fluvial system that were not included in either Sections 2.1 or 2.2 above. As these processes were less important when considering gravel-only rivers it can be inferred that these factors are necessary for a GST to evolve, instead of the gradual gravel downstream fining that occurs upstream. A flow diagram of the forms and processes important in gravel-sand bed rivers is shown in Figure 2.4 below.

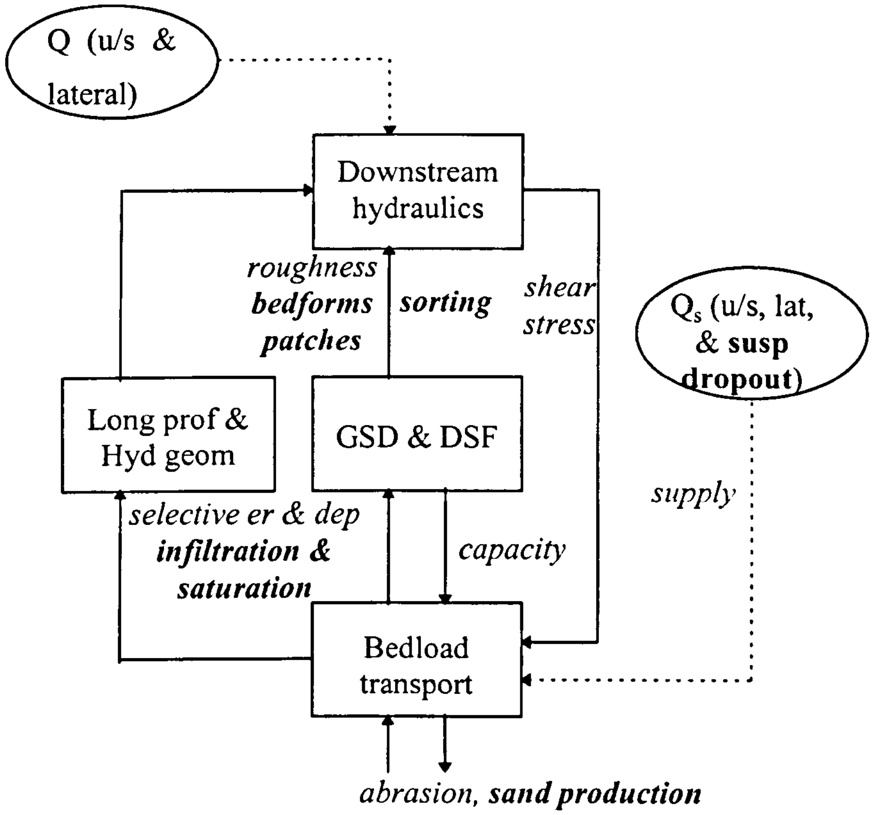


Figure 2.4: Forms and processes operating in a stream with a GRAVEL-SAND sediment present on the bed. The terms in **bold** indicate factors which increase in importance as the sand proportion increases downstream. Abbreviations are as in Figure 2.1 and 2.3 apart from: susp dropout is suspension dropout.

If Figure 2.4 is entered from the top left, unsteady discharge (**Q**) flows through the river, supplied from **upstream** and **lateral** sources. If the slope decreases and the discharge increases in a streamwise direction the **downstream hydraulics** acting on the channel bed will vary generally leading to a decrease in **shear stress**. Sediment is supplied (Q_S) from **lateral** sources and, in association with the reduction in bed slope, sediment that was carried in suspension in the water column may be dropped out as the shear stress falls **(suspension dropout)**. If suspended sediment is dropped then this material will either be deposited on the channel bed or may continue to be transported as **bedload**. As sand is deposited on the channel bed during periods of low to moderate flow the fines will **infiltrate** the coarser bed material, potentially leading to the **saturation** of the gravel-bed by sand. These fines therefore begin to dominate the bed sediment mix and are preferentially entrained over larger grains as discharge and shear stress increase during periods of high flow (**selective erosion and deposition**). The deposition of large volumes of fine sediment may reduce the **long profile** bed slope, encouraging the further deposition of sand and a break of slope to develop. As the bed slope and therefore shear stress change over a short distance a spatially accelerated rate of downstream fining may develop. The **abrasion** of gravels may augment **sand production** either through the gradual wearing of coarse grains or the spontaneous breakdown of particles caused by lithological weaknesses.

As sand infiltrates and begins to saturate the gravel bed, areas of this fine sediment will form on the bed surface. Due to the low roughness exhibited by these areas it may be expected that gravel will be transported rapidly across them as a result of the coarser sediments' exposure to the flow. A high degree of localised grain size sorting may therefore occur as the gravel and sand sizes organise themselves into discrete strips or **bedform patches**, each with different hydraulic characteristics. It can also be expected that, at low to moderate flows, sediment from the fine patches will be entrained preferentially to that from the gravel patches. These processes will lead to further **sorting** of both sand and gravel. If the sand patch was not present, however, this fine sediment would not be available for transport as it would be hidden beneath and between the coarser grains. Downstream of the zone exhibiting both gravel and sand patches on the bed surface, therefore, a sand-dominated bed may be expected. This spatially rapid change in bed **GSD and downstream fining** is a result of the increased mobility of the sand fraction with respect to the gravel and occurs because there is an increasing volume of sand present in the bed downstream.

As noted in Chapter 1, the distal reaches along a river often exhibit a spatially accelerated rate of downstream fining with a relatively sudden switch from a gravel-dominated to a sand-dominated bed. The fact that the switch tends to occur over an extremely short distance relative to the fining elsewhere along the river, and is often associated with an order of magnitude reduction in the bed slope, indicates that some threshold or non-linearity may be operating. This threshold may be related to a changing hydraulic regime upstream and downstream of the transition.

For a GST to occur there must be additional sources of sand to provide sediment for deposition onto a gravel-bed channel downstream. As noted above if the size of sediment supplied to a stream is altered, the river channel will change its geometry and bed GSD to accommodate this. The degree of influence that the fine sediment will exert over the hydraulic properties of the bed is poorly understood, as is the interaction between the fine and coarse fractions present in the bed GSD. Although a detailed investigation of these processes lies outside the scope of this thesis a review of the findings of previous studies may aid in the interpretation of some of the results of the current research regarding the form, characteristics and development of a GST. The possible sources of fine sediment required to allow a GST to form are detailed below and include: lateral inputs; abrasion to sand; dropout from suspension; and size selective bedload sorting. The latter two sources may be heightened in importance by the presence of the break of slope which reduces the shear stress over a short distance.

An important question to consider for the current research is why treat gravel-sand bedded streams separately to those that are gravel-bedded? There are several important aspects that should be noted to answer this question, some addressed by Simons and Simons (1987). Firstly, sand in the bed surface will extend the range of the GSD complicating further the size-selective entrainment of grains, introducing the possibility of sediment sorting through transporting grains in suspension rather than solely as bed load. Secondly, following on from the first aspect, the presence of sand, and consequent reduction in bed sediment sorting, may lead to complex process feedbacks through altering the near bed hydraulic and sedimentary characteristics. Finally, the bed GSD of a gravel-sand stream is often bimodal in characteristic, with peaks in the medium to coarse gravel and medium sand sizes but with a relative paucity in the coarse sand and fine gravel sizes. The presence of this grain-size gap between the two modes provides further support for the decision to treat sand and gravel streams separately.

Although not ubiquitous throughout gravel-sand rivers, the causes and impacts of the bimodal bed GSD often exhibited by gravel-sand streams have been investigated in some detail. To change a unimodal gravel into a gravel-sand mixture the proportion of sand must be increased through some mechanism. The bed GSD, however, will only remain bimodal if there is some reason for the existence of the grain-size gap. Three main mechanisms are thought to be responsible for this (from Sambrook-Smith, 1996):

1. The parent material may be of a particular type that, when weathered, produces only sand and coarse gravel-sized sediment (Milhous, 1982; Shaw and Kellerhals, 1982; Wolcott, 1988).
2. The preferential break down of fine gravel sizes where these sediments are readily crushed into their constituent sand-sized grains (Yatsu, 1957; Shaw and Kellerhals, 1982; Kodama, 1994a) leaving only sand and coarse gravel sizes on the bed surface.
3. Preferential entrainment of fine gravel leaving only coarse gravel and sands present (Russell, 1968).

It is likely that, in many cases, a combination of the above processes is responsible for the continued presence of a grain-size gap (Sambrook Smith, 1996). Shea (1974) suggested that bimodality and the associated grain-size gap did not occur, arguing instead that the few data sets that were then available had either been misinterpreted or analysed incorrectly. Reid *et al* (1997) agreed that a careful sampling strategy was required if bed size distributions were to be adequately characterised and individual depositional units needed sampling. Failure to do so may result in unrepresentative bimodal GSDs being assigned to particular areas of a river bed. If GSTs occur in streams which do not have strongly bimodal bed GSDs, however, other processes must be responsible for the creation of spatially rapid downstream fining, rather than simply the existence of a grain size gap, and these processes also require closer examination.

The remainder of this chapter, therefore, aims to outline the main sources for the increased proportion of sand that may cause bimodality and initiate a GST, investigate how this impacts on the sediment transport processes occurring in a stream, and to show how the fine sediments alter the characteristics of the stream bed itself.

2.3.1 Sources of sand

Lateral inputs of sediment

Large volumes of fine sediment can be provided by lateral inputs and in some situations this can occur with little or no associated change in bed slope. An example of a stream that receives fine sediment in this way is the Red Deer River, Alberta, Canada (Campbell 1970;1977). The GST in this stream occurs where tributaries of the river, which drain badlands, enter the main stream. Large volumes of fine sediment are supplied from easily erodible banks. The excess sand input clogs and buries the gravel bed and these fines are only removed during rare, high-magnitude flood events (Campbell, 1977). A break of slope in the channel long profile is not associated with the switch from a gravel to a sand bed as the extra quantity of fine load is traded off against the lower overall bed grain size (Sambrook Smith, 1994).

The input of fine sediment from a lateral source can also be the result of human impact in a catchment. Knighton (1989;1991;1999) found that mine waste supplied to the Ringarooma basin, Tasmania, caused a change in the dominant sedimentological characteristics of this stream. He noted, however, that the very large inputs of sand required to cause these changes mean that they only occur in specialised conditions where large volumes of fine sediment are available for entrainment by a river (Knighton, 1999).

A bimodal bed GSD can also be caused by inputs of sediment with a range of sizes that has a bimodal GSD, caused by in situ breakdown through weathering at source. Wolcott (1988) suggested that the size distribution of the parent input material may be largely responsible for the existence of a grain-size gap in some streams.

Rice (1999, pers comm) argued that lateral sources, such as tributaries, were unlikely to act as a source of fine sediment, instead tending to provide particles coarser than those present in the mainstem of the channel. Due to the relative size of a tributary, compared to the main channel, any fine sediment that is supplied is transported rapidly downstream due to the increased competence and capacity of the larger channel. A GST is therefore unlikely to be found immediately downstream of a tributary and these lateral channels are more likely to be associated with a coarsening of the main channel bed (Rice and Church, 1998; Rice, 1999).

Abrasion to sand

Yatsu (1955;1957) and Kodama (1992;1994a) suggested that coarse gravel lithologies, in high energy conditions, can break down, through weathering and abrasion, into medium gravel and sand sized particles. This process will give rise to a bimodal bed GSD and consequently will increase the proportion of fine sediment present in the channel bed. Yatsu and Kodama argued that in some rivers this process alone could explain the whole of the downstream fining phenomenon. It seems likely, however, due to the high compressive stresses required to break down river gravels, that abrasion is only an important factor in relatively large and active channels (Sambrook Smith and Ferguson, 1995). Yatsu (1955) stated that a bimodal bed GSD could also be formed by crystalline granitic rocks which have a tendency to be broken down, when they are abraded to approximately 10 mm in diameter, to give sand-sized sediment. Yatsu suggested that the material of sizes between the modes may be structurally unstable accounting for its dearth in the bed GSD.

Abrasion of some lithologies is assumed to give silt-sized products which are removed as washload. Ikeda (1970) noted that limestone, mudstone, slate and granite gravels may be more susceptible to weathering than other rock types, although not all of these lithologies would abrade or break down to produce sand-sized sediments. High-energy abrasion experiments carried out by Kodama (1992), however, suggested that andesite would readily break down into sand through granular disintegration. A reduction in the amount of andesite gravel and an increase in the proportion of sand downstream led Kodama to suggest that this process was the likely cause of a reduction in bed grain size in the Watarase River, Japan. Kodama also found that, in mixed-size sediments, finer gravels may be preferentially crushed by coarser grains creating a bimodal distribution in the bed GSD with peaks in the medium to coarse gravels and sand sizes.

Yatsu (1955;1957) indicated that the spatially rapid size reduction caused by the breakdown of crystalline rocks was associated with a decrease in bed slope, although no evidence was actually supplied in his work to support this hypothesis. A similar explanation for the rapid bed grain size reduction and break of slope witnessed in the Siret River, Romania was suggested by Ichim and Radoane (1990). Here the dominant sandstone and quartz bed sediments were thought to generate large amounts of sand through abrasion.

Shaw and Kellerhals (1982) also agreed that abrasion of fine gravels (less than 10 mm diameter) was an important process generating changes in sedimentological characteristics along a stream. This abrasion created finer sands that could be carried in suspension until, with decreasing shear stress downstream, these were deposited on the channel bed. This deposition was though to cause a bimodal bed GSD. The fine sand mode of this distribution increased in magnitude downstream with respect to the gravel mode as the gravel was continually abraded.

The process of abrasion can also cause disparities in the size of bedload and bed sediment, even if these are sampled at the same site. In situ weathering can weaken the particles on the bed surface (see Bradley, 1970; Parker, 1991a), and these will often break down soon after entrainment (Kodama, 1992).

Break of slope

The GST is often, although not always, associated with an order of magnitude reduction in bed slope over a short distance (Kodama, 1994b; Sambrook Smith and Ferguson, 1995; Talling, 2000). The consequent decrease in stream energy causes material in transport to be deposited and less entrainment of surface sediments occurs (Yatsu, 1957; Kodama, 1994b; Sambrook Smith and Ferguson, 1995). It is unclear, however, if this sediment deposition causes, or is a result of, the decrease in channel slope and therefore shear stress. A spatially rapid reduction in bed slope will reduce bed shear stress leading to both a dropout of sediment from suspension and a decrease in the amount and size of particles that can be transported as bedload. It should be noted, however, that in some streams a break of slope is not concurrent with a rapid reduction in bed grain-size (Campbell, 1970;1977; Shaw and Kellerhals, 1982; and the Vedder Canal investigated for the current research).

Local base level control, such as a lake, fan, or main channel, can also cause a relative increase in sand supply to the bed material as the amount of gravel transport decreases. Pickup (1984), in describing the Fly River, Papua New Guinea, suggested that a rise in sea-level caused a reduction in water surface slope and therefore decreasing shear stress through a backwater effect. Gravel deposits are therefore left as a lag, and greater deposition of sand takes place. With low slopes downstream, and sand as the dominant sediment size being supplied, the transition zone rapidly becomes fully sand-bedded, causing vertical grain-size changes in the alluvial basin. These findings mirror those of Mulder and Syvitski (1996) who argued that a fall in sea level would result in a greater supply of fluvially-transported sediment to the continental shelf.

Dropout from suspension

Another possible source of the sand necessary to create a GST is from suspended sediment dropped on to the stream bed (Paola *et al*, 1992). Dade and Friend (1998) suggested that a change in transport mode of sand from suspension to bedload occurred in many rivers as shear stress fell downstream. This change in transport mechanism from suspension to bedload may be responsible for rapid reduction in bed slope over a short distance related to a spatially sudden sediment supply as the shear stress falls below that required to suspend the modal sand size. This source of sediment may have an impact on the bed surface sedimentology in the stream.

In rivers where rapid downstream fining occurs between a gravel and sand bed without a measurable change in bed slope, however, (such as those studied by Shaw and Kellerhals, 1982; and the Vedder Canal investigated for the current research) then this process cannot be held responsible as the source of sand to cause the downstream variations in sedimentology. It is also important to note that, although dropout from suspension can provide sand to the bed of a stream, other processes are required to cause this sand to generate a GST. The deposition of sand may generate a break of slope, leading to further deposition of sand, or cause changes in the bed surface sedimentology, altering the characteristics of the bed. These additional processes are equally important in generating a GST as the initial supply of sand from suspension dropout.

Selective bedload sorting

As noted in Section 2.2.1 selective sorting is the process which moves mixed-size sediments at different rates, possibly by distinct processes. Russell (1968) claimed fine to medium sands would be transported through saltation with medium to coarse gravels moving through sliding or rolling along the bed. The intervening coarse sands and fine gravels were thought to be transported by both mechanisms. These combined processes would act to cause a preferential entrainment of the grain-size gap fractions. These sediments would be transported through the reach more rapidly, and due to their increased mobility, would undergo a higher degree of abrasion in comparison to the coarser and finer grains. As noted above, however, more recent studies, beginning with the investigations of Paintal (1971), have questioned the reliability of Russell's hypothesis in explaining sediment transport in natural streams.

Yatsu (1955;1957), when studying aggrading fan systems, found that the sudden transition to a sand bed was always associated with a break of slope. Shaw and Kellerhals (1982), investigating streams in Alberta, Canada, whose slopes were essentially predetermined by the dip of the North American Great Plains, found no spatially rapid reduction in bed slope in the region of the switch from a gravel to a sand bed. The abruptness of some GSTs in rivers which have no pronounced break of slope, and no obvious lateral input of fines indicates that some threshold or non-linearity must be present in the entrainment, transport and depositional processes of bedload in these streams. The lack of a break of slope, and consequent reduction in shear stress, removes dropout from suspension as the potential source of the sand required to cause the spatially rapid downstream fining associated with a GST.

If selective transport of finer grains occurs then one can conclude that finer particles in the bed will move more frequently and further than coarser particles. As noted above, however, many researchers (such as Parker *et al*, 1982; Wilcock, 1998) disagree with this hypothesis, stating that a small proportion of sand present in bed dominated by a gravel framework will be unavailable for entrainment until the gravel has been moved. This process would tend to move the system away from selective transport of finer grains, towards equal mobility at entrainment of all sediment sizes present at the bed surface.

Lisle (1995) proposed an additional process that could aid size selective sorting through differential fractional transport rates. He suggested that when some of the fine sediment present in the bed was entrained it was transported rapidly through the reach without deposition and storage in the bed. This process was thought to be less influential than, and subsidiary to, the selective entrainment process itself.

2.3.2 Sand transport

Bedload transport

At low shear stress bedload is size selective and once some sand has been added to the bed (whether through abrasion, suspension dropout or due to a gradual reduction in shear stress downstream and therefore deposition from bedload) its availability for transport at medium and high flows increases This causes a rapid change in the relative amounts of sand and gravel being entrained. Many sediment transport equations use a single average grain size to characterise the bed GSD (for example Meyer-Peter and Muller, 1948). In natural channels, however, particularly those with both sand and gravel present on the bed surface, this technique is likely to offer a poor estimate of the transport characteristics of individual grain sizes which move at different rates. An overall transport rate predicted by the median grain size will under-predict the transport rate of sand and may be several orders of magnitude larger than the transport rate for gravel fractions (Leopold, 1992). As noted above, when investigating the processes occurring in mixed size sediments, such as those present in gravel-sand bed streams, the transport rates of individual size fractions require calculation (Parker *et al*, 1982). Using this technique the rates of movement experienced by different size fractions can be investigated.

Wilcock (1998) argued that if only small amounts of sand were present in the bed, or being deposited from transport, the grains would settle in the interstices between the gravel particles. Consequently, there will be little or no sand directly exposed to the flow. In order to entrain the sand present on or near the bed surface the river must first remove the coarser gravels within which the sand is stored. In this case the critical shear stress for the incipient motion of sand can be thought of as essentially the same as that for gravel. If this hypothesis were true for the length of a channel's long profile, however, the bed GSD would remain constant along the river as selective sorting by grain size would not occur. In this case the only processes that could be responsible for downstream fining would be lateral inputs of fine sediment or abrasion. Downstream fining has been noted in streams which have little or no lateral inputs and can occur over a distance short enough to make size reduction through abrasion negligible. For a GST to occur, therefore, there must be an important process, or processes, missing from the equal mobility hypothesis as the proportion of sand in the bed surface increases.

The transport rates of gravel and sand were thought, by Wilcock (1998), to depend on the proportion of sand present on the bed surface as it is these sediments upon which the transport rate immediately depends. The amount of fine sediment can define the respective amounts of sand and gravel available for transport by the flow and this also influences the mobility of each fraction (Ikeda and Iseya, 1988; Kuhnle, 1993a,b; Wilcock and McArdell, 1993; Wilcock, 1997; Wilcock and McArdell, 1997; Montgomery *et al*, 1999). The fractional sediment transport rate (q_i) is proportional to both the availability of a particular size (F_i) and the excess shear stress (see Equation 2.21). Wilcock (1998) noted that the relative transportability of sand and gravel modes could be investigated by considering the ratio of the critical shear stress for entrainment of sand (τ_{cs}) over the critical shear stress for entrainment of gravel (τ_{cg}). The ratio, τ_{cs}/τ_{cg}, was calculated from values of τ^*_{ci} (the dimensionless critical shear stress required to entrain the ith size fraction) for the maximum and minimum limits of sand content. These values are shown in Table 2.1 below.

Table 2.1: Approximate values of dimensionless critical shear stress at the limits of bed surface sand content (from Wilcock, 1998).

	Clean gravel (sand (F_S) = 0%)	Clean sand (sand (F_S) = 100%)
τ^*_{cs}	0.04	0.01
τ^*_{cg}	$0.04(D_g/D_s)$	0.04
τ_{cs}/τ_{cg}	1	$4(D_s/D_g)$

When the bed contains no sand τ_{cs}/τ_{cg} is approximately equal to 1. When the bed contains 100% sand τ_{cs}/τ_{cg} is approximately equal to $4(D_s/D_g)$. In many rivers, since D_g is one or more orders of magnitude larger than D_s, the decrease of τ_{cs} with a reduction in the proportion of sand in the bed is larger than the reduction of τ_{cg}. This fact suggests that sand becomes relatively more mobile as the proportion of fines in the bed increases. The nature of the reduction in the critical shear stresses for the entrainment of both sand and gravel with increasing proportions of sand, however, requires further investigation.

Wilcock (1998) investigated the impact that the proportion of sand in the bed had on both τ^*_{cg} and τ^*_{cs} by analysing sediment transport data from four rivers and one flume experiment. The bulk near surface bed sand fraction in these streams varied between 15 and 59%. Wilcock found that the shear stress values required to entrain a small amount of gravel or sand decreased rapidly between a bulk bed sand fraction of 15 to 22% and 34%. This finding suggests that a shift from a gravel-framework to a sand-matrix bed, required to reduce the critical shear stress for entrainment, occurs over a small increase in the bulk near surface bed sand content.

Suspended sediment transport

The downstream decrease in shear stress, and hence settling velocity, which occurs in all rivers with a concave long profile, could lead to a major increase in sand bedload as the shear stress falls below the settling velocity of the modal sand size carried in suspension. This would allow sand to persist on the bed unprotected by coarser grains (Shaw and Kellerhals, 1982; Paola *et al*, 1992). Surface sand content is likely to be highest where the suspended sand is coarse, so that viscosity is of little importance to its deposition. In some streams there may therefore be an overlap in the GSDs of bedload and suspended load depending on flow characteristics and sediment supply (Andrews, 2000).

The upward component of turbulence provides the lift necessary to entrain a grain into the flow and supports a grain during transport. Once in suspension the force exerted by gravity on a particle is a fundamental control in defining whether that particle is deposited. The interaction between this force and the upward component of turbulence often results in a vertical distribution of suspended sediment in which concentration and average grain size decrease with distance above the bed (Lapointe, 1992). As a particle falls through the fluid, the forces opposing its motion increase, until it reaches its terminal, or fall velocity. This fall velocity increases with grain diameter. For a grain to remain in suspension the fall velocity must be exceeded sufficiently often by vertical pulses associated with flow turbulence which are proportional to the shear stress or shear velocity. For any given flow condition there is therefore a maximum grain size that can be transported in suspension. Once the shear velocity falls, below the settling velocity of a particle, deposition occurs. As the settling velocity for a particle is closely related to its size the coarsest grains are deposited first, followed by finer grains as the shear velocity continues to fall.

A reduction in the variation of fall velocity for different sizes of finer grains was proposed by Peloutier *et al* (1997) to be caused by the turbulent entrapment of the finest grains combined with the low probability of infiltration of saltating particles. It was noted, however, that the median size of the fine material deposited decreased downstream due to selective deposition. In the field dropout from suspension will be influenced by both shear stress and bed GSD. In these natural conditions gravel surface layers have more overall influence on the settling velocities of coarser particles than finer ones. The rate of infiltration of the deposited suspended sediment into a gravel bed, however, was not found to be strongly related to the size of the bed sediment (Peloutier, *et al*, 1997). Carling (1984) found that the infiltration rate of fine sediment into a gravel bed was strongly related to the suspended sediment concentration in the flow, and for very high concentrations sand was deposited on the bed from suspension until it formed ripples on the bed surface.

Sambrook Smith (1994) presented an analysis of the estimated maximum size of material that could be carried in suspension upstream and downstream of the GST in a number of streams and laboratory flume runs. In most cases the size of suspended sediment decreased, although in many streams, after the transition, grains of up to 0.5 mm could be carried in suspension. Sediment of this size is coarser than the fine mode present in the bed beyond the GST in the majority of streams studied to date. This analysis suggests, therefore, that while important, dropout from suspension should not be thought of as the sole cause of GSTs.

2.3.3 Sand deposition

Infiltration of fines into a gravel bed

Diplas and Parker (1985) suggested that sand moving as bedload will infiltrate into the subsurface to a depth of between 2.4 and 4.1 D_{90}. The remaining pore space in the gravel bed above this depth is then filled until this is saturated with sand. Once this occurs any additional sand deposited will remain on the surface, available for entrainment at shear stresses considerably lower than if all the sand available was present only beneath the gravel bed surface. Diplas and Parker also found that the Shields stress affected the depth of infiltration with lower infiltration depths at low stresses. This process may act as a mechanism to decrease the capacity of a gravel bed to hold interstitial sand with distance downstream as slope, and therefore shear stress, decrease.

Einstein (1968) stated that fine sands (of about 0.2 mm in diameter) will infiltrate to the bottom of a gravel bed, and this bed will then fill with fines from the bottom up, with none appearing on the surface (and hence available for size-selective transport) if there is space in the pores below. Experiments carried out by Frostick *et al* (1984) using sediment traps support these findings. Wathen *et al* (1995), however, indicated that sand transport can occur in areas of low bulk sand content at shear stresses below that necessary to entrain the armour layer. Beschta and Jackson (1979) and Allan and Frostick (1999), however, argue that if the sand fraction is coarser (approximately 0.5 mm in diameter, depending on the pore size) these particles may be able to bridge pore spaces near the surface, forming a seal, and hence inhibit deeper infiltration of fines.

Saturation of a gravel bed by sand

Carling and Reader (1982) stated that when the bulk sand proportion in a gravel-sand bed stream is lower than 20% the river bed would comprise an interlocked gravel-framework. As the proportion increases above 20%, however, the grains within the framework begin to lose contact with each other and as the proportion exceeds 40% the gravel framework is replaced by a sand matrix containing discrete gravel clasts. Although more research into these processes is required it is generally recognised (Carling and Reader, 1982; Carling, 1984; Peloutier *et al*, 1997; Wilcock, 1998) that as the sand approaches 20% of the bulk bed volume the gravel grains forming the bed framework begin to lose contact and as the sand approaches 30 to 35% the bed becomes a matrix supported structure with sands being the dominant grain size found at the surface. In a matrix supported bed such as this sand is present on the bed surface and the transport rates of this fine fraction should approach those of a purely sand bed. For a gravel clast to be transported the sand surface needs to be removed by scour, and the entrainment of this coarser grain is no longer dependent on the adjacent gravel clast, as would be the case in a gravel-framework bed. There is therefore likely to be a large variation in both τ^*_{cg} and τ^*_{cs} as the proportion of sand in a gravel bed rises from less than 20% to more than 40% (Ikeda and Iseya, 1987;1988).

Sambrook Smith *et al* (1997) stated that the hydraulic properties of a sand-gravel bed may be more closely related to the grain size characteristics of the bed when this is defined using the area of the bed covered by sand or gravel, rather than bulk volume measurements of the bed GSD.

Grain size patches in gravel-sand bed streams

Sand present on the surface of a gravel bed often congregates in patches (Ashworth and Ferguson, 1989; Sambrook Smith, 1994; Paola and Seal, 1995; Lisle and Hilton, 1999) making it available for entrainment even at low width-averaged concentrations (Wathen *et al*, 1995). In the area of the fine patch sand will be available for transport at a shear stress lower than it would have been if it was hidden in the interstices between gravel grains, leading to a decrease of τ_{cs} with respect to τ_{cg} (Jackson and Beschta, 1982; Carling, 1983; Diplas and Parker, 1985; Iseya and Ikeda, 1987; Whiting *et al*, 1988; Ferguson *et al*, 1989; Diplas and Parker, 1992; Lisle, 1995; Paola and Seal, 1995; Wilcock, 1998). This selective transport will lead to the development of larger, more numerous, well sorted fine patches downstream and therefore provide sand for the formation of a GST. Entrainment from each grain size patch may approach equal mobility but on a width-averaged basis selective transport will occur (Church *et al*; 1991; Wathen *et al*, 1995; Paola and Seal, 1995). These fine patches therefore contain the first bed material to be entrained during rising stages, the most erodible at bankfull stage and the last to be deposited during waning stages (Andrews, 1979; Lisle, 1979; Meade, 1985; Komar, 1987; Sear, 1996; Wilcock *et al*, 1996). Wilcock (1993) suggested that in strongly bimodal sediments hiding and protrusion effects would be reduced, and the critical shear stress for a particular grain size may approach that of its Shields equivalent through size segregation on the bed surface.

Downstream hydraulics

When a fine or medium gravel, such as that found immediately upstream of many GSTs, becomes partially covered by sand, possibly deposited from upstream or lateral sediment sources, the hydraulic roughness of the bed decreases rapidly. As the saturation threshold is approached a small increase in the amount of sand present on the bed surface causes the hydraulics to switch to those associated with a purely sand bed channel. Decreases in bed roughness will cause shallower flow and therefore a decrease in shear stress (Sambrook Smith and Ferguson, 1996). Another consequence of this is that a bed which is hydraulically smoother will have a lower pivot angle for gravel entrainment and this may affect sediment transport by creating a sand bed with gravel overpassing (Wilcock, 1993; Sambrook Smith *et al*, 1997). These processes will lead to an increase in the amount and frequency of sediment being transported through the reach.

2.4 Summary

From the information presented in this chapter it can be seen that the dominant sediment type found on the bed of alluvial channels changes in a downstream direction. The sedimentology can switch abruptly from unimodal gravel, through mixed sands and gravels, to dominantly sand over a short distance. Over the course of this switch additional sorting forms and processes occur as sediments congregate in patches or stripes of ambient grain size. A stream bed can only maintain a gravel surface if the rate of sand deposition does not exceed the pore space available in the gravel framework. Once sand begins to accumulate on the surface, positive feedback occurs through the increased mobility of the fine grains. Patches of fine sediment are entrained at shear stresses below that needed to move sediments from the coarser patches. These fine patches can form even when the width-averaged sand content is relatively low. As finer, sand-sized sediments are increasingly mobile these grains clog up any gravels remaining at the bed surface, leading to a sand-dominated bed in only a short distance. A change therefore occurs, from a gravel-framework deposit with sand infill to a sand-matrix supporting gravel clasts, which decrease in number downstream as shear stress falls.

The influence of the magnitude of the bed sand fraction on critical shear stresses for the movement of both gravel and sand produces a mechanism for the formation of a GST. The transition is often located where the transport capacity of the river is reduced relative to the sediment load. This can result from a reduction in bed slope, a backwater or a lateral input of fine grained sediment. A break of slope, backwater or lateral input in the region of a GST is not ubiquitous, however, and a GST can form without their presence. As the GST often occurs over a distance shorter than that expected by the gradual change in bed hydraulics or through abrasion, its abruptness suggests that a discontinuity exists in sediment transport processes as the bed sand content increases. Sambrook Smith (1994) and Parker (1998) suggest that even with a smooth increase in the transport rate, with reducing grain size, a GST can occur as a result of a gap in the bed GSD in the fine gravel sizes. An alternative explanation is that a small increase in the proportion of sand contained within a gravel framework leads to a large increase in the relative mobility of both gravel and sand. The decrease in shear stress required to entrain the sand fraction is of a larger magnitude, however, and the fines can therefore be transported at lower discharges than the coarser gravels. The enhanced mobility of the sand accelerates the rate of hydraulic sorting in the channel reach exhibiting both sands and gravels on the bed surface (the united gravel-sand zone) to the extent that sand, and not gravel, is preferentially transported downstream of this region, creating a GST.

For this process to occur there must be a rapid change in the proportion of sand at the bed surface with only a small increase in the bulk sand content. This would act to increase the availability of sand and also alter the near-bed hydraulics of the channel which influence the mobility of sands and gravels. From the evidence presented above it appears that a threshold exists in the percentage of sand present in the bed, which governs whether the stream behaves as it is sand or gravel bedded. Research indicates that this value is likely to fall between 20 and 40% of bulk bed sand content, although sand can be present on the surface at bulk contents as low as 10%.

2.5 Research questions

The primary aim of the current research is to elucidate the forms and processes occurring along a gravel bed river as the proportion of sand increases. This over-riding aim is tackled by characterising the sedimentology of a GST, assessing how channels evolve in the region of the GST, and investigating whether these transitions can be modelled using width-averaged bedload sorting alone. This thesis will attempt to answer the following specific questions within the three objectives:

Characterisation of contemporary GST sedimentology:

- How does bed grain size change in a streamwise direction through a GST?
- To what extent does lateral sorting of grain sizes occur in the united gravel-sand reach?
- Can a quantitative assessment of the proportion of sand in the bed be obtained by a qualitative observation of bed surface sedimentology?
- How can gravel-sand sediments be sampled to obtain representative bed GSDs for a particular distance downstream?
- What proportion of bulk sand is required to cause a switch in bed surface sedimentology from gravel-framework to sand-matrix?
- What is the relationship between bulk and areal sand content in gravel-sand bed streams?

Channel change and mechanisms of GST formation:

- To what extent does size selective transport occur in distal fine gravels immediately upstream of the GST?
- Is there evidence of processes causing sorting of fine gravel into patches in the united gravel-sand reach?
- How does the surface morphology and sedimentology of the gravel front and united gravel-sand reach evolve over time periods of 10^0 to 10^2 years?
- Are lateral inputs of fine sediment crucial for the formation of a GST?
- Do GSTs form without a sharp reduction in bed slope over a short distance?

Modelling the GST through width-averaged bedload sorting:

- How does a one-dimensional model of size selective gravel bedload sorting need to be modified to be capable of simulating gravel-sand mixtures?

- Can a GST be formed through numerical modelling of width-averaged size selective bedload sorting alone?
- Can a GST be formed through numerical modelling of width-averaged size selective bedload sorting in the presence of a spatially rapid break of slope?
- Can a GST be formed through numerical modelling of width-averaged size selective bedload sorting and the overwhelming of a gravel bed by sand?
- How can a one-dimensional model of size selective gravel bedload sorting be improved to simulate accurately the behaviour of gravel-sand mixtures?

2.5 Thesis outline

In Section 1.6 of the previous chapter a general thesis structure was presented. The present chapter detailed the research relevant to downstream fining and GSTs that has been carried out to date and posed a number of specific questions that require investigation. The results of investigations carried out for the current research to answer these questions will be presented in the forthcoming chapters.

Chapter 3 outlines previous research that has been carried out on the two fieldsites chosen for detailed study as part of the current research. Details of the bed surface sedimentology of contemporary GSTs is presented in Chapter 4. Here the importance of both streamwise and lateral sorting is assessed. The most effective methods to characterise and sample these sediments are investigated, as is the relationship between bulk and areal sand content. The morphology of the channel is also quantified to check changes in cross section and long profile form. Chapter 5 concentrates on processes occurring in GSTs and change in the GST reach of channels over time. A tracer study of sorting processes in the distal gravel and united gravel-sand reach is employed and medium to long term change is analysed through subsurface probing in the GST zone. Cross-sectional change over time is assessed to investigate the importance of lateral inputs of fine sediment.

Chapter 6 examines the characteristics of physically-based numerical models that have the capacity to simulate downstream fining and GSTs to assess which is most suitable for application to the current research. In Chapter 7 the difficulties associated with modifying the chosen gravel-only model, to allow it to simulate gravel-sand sediment mixtures, are reported. A sensitivity analysis of the updated model is also included. In Chapter 8 a series of runs using the updated model to simulate gravel-sand mixtures are reported. These include attempts to use data collected in the field as initial conditions, and runs analysing the impact that a break of slope and the overwhelming of a gravel bed by sand has on the bed GSD.

Chapter 9 draws together both the field and modelling aspects of the research to assess which forms and processes may be most important in generating a GST. Internal model validation is discussed, along with how the field data can be used to suggest possible improvements to the numerical model employed for the research. Further analysis of the field data is also undertaken, a number of implications of the current research are suggested and potential further work that could assist in future study of gravel-sand bed rivers is outlined. Chapter 10 answers the research questions posed above using evidence presented in the thesis.

Chapter 3. Field context: previous work on Allt Dubhaig and Vedder River

This chapter will form the foundation for Chapters 4 and 5 which are concerned with the characteristics, morphology and evolution of a GST. It features a review of past work on the field sites studied in detail for the current research, Allt Dubhaig and Vedder River. The aim is to provide a context into which the research findings can fit as it is useful to understand the general morphology of the rivers which are being investigated. Both Allt Dubhaig and Vedder River have undergone extensive field investigation by numerous researchers in the past and much of this study is pertinent to the analysis presented elsewhere in this thesis.

3.1 Choice of fieldsites

The main aim of the fieldwork programme is to elucidate the forms present and processes occurring in rivers which have gravel-sand sediment mixtures present on their bed surface and exhibit a GST. This is achieved through detailed characterisation of GSTs, analyses of their evolution and an attempt to numerically simulate a GST through width-averaged size selective bedload sorting alone. If model simulations fail to represent the field conditions then detailed knowledge of the actual forms and processes occurring may assist explanation of why this failure occurred. The relative importance of the processes, not included in the model, for the formation of a GST may then be gauged.

As noted in Chapters 1 and 2 it is generally assumed that, in rivers with no lateral sediment inputs, downstream fining occurs through some mix of abrasion and selective sorting. The relative importance of each process, however, has been the source of debate. As the purpose of this study was to investigate whether downstream fining between a gravel and a sand bed can occur through sorting alone the chosen fieldsites had to satisfy the following criteria: exhibit spatially rapid GSTs (relative to their size); have predominantly resistant bed sediments so that the importance of abrasion can be considered negligible; and as far as possible be free from lateral inputs of water or sediment in the transitional reach since these can unnecessarily complicate the downstream fining processes which is being investigated (Knighton, 1980; Rice, 1998).

The particular fieldsites studied for the current research were also chosen in an attempt to show that GSTs are not simply a manifestation of the conditions present in a specific river, rather that they are a phenomenon that occur at both field sites which exhibit different morphological conditions. In the two rivers chosen for further study the GST reaches were split up into a number of cross sections in an attempt to characterise the reaches as a whole. Details of the sites and the frameworks set up for investigation of the channel can be found in the following sections. Further discussion of the data collection framework can also be found in Chapter 4.

3.1.1 Allt Dubhaig

Allt Dubhaig is a relatively small headwater tributary of the Tay, which drains about 17 km^2 in the central Scottish Highlands (Figures 3.1 and 3.2). It has an alluvial channel which comprises gravel and sand sized sediments and is approximately 3.5 km in length with an average width of 10 m. The river cuts through hummocky moraine at its upstream end which supplies a wide range of sediment sizes. These moraines have been interpreted as relics of outlet glaciers fed by the plateau ice caps from the west and the east during the Loch Lomond Stadial (Sissons, 1974). Towards its upstream end the channel crosses several bedrock sills which impede degradation and are exposed. This implies negligible aggradation at these sites although it is likely that this occurs further downstream. Almost all of the water and sediment flowing through the channel enters from upstream since there are no major tributaries and little lateral migration, particularly in the GST reach.

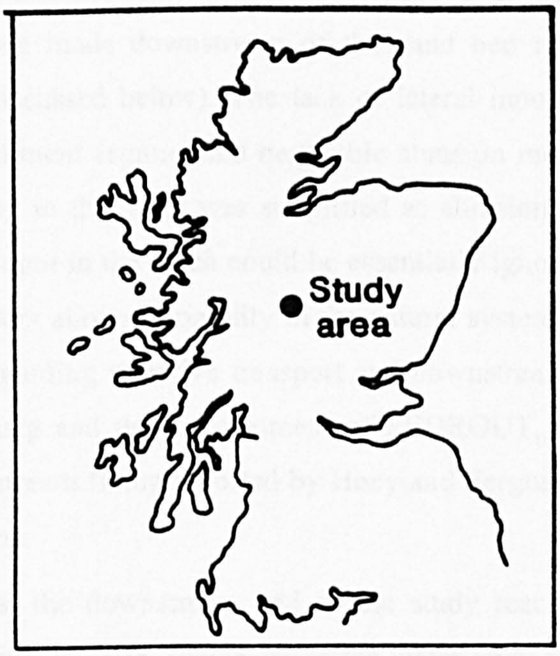


Figure 3.1: The location of Allt Dubhaig in Scotland. From Ferguson and Ashworth (1991).

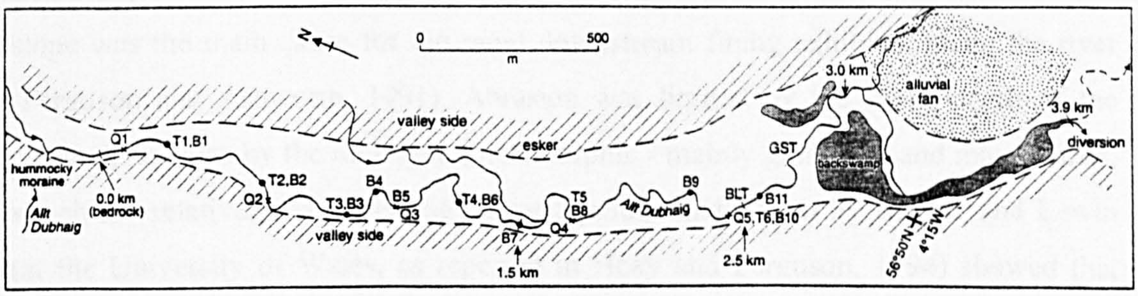


Figure 3.2: Map of Allt Dubhaig showing gauging stations (Q1 - Q5), bed-load trap (BLT), gravel-sand transition (GST), tracer-pebble seeding positions (T1 - T6), bulk bed sampling points (B1 - B11) and the diversion dam (diversion). From Ferguson *et al* (1996). The stream flow is from left to right.

The river has experienced little human disturbance in the recent past although some changes have been made downstream of the sand bed reach (in the form of the diversion dam, discussed below). The lack of lateral inputs, which could influence discharge and sediment regime, and negligible abrasion meant that investigating the downstream fining in this river was simplified as abrasion and anomalously fine or coarse sediment input in the reach could be essentially ignored. The removal of these complicating factors allows variability in the natural system to be constrained. Much previous work regarding selective transport and downstream fining has been carried out on the Dubhaig and the development of SEDROUT, the numerical model for simulating downstream fining reported by Hoey and Ferguson (1994;1997), uses this river as a prototype.

An alluvial fan at the downstream end of the study reach, and the presence of a diversion dam (built in 1935, which raises the water level by up to 1 m), imposed a local base level. Beyond the fan the channel steepened and returned to a gravel bed. The channel long profile was strongly concave and the water surface slope (presented in Figure 3.3) decreased from approximately 0.02 at the upstream end, to 0.002 towards the distal end of the gravel reach and then decreased rapidly with distance to 0.0002 in the region of the GST as the local base level was approached (Sambrook Smith, 1994). This decrease in the channel slope (and hence shear stress) was associated with a reduction in the bed grain size. The original channel long profile was inherited from the last deglaciation and it was hypothesised that the decrease in slope was the main cause for the rapid downstream fining exhibited along the river (Ferguson and Ashworth, 1991). Abrasion was limited by the short length of the channel, and also by the rock type (metamorphic - mainly granulites and mica schists, which are relatively hard wearing). Investigations undertaken by Brewer and Lewin (at the University of Wales, as reported in Hoey and Ferguson, 1994) showed that circular flume abrasion tests provided weight loss per kilometre of transport figures averaging 0.08%. According to these findings for the amount of downstream fining observed in Allt Dubhaig to take place solely through abrasion the particles would need to be transported for several thousand kilometres.

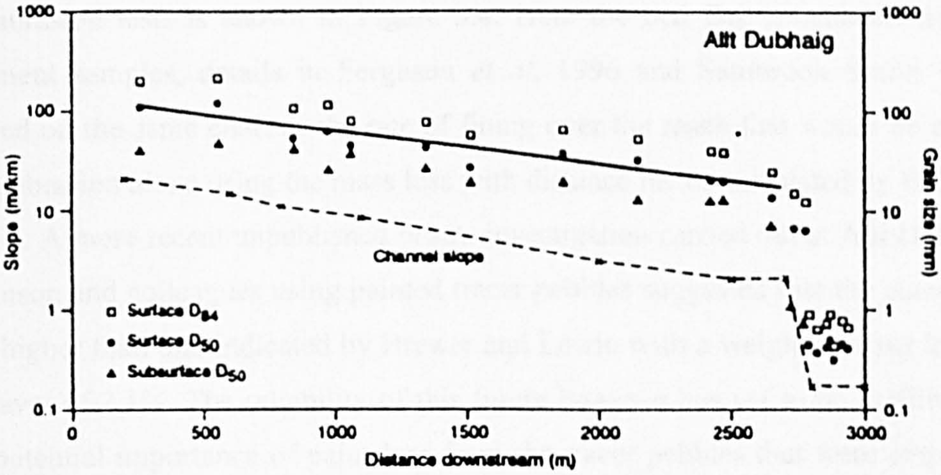


Figure 3.3: The water and bed surface slope of Allt Dubhaig. The variation in bed grain-size and the GST is also shown. From Sambrook Smith and Ferguson (1995).

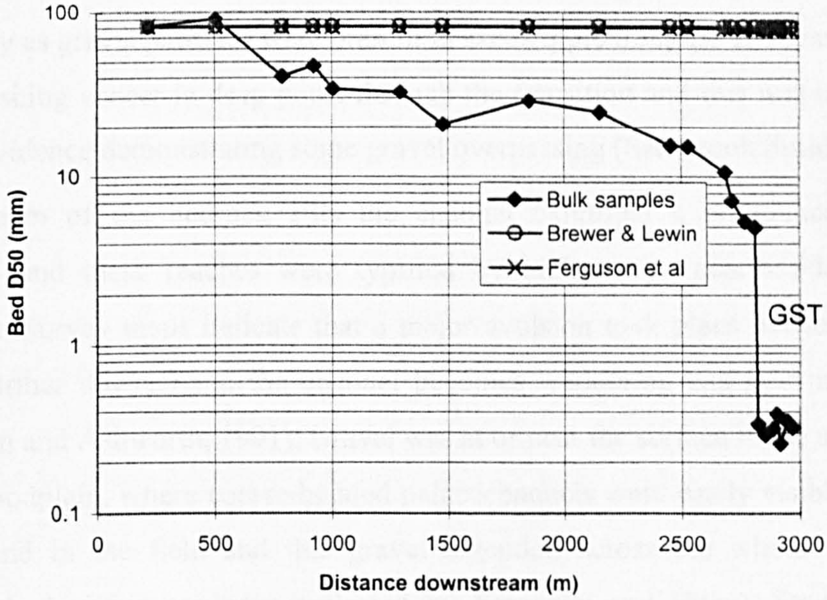


Figure 3.4: D_{50} of Allt Dubhaig bulk bed GSD samples with distance downstream compared with the rate of fining predicted by the abrasion tests of Brewer and Lewin and Ferguson *et al*. In the abrasion plots the grains are assumed to be ellipsoids. Field data provided by Ferguson and Sambrook Smith.

A plot of the downstream fining observed in Allt Dubhaig against that predicted by the abrasion tests is shown in Figure 3.4. Here the bed D_{50} (calculated from bulk sediment samples, details in Ferguson *et al*, 1996 and Sambrook Smith 1994) is plotted on the same chart as the rate of fining over the reach that would be produced from abrasion alone using the mass loss with distance factor calculated by Brewer and Lewin. A more recent unpublished *in situ* investigation carried out at Allt Dubhaig by Ferguson and colleagues using painted tracer pebbles suggested that the abrasion rate was higher than that indicated by Brewer and Lewin with a weight loss per kilometre of travel of 2.1%. The reliability of this figure however has yet to be verified due to the potential importance of paint loss from the tracer pebbles that were employed in the study. It is likely that the actual figure lies somewhere between the two values and it is therefore clear that other processes must be operating to cause the amount of fining observed.

Sambrook Smith (1994) noted the lateral sorting of gravel and sand sizes into adjacent tongues in the GST reach. He also noted that overpassing of the gravel over the sand was likely as gravel particles were present in sand-ripple troughs. The gravel remained as a vanishing veneer in deep pools through the transition and this was interpreted as further evidence demonstrating some gravel overpassing (Sambrook Smith, 1994).

Downstream of the bedrock sills the channel exhibited a pronounced step-pool sequence and these reaches were typified by high-energy near-braided channels. Ordnance Survey maps indicate that a major avulsion took place between 1860 and 1930. Further downstream the channel becomes wandering and then more straight (Ferguson and Ashworth, 1991). Gravel was at or near the surface in the upper reaches of the floodplain, where coarse-bedded palaeochannels were easily visible both in air photos and in the field and this gravel extended across the whole valley width (unpublished coring evidence collected by Ferguson and Hoey). Downstream the valley floor became marshy, possibly associated with the backwater effect of the dam. Coring of the floodplain in this region revealed that, away from the main channel, the sediments were predominantly fine grained silty sands with peat. Alluvial sediments were only found within about 20 m of the present day channel (Ferguson and Hoey, 1996, pers comm; Ferguson *et al*, 1996). From this information it was inferred that the lower reaches of the channel have migrated laterally very little in the late Holocene and the alluvial sediments have prograded over fine valley fill (possibly a relic sediment deposited in an ice-dammed lake). This evidence of upstream avulsion, with lateral downstream stability, together with progradation indicates that the channel has been aggrading, and continues to do so. This was due to the highly concave nature of the long profile which forced deposition downstream as shear stress decreased.

Data concerning the degree of selective transport occurring in the stream between 1991 and 1993 was obtained from a bedload trap in a straight near-rectangular reach approximately 300 m upstream of the GST (see Wathen *et al*, 1995 for full details). Discharge variations were logged at five sites in 1991-1993 (Q1-Q5 in Figure 3.2) and the record from the upstream gauging station (Q1) covered 1988-1997. The regime at this station was very flashy, with a bankfull discharge of $11 m^3/s$ which was exceeded several times a year in response to rain falling on snow present in the catchment, or locally heavy rain (Ferguson *et al*, 1996). The trap was emptied after almost every flood and showed that, below a peak shear stress of 11 Pa, negligible bedload transport occurred. As shear stress increased, however, transport rate increased and bedload became coarser with maximum bedload diameter increasing with peak stress (see Figure 3.5). Below 20 Pa the load was dominated by sand and above 30 Pa by gravel with bimodal sand and gravel being transported at floods with peak stresses between these values. The findings of Wathen *et al* (1995) showed the preferential transport of sand and finer gravels over coarser gravels. The coarse fractions of the bed were under-represented in the bedload, although less so with increasing shear stress. Averaging fractional transport rates over each flood event showed a small statistical departure from equal mobility for all sizes. Where sediment was > 2 mm in size the estimated hiding factor (x in Equation 2.14) was 0.10. This value is almost identical to the value specified in Parker's (1990) bedload function that is adopted for use in SEDROUT (see Chapter 7).

A reach of nearly 300 m, in which the GST occurs in the Dubhaig, was divided into 27 cross sections which were studied in detail. The number and sites of these sections was decided by Sambrook Smith and are discussed in his thesis (Sambrook Smith, 1994). For the purposes of this research it was deemed sensible to continue with investigations at these sections so that temporal as well as spatial change of GSTs could be characterised.

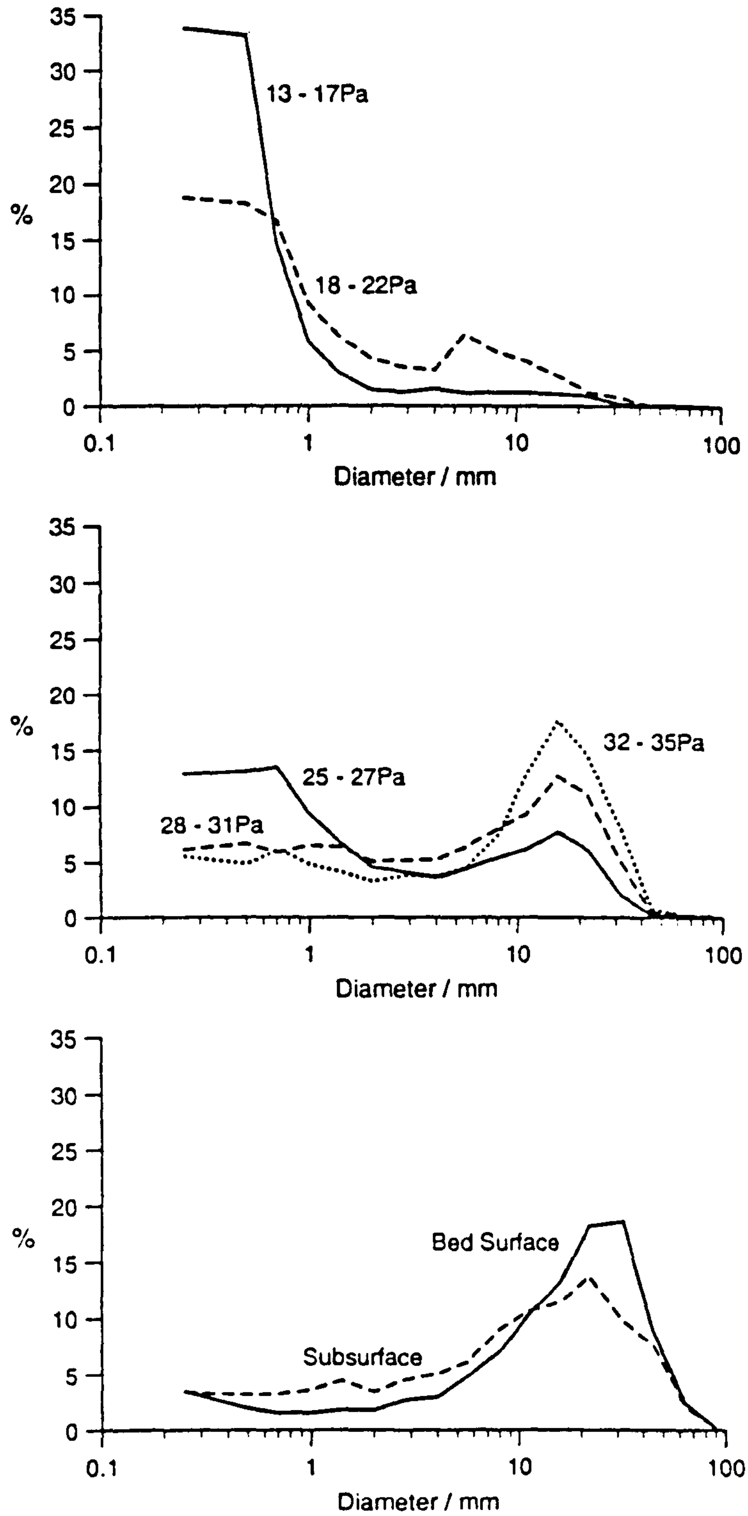


Figure 3.5: Allt Dubhaig bed load grain size distributions for the representative single events of different peak shear stresses. Shows the changing pattern of bimodality: modal fine and coarse sizes change little with event magnitude but their relative proportions alter. From Wathen *et al* (1995).

3.1.2 Vedder River

Vedder River was chosen for further study so that the broader applicability of the forms and processes thought to be important in the dynamics of gravel-sand mixtures could be tested in a river an order of magnitude larger than Allt Dubhaig. The transferability of SEDROUT can also be investigated on this second river. The Vedder has been studied in detail by the University of British Columbia (UBC) research group led by Mike Church.

Vedder River drains 1230 km^2 and is a south bank tributary of Fraser River about 160 km east of Vancouver, British Columbia, Canada (see Figure 3.6). The river flows off the Cascade Mountains (where it is called the Chilliwack River) through a rock ridge onto an aggrading alluvial fan at Vedder Crossing, downstream of which the channel is referred to as Vedder River. The river is constrained by flood dykes, and has only been flowing along its present course for about 100 years (McLean, 1980). A further 8 km downstream of Vedder Crossing the river was channelised in 1928 and is referred to as Vedder Canal. It is in this reach that the transition from a gravel to a sand bed occurs (Martin and Church, 1995).

Sinclair (1961) noted the major requirements of Vedder Canal: to contain the Fraser River backwater, which extended up Vedder River during the spring freshet; and to pass flood flows on Vedder River without scour to the banks, or deposition of sediments, which would lead to aggradation and therefore decreased channel capacity. The canal was planned to be self-scouring and although the design tried to account for movement of gravel in Vedder River it was still thought that it would be necessary to remove gravel bars from the entrance of the canal. The final design required 2.1 m of channel excavation with dykes spaced 152.4 m apart (at the crest). The channel slope was designed to be 0.00028. The canal was completed in 1924, and from aerial photographs it appeared that the bed of this channel was dominated by fine grained sediments (Church, 1998, pers comm). Since then Vedder River has been flowing down this straight route (McLean, 1980) and for the purposes of this research, this situation is ideal, as the river can essentially be considered as a large natural flume. McLean (1980) noted that in 1963 the channel slope in the Canal was 0.00032. This indicates that some aggradation had taken place since the Canal was completed. The presence of alternating bars in the Canal was also noted by McLean, who suggested that the river was attempting to develop meanders in this reach.

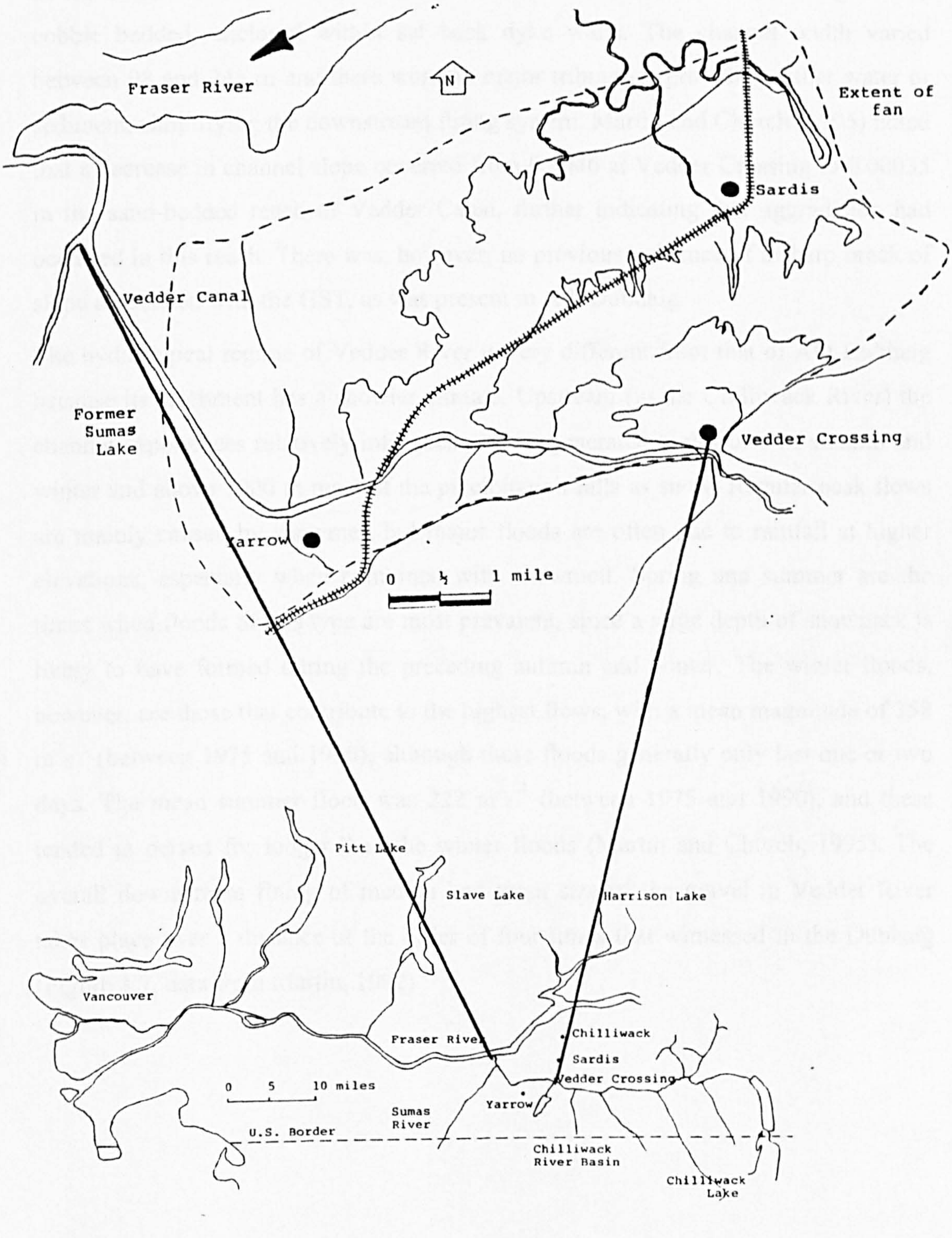


Figure 3.6: The location of Vedder River in British Columbia, Canada. Note the straight "Vedder Canal". From McLean (1980).

In the 8 km reach of Vedder River upstream of the Canal the channel is generally cobble bedded, enclosed within set back dyke walls. The channel width varied between 98 and 245 m and there were no major tributaries providing either water or sediment, simplifying the downstream fining system. Martin and Church (1995) noted that a decrease in channel slope occurred from 0.0046 at Vedder Crossing to 0.00035 in the sand-bedded reach of Vedder Canal, further indicating that aggradation had occurred in this reach. There was, however, no previous evidence of a sharp break of slope associated with the GST, as was present in Allt Dubhaig.

The hydrological regime of Vedder River is very different from that of Allt Dubhaig because its catchment has a snowier climate. Upstream (in the Chilliwack River) the channel experiences relatively infrequent, storm generated high flows in autumn and winter and above 1000 m most of the precipitation falls as snow. Regular peak flows are mainly caused by snowmelt but major floods are often due to rainfall at higher elevations, especially when combined with snowmelt. Spring and summer are the times when floods of this type are most prevalent, since a large depth of snowpack is likely to have formed during the preceding autumn and winter. The winter floods, however, are those that contribute to the highest flows, with a mean magnitude of 358 m^3s^{-1} (between 1975 and 1990), although these floods generally only last one or two days. The mean summer flood was 222 m^3s^{-1} (between 1975 and 1990), and these tended to persist for longer than the winter floods (Martin and Church, 1995). The overall downstream fining of median bed grain size of the gravel in Vedder River takes place over a distance of the order of four times that witnessed in the Dubhaig (Figure 3.7, data from Martin, 1992).

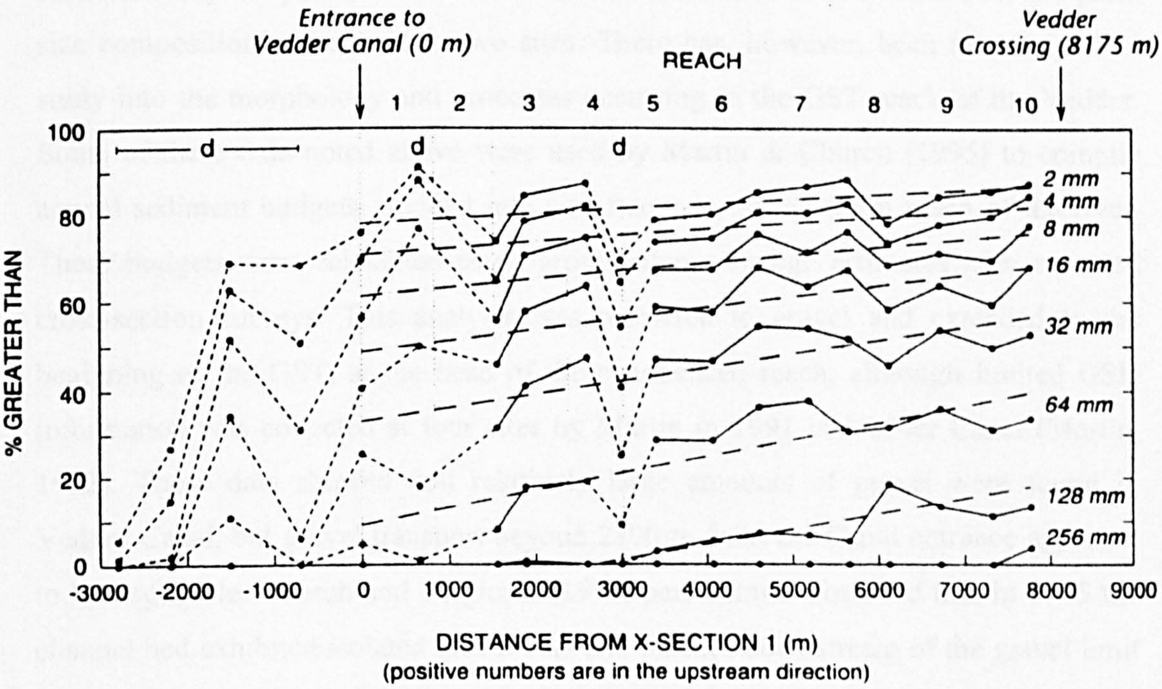


Figure 3.7: GSDs of Vedder River represented as a proportion of material coarser than given sizes at various positions along the stream. The trend lines are regression fits to the data within the surveyed reach. The data marked ‘d’ were deleted from the analysis because “do not fit the trend evident in the balance of the data”. From Martin and Church (1995).

Like Allt Dubhaig, Vedder River has a large database of morphological, sedimentological, hydrological, historical and process data as the river was studied by various workers attached to the UBC research group, together with provincial engineers. This database includes streamflow records, surveys of closely spaced cross sections every 1-3 years, detailed bulk GSDs, and measurements of the flux and grain size composition of bedload at two sites. There has, however, been relatively little study into the morphology and processes occurring in the GST reach of the Vedder. Some of these data noted above were used by Martin & Church (1995) to compile annual sediment budgets, divided into size fractions, for an 8 km reach of the river. These budgets were calculated using gross volume change estimates after repeated cross-section surveys. This analysis was restricted to gravel and extended to the beginning of the GST, at the head of the channelised reach, although limited GSD information was collected at four sites by Martin in 1991 in Vedder Canal (Martin, 1992). These data showed that relatively large amounts of gravel were found in Vedder Canal, but gravel transport beyond 2800 m from the Canal entrance appeared to be negligible. Church and Ferguson (1997, pers comm) observed that in 1995 the channel bed exhibited isolated gravel bars and patches downstream of the gravel limit inferred from the limited GSD data collected in the Canal by Martin (in 1991); this suggests the GST is prograding and gives an opportunity to investigate the processes involved. As noted above, there has only been very limited investigation into the morphology and sedimentology of Vedder Canal. A new framework for analysis was therefore required. Twelve cross sections were set up across the channel at approximately 300 m intervals. These were numbered N1 to N12 in downstream order and data collection similar to that done on the sections in the GST reach of Allt Dubhaig was carried out at each of these. Four of these sections coincided with repeat grain size sampling of Vedder Canal undertaken by Martin in 1991.

3.2 The GST in Allt Dubhaig and Vedder River: similarities and differences.

From the details of the two rivers outlined above it is clear that both have gravel-sand sediment mixtures on the bed and exhibit a spatially rapid GST. There are, however important differences between them including the fact that they are an order of magnitude different in width, have been influenced by differing anthropogenic activities and have a different environmental history. Both the Dubhaig and the Vedder have had some degree of human influence, which has altered the sedimentary dynamics of the reaches which contain the contemporary GSTs. As this anthropogenic activity is vastly different in both rivers, however, it should not be thought of as the main cause of rapid downstream fining between the gravel and sand beds. Choosing these fieldsites with their contrasting physical environments is likely to give a better constraint on the important processes occurring in streams with gravel-sand sediment mixtures present on the bed. Neither river has any major inputs of water or sediment in the study reach, and both have a relatively long history of study.

Both Allt Dubhaig and Vedder River exhibit downstream fining of the river gravels, a prograding gravel front upstream of a spatially rapid GST and aggradation occurring in their gravel reaches. While the Dubhaig GST is coincident with a sharp reduction in slope there is no evidence to date to suggest that a break of slope is present in the region of the GST in the Vedder. Vedder River is also less strongly concave in its gravel reach than the Dubhaig.

Martin and Church (1995, p.351) noted that "the sand fraction of the bed material in Vedder River falls within the range 14-22% and is matric infill within the gravel framework". This interpretation follows that of Ferguson and Ashworth (1991, p.69), who, when discussing the Dubhaig, inferred "that sand is winnowed from between surface cobbles and infiltrates to form a matrix at depth". In their study, Martin and Church (1995) treated sand as wash load, and chose to restrict their sediment budget of the Vedder to gravel. In Vedder River, where the main sediment budget investigation was carried out, this simplification may be valid, but in the upper reaches of Vedder Canal, and downstream where the percentage of fine grained sediment in the bed increases, the importance of sand for fluvial processes needs to be considered in detail. Ferguson and Ashworth (1991, p.69), following field observations in the upper and middle reaches of the Dubhaig, stated that "surface sand is restricted to transient pockets in the lee of protruding cobbles, until just past (the GST) reach". Similar small patches of sand immediately downstream of obstacles can be seen in Vedder Canal, upstream of the GST. Work carried out on both rivers has noted, therefore, that sand was present on the surface of the gravel bed. This sand, however, exists in too small a quantity in a sufficiently high-energy environment and its influence on bed surface sedimentology is minimal. Further downstream, however, as the magnitude of the fine mode increases, and bed shear stress decreases, the importance of sand in terms of bed surface sedimentology is likely to increase.

Chapter 4. Characteristics of contemporary Gravel-Sand Transitions

This chapter will provide details of the characteristics of the contemporary GSTs in Allt Dubhaig, Scotland and Vedder Canal, Canada. All the data presented here were collected during the current research. The techniques of data collection will be outlined and the results presented. In all cases, methods and results from Allt Dubhaig are outlined first, followed by those from Vedder Canal. In Chapter 5 the data will be compared with those collected in both rivers during previous research to elucidate changes in the characteristics of the GST zones over time. In addition, the influence that altering the sampling techniques can have on the results will be analysed. The implications of the results presented here will be discussed in detail in Chapter 9.

Referring back to Chapter 1, the main objectives of fieldwork were: to assess what forms and processes are occurring in the GST reach of the rivers and to outline how the transition zones change in the short to medium term. Using this information the accuracy on numerical models of gravel-sand stream behaviour and the conceptual model of GST form can be developed.

4.1 Framework for analysis

4.1.1 Allt Dubhaig

As outlined in the previous chapter, the GST of Allt Dubhaig has been described in some detail by Sambrook Smith (1994) and Sambrook Smith and Ferguson (1995). Twenty-seven cross sections were set up by Sambrook Smith in 1992, perpendicular to the flow direction, from the distal-most gravel bar (last gravel bar) at cross section 1, through the united gravel-sand bed reach, to the sand-dominated channel bed (a distance of almost 300 m). The positions of the cross sections are shown in Figure 4.1.

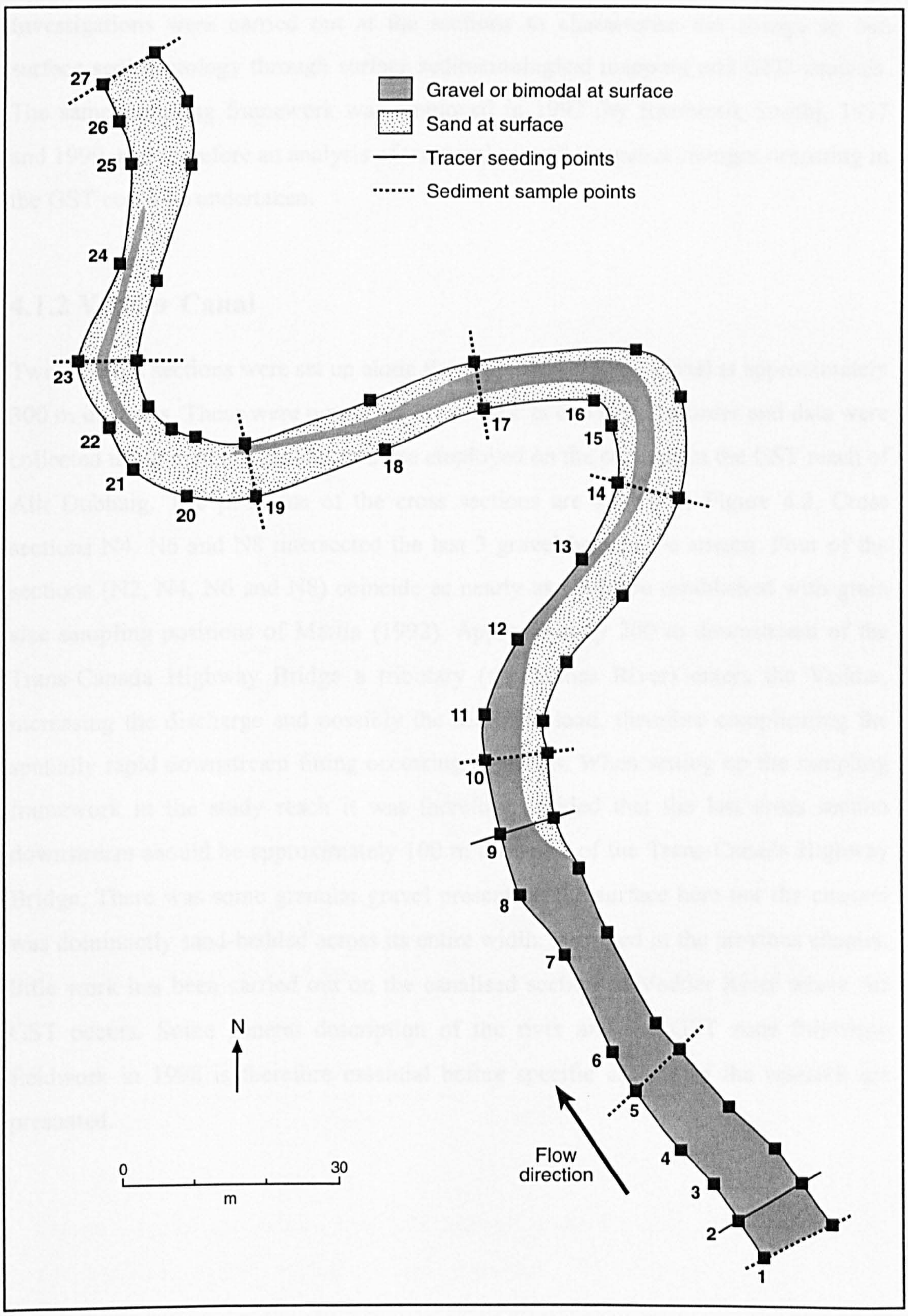


Figure 4.1: The gravel-sand transition reach of Allt Dubhaig in 1997. Positions of cross sections, sediment sampling points, tracer seeding points and bed surface sedimentology are shown.

Investigations were carried out at the sections to characterise the change in bed surface sedimentology through surface sedimentological mapping and GSD analysis. The same sampling framework was employed in 1992 (by Sambrook Smith), 1997 and 1999, and therefore an analysis of temporal as well as spatial changes occurring in the GST could be undertaken.

4.1.2 Vedder Canal

Twelve cross sections were set up along the channel in Vedder Canal at approximately 300 m intervals. These were numbered N1 to N12 in downstream order and data were collected using methods similar to those employed on the sections in the GST reach of Allt Dubhaig. The positions of the cross sections are shown in Figure 4.2. Cross sections N4, N6 and N8 intersected the last 3 gravel bars in the stream. Four of the sections (N2, N4, N6 and N8) coincide as nearly as could be established with grain size sampling positions of Martin (1992). Approximately 200 m downstream of the Trans-Canada Highway Bridge a tributary (the Sumas River) enters the Vedder, increasing the discharge and possibly the sediment load, therefore complicating the spatially rapid downstream fining occurring upstream. When setting up the sampling framework in the study reach it was therefore decided that the last cross section downstream should be approximately 100 m upstream of the Trans-Canada Highway Bridge. There was some granular gravel present on the surface here but the channel was dominantly sand-bedded across its entire width. As noted in the previous chapter, little work has been carried out on the canalised section of Vedder River where the GST occurs. Some general description of the river and the GST zone following fieldwork in 1998 is therefore essential before specific aspects of the research are presented.

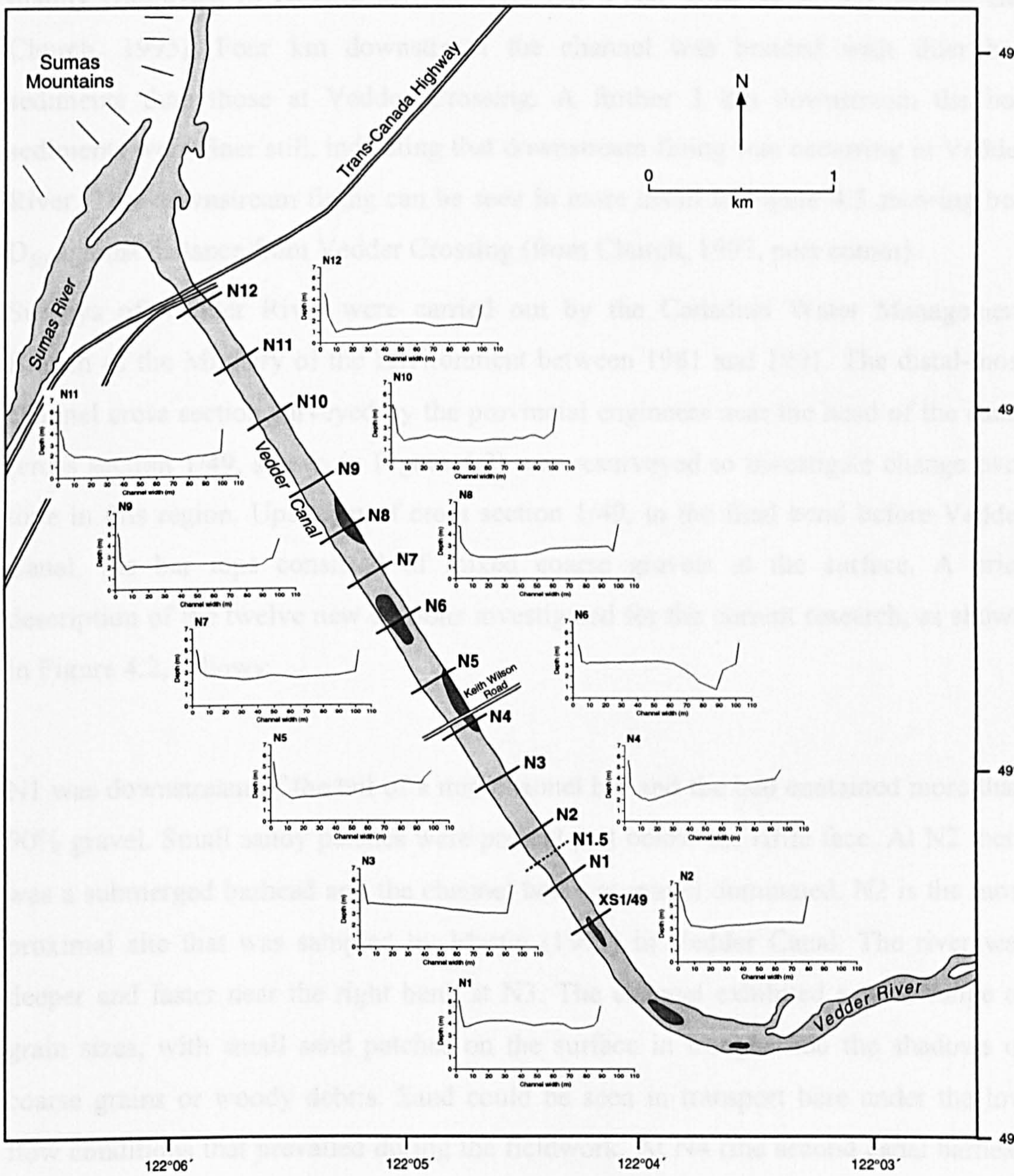


Figure 4.2: The gravel-sand transition reach of Vedder Canal in 1998. Positions of cross sections and cross section profiles are plotted. Gravel bars are shown as darker areas in the stream. Flow is from bottom right to top left.

Preliminary Observations

At Vedder Crossing (shown in Figure 3.6 in the previous chapter) the channel bed was mainly comprised of cobbles and the river has a bed slope of 0.0046 (Martin and Church, 1995). Four km downstream the channel was braided with finer bed sediments than those at Vedder Crossing. A further 3 km downstream the bed sediments were finer still, indicating that downstream fining was occurring in Vedder River. This downstream fining can be seen in more detail in Figure 4.3 showing bed D_{50} against distance from Vedder Crossing (from Church, 1997, pers comm).

Surveys of Vedder River were carried out by the Canadian Water Management Branch of the Ministry of the Environment between 1981 and 1991. The distal-most channel cross section surveyed by the provincial engineers near the head of the canal (cross section 1/49, shown in Figure 4.2) was resurveyed to investigate change over time in this region. Upstream of cross section 1/49, in the final bend before Vedder Canal, the bar tops consisted of mixed coarse gravels at the surface. A brief description of the twelve new sections investigated for the current research, as shown in Figure 4.2, follows:

N1 was downstream of the tail of a mid channel bar and the bed contained more than 90% gravel. Small sandy patches were present just below the riffle face. At N2 there was a submerged barhead and the channel bed was gravel dominated. N2 is the most proximal site that was sampled by Martin (1992) in Vedder Canal. The river was deeper and faster near the right bank at N3. The channel exhibited a wide range of grain sizes, with small sand patches on the surface in troughs and the shadows of coarse grains or woody debris. Sand could be seen in transport here under the low flow conditions that prevailed during the fieldwork. At N4 (the second canal barhead sample site of Martin (1992)) the bed was mainly gravel-framework with some interstitial sand but there was much local variability in surface grain size. More sand was present at the surface than was evident at the cross sections upstream. N5 was mainly gravel framework with some sandy patches. The bed at N6 was mainly gravel framework with some sand. There was a gravel bar present at this section, although Church and Ferguson (1998, pers comm) remember this site to have been more sandy in 1995. N6 was the third canal barhead sample site of Martin (1992).

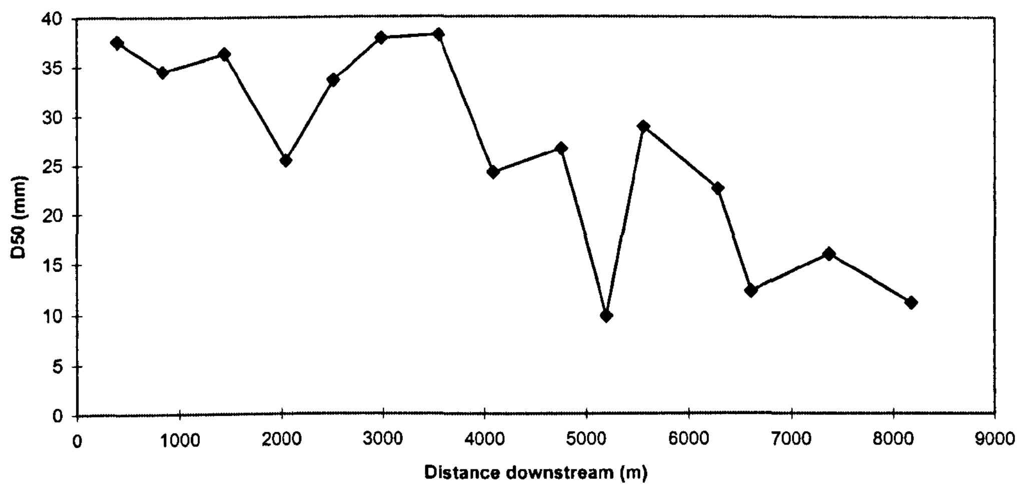


Figure 4.3: Change in bed D_{50} in Vedder River from Vedder Crossing to the head of the canal (from Church, 1997, pers comm).

There was more sand on the bed surface than upstream and a medial gravel bar was present in the channel. The channel bed was dominated by a gravel framework. Downstream of N6 sand streaks began to appear on the bed surface. The bed was beginning to switch from a framework supported gravel bed with some sand, to a matrix supported sand bed with some gravel. Downstream there were fine gravel deposits in the troughs of large sand ripples/dunes. Approximately 100 m downstream of the tail of the N6 bar was N7. N8 was at the head of a right bank gravel bar, which was the last visible gravel bar. This section was the fourth and distal-most canal barhead sample site of Martin (1992). In 1998 the bed was mainly a mix of gravel framework with some sand and a sand matrix with some gravel. Downstream of N8 the GST occurred and N10 exhibited some of the documented features of a GST with large sand patches and streaks present on the bed (see Sambrook Smith, 1994). N11 and N12 were dominantly sand bedded with ripple and dune features present. Approximately 400 m upstream of the Highway Bridge the channel was sand bedded, although some granular gravel was present, especially in troughs. At the distal sand-dominated sections there was a pattern of submerged sand sheets with sinuous thalwegs around them.

4.2 Methods of measurement

As noted in the introduction, fieldwork was undertaken to characterise the form of a GST in two different rivers. The data necessary for understanding the morphology of the GST reach include the long profile, channel cross sections and bed GSDs. Information concerning the bed surface sedimentology is useful for characterising the GST zone and how this differs from the gravel-dominated or sand-dominated reaches of the channel bed. Data were therefore collected in the field using a combination of water surface and cross section surveys, bulk bed samples for GSD analysis and mapping the bed surface sedimentology in the GST zone. Linking the surficial and bulk grain size samples will allow a comparison of bulk against areal extent of different GSDs in the transition zone. The information gained from this research will be linked to probing data collected for the current study, in the GST reach of Allt Dubhaig, and surface sedimentological data collected previously and discussed in

Chapter 5 to assess change over time in the GST zone. A detailed methodology of the field investigations can be found in the following section.

4.2.1 Long profile surveys

In Scotland the water surface elevation was measured from the bedload trap (shown in Figure 3.2) through to the sand-dominated reach, including each of the 27 cross sections. This survey was carried out using a Leica total station.

In Canada a long profile survey was carried out with a Sokkisha automatic level. The cross section end pegs were levelled along the length of Vedder Canal (approximately 4 km) in three successive traverses along the left bank, each of which was closed to identify errors. The closure errors were divided evenly across all points and are shown in Table 4.1. The end peg points were combined with cross section surveys (see Section 4.2.2 below) to give bed and water surface elevations. These were then combined and tied into known Survey of Canada datum points near the river.

Table 4.1: Closure errors of the cross section peg survey along Vedder Canal.

Survey end points	Distance between end points (m)	Closure error (m)
N1 - N4	1149	-0.15
N3 - N8	1606	0.03
N8 - N12	1354	0.11

4.2.2 Cross section surveys

Allt Dubhaig cross section surveys were carried out through the transition zone at each of the 27 cross sections noted above. Both the 1992 (Sambrook Smith) and 1997 (current study) surveys were tied to the same datum to allow direct comparison. Measurements were taken at approximately half metre intervals at each section from the left bank peg to the right bank peg. It should be noted that the error in height measurements taken with the total station depends on the distance between the EDM and the reflecting pole. Approximate errors may be of the order +/- 0.05 m over the distances measured for the cross section surveys (Hoey, pers comm, in Sambrook Smith, 1994). Errors will increase with distance between the EDM and reflecting pole.

In Vedder Canal cross section surveys were carried out with a Sokkisha automatic level. Cross stream survey intervals at each cross section were approximately 5 m. Cross section 1/49 was also resurveyed to allow comparison with earlier surveys by engineers from the Canadian Water Management Branch.

4.2.3 Bed surface sedimentology

In Allt Dubhaig the bed surface sedimentology was mapped during the cross section surveys carried out in 1992 (by Sambrook Smith) and 1997 (for the current study). In 1997 the type of material present on the bed surface was noted each time the reflector pole for the total station was moved. The bed was described as gravel, sand or bimodal depending on the proportions of each sediment type.

In Vedder Canal the changes in bed surface sedimentology were noted each time the surveying staff was moved to a new survey point. A technique adapted from Kodama (1994b) was employed because of the large range of sediment types present on the bed surface in Vedder Canal. This involves a qualitative classification of the sediments to understand the spatial variation in grain size occurring in the GST. The scheme was not utilised when describing the Dubhaig bed sediments because the 1992 survey had not employed this method and so comparison over time would be more difficult if another notation scheme was introduced. The bed sediment changes also occurred over a shorter distance in the Dubhaig, both laterally and longitudinally, compared to the Vedder. The notation for the scheme used in the Vedder is shown in Table 4.2.

Table 4.2: Notation scheme for classifying the bed surface sedimentology in Vedder Canal. Adapted from Kodama (1994b).

Notation scheme	Bed surface sedimentology
1	100% gravel (negligible sand)
2	90-100% gravel
3F	Gravel framework with some sand
3M	Sand matrix with some gravel
4	90-100% sand
5	100% sand (negligible gravel)

Kodama was consulted during the preparation of this scheme (in 1998) to clarify how to classify the sedimentology since the original paper was written in Japanese. The order of classification was reversed, from that presented in Kodama (1994b), so that lower numbers indicated gravel beds and higher numbers indicated sand beds. In the case of a river bed exhibiting downstream fining, therefore, the numbers generally increase downstream. The classification containing negligible sand (1) was altered from the original scheme where it was used to classify eroding areas on bar surfaces. Type 3 originally covered a wide range of surface sediment types and this classification was split into two depending on whether the surface was dominated by sand matrix or gravel framework supported sediments.

4.2.4 Bed GSDs

In Allt Dubhaig bed surface samples were collected for the present research at various positions in the transition zone in 1997 and 1999. The method of Sambrook Smith (1994) was followed for the current research to allow temporal change to be investigated. This investigation into change over time is presented in Chapter 5. It should be noted that, to quantify the patchiness of a transition zone, Sambrook Smith took samples "from different facies within the mixed sediments of the transition reach" (Sambrook Smith, 1994, p.39). These samples were taken, however, without regard to the width of channel occupied by the different sedimentological units, and the specific GSDs of the different patches were not investigated. Rather, the individual samples were bulked together and sieved as one. In 1999 samples were collected and sieved from the different patches to illustrate the different GSDs present. These individual samples were then combined to give a width-integrated GSD for the particular distance downstream with each patch sample weighted depending on the proportion of the width occupied by the different facies. The positions of these samples was decided after consulting the bed surface sedimentological map surveyed in 1997. Where only one sample was taken at a particular cross section in 1997 and 1999, individual subsamples were taken across the whole channel width at spacings of no more than 0.5 m to ensure the sample was spatially homogeneous. A list of the positions and numbers of the samples collected for GSD analysis are shown in Table 4.3 and positions of the patch samples is shown in Table 4.4.

Table 4.3: Numbers of samples collected at different cross sections in different years at Allt Dubhaig. Note: samples collected in 1992 were reported in Sambrook Smith (1994).

Year	Cross section number							
-	1	5	10	14	17	19	23	27
1992	1	1	1	1	1	1	1	1
1997	1	1	1	1	1	0	0	0
1999	1	1	2	3	2	1	1	1

Table 4.4: Lateral positions of patch grain size samples collected in 1999 in the gravel-sand reach of Allt Dubhaig. Distance across the channel is shown (in m) and the sedimentology is included in brackets. RB and LB are right bank and left bank respectively.

Cross section number		
10	**14**	**17**
LB-7 (gravel)	LB-5 (sand)	LB-5 (sand)
7-RB (sand)	5-6 (bimodal)	5-8 (bimodal)
-	6-RB (sand)	-

The depth of the largest clast was taken as the maximum sampling depth and sediments were sieved at half-phi size intervals to 0.18 mm. Following the sample size criteria of Church *et al* (1987), the largest grain was never more than 1% of the total mass of the sample.

In the Vedder the bed at sections N4 to N12 was sampled to quantify bulk bed GSD change both upstream and downstream of the GST. This sampling included three of the sites (N4, N6 and N8) featured in Martin (1992), to investigate change over time. These sites were excluded from the analysis in Martin (1992) and Martin and Church (1995, p.353) because "they do not fit the trend evident in the balance of the data" which was mostly collected in the gravel reach upstream of Vedder Canal. At these sites subsurface samples were taken at bar head locations using the same technique as Martin (1992) which was documented in Church *et al* (1987). The surface sediments were removed to the depth of the largest grain over an area of approximately 2 m by 2 m. The subsurface sediments were then weighed and sieved to 0.18 mm such that the mass of the largest grain never exceeded 1% of the total sample size. This technique can only be carried out on exposed bars since, if sampling was carried out under water, the fine sediments from the subsurface would be winnowed once the coarse surface layer had been removed.

Samples were also taken at N4 to N12 using a dredge technique so that variations in the GSDs produced by the two sampling methods could be investigated and grain size changes through the GST could be characterised. No gravel bars were present between N1 and N3 so the Martin (1992) barhead site at N2 could not be resampled using the technique of Church *et al* (1987). Consequently it was decided to take a dredge sample at a cross section with a suitable access point. This was found approximately 50 m upstream of N2 and was called N1.5 (see Figure 4.2). The dredge technique involved dropping a collection device (shown in Figure 4.4) onto the bed of the channel and dragging it along to scoop up and retrieve sediments. The mesh on the dredge was 0.2 mm and its mouth aperture was 160 mm. Dredges were taken across the whole channel width at spacings of no more than 10 m to ensure the samples were spatially homogenous. This type of sampling could be carried out in the wetted perimeter of the channel unlike, the barhead method above. It was therefore useful for gravel-dominated sections which had no exposed bars but where large weights of sample were required to be representative of the bed grain size population as a whole. In the sand-gravel and sand reaches a number of different dredge samples were taken across each of four sections (N9 to N12), in areas of different surface sedimentology to quantify the degree of lateral variation in GSD of the bed sediments. A list of the positions of these individual samples is shown in Table 4.5. A width-averaged GSD was also calculated for these sections with each patch sample weighted depending on the proportion of channel width occupied.

Table 4.5: Lateral positions of facies grains size samples collected in the gravel-sand reach of Vedder Canal. Distance across the channel is shown (in m) and the sedimentology (Kodama notation) is included in brackets. RB and LB are right bank and left bank respectively.

Cross section number			
N9	**N10**	**N11**	**N12**
LB-65 (4/5)	LB-18 (4)	LB-17 (3M)	LB-64 (4/5)
65-RB (5)	18-43 (4/5)	17-49 (4)	64-RB (4)
-	43-RB (4)	49-RB (5)	-

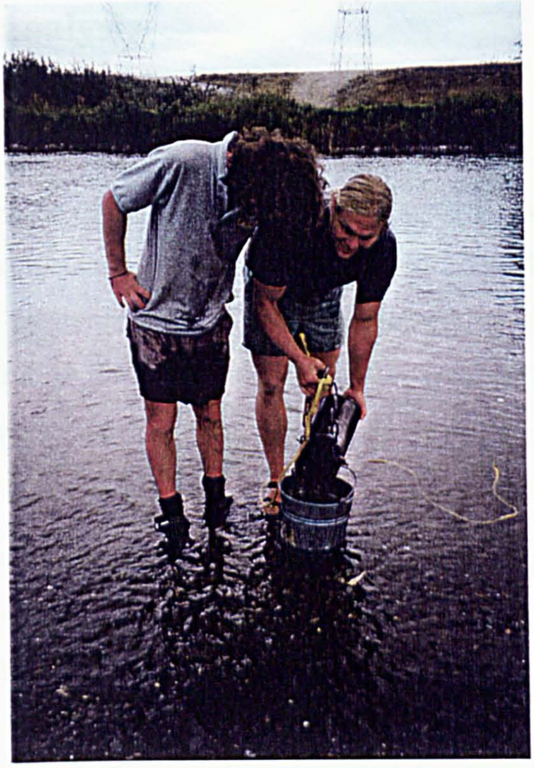


Figure 4.4: Pipe dredge used to collect bed surface sediment samples from Vedder Canal.

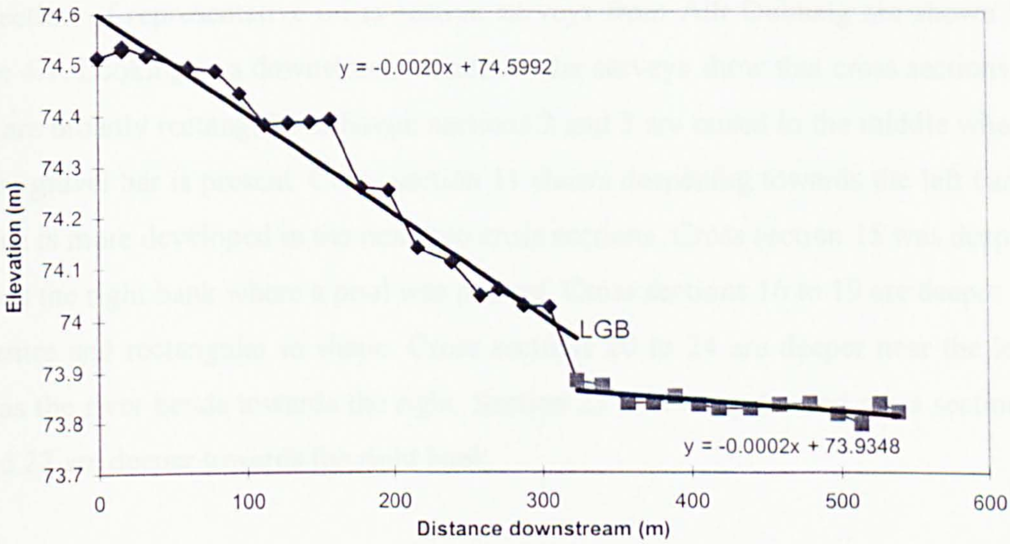


Figure 4.5: Allt Dubhaig water surface profile across the gravel-sand transition. Position of the last gravel bar (LGB) is shown.

4.3 Results of analysis

4.3.1 Long profile surveys

The water surface profile of the Dubhaig taken at low flow shows that a break of slope occurred a very short distance downstream of the last gravel bar (see Figure 4.5). The actual GST occurs approximately 60 m downstream of the break of slope. The gradients before and after the break of slope in the water surface profile were 0.0020 and 0.0002 respectively.

After plotting both the bed surface and the water surface long profile of the Vedder it was evident that there was no break of slope in the region of the GST (see Figure 4.6). The most obvious change in gradient between each of the surveyed sections occurred at N4. Gradients for the bed slope upstream and downstream of this point were 0.0008 and 0.0005 respectively. Taking all 12 sections together the gradient was 0.0006.

4.3.2 Cross section surveys

A selection of representative cross section surveys from Allt Dubhaig are shown in Figure 4.7. Looking in a downstream direction, the surveys show that cross sections 1 to 10 are broadly rectangular although sections 2 and 3 are raised in the middle where the last gravel bar is present. Cross section 11 shows deepening towards the left bank and this is more developed in the next two cross sections. Cross section 15 was deeper towards the right bank where a pool was present. Cross sections 16 to 19 are deeper in the centre and rectangular in shape. Cross sections 20 to 24 are deeper near the left bank as the river bends towards the right. Section 25 is rectangular and cross sections 26 and 27 are deeper towards the right bank.

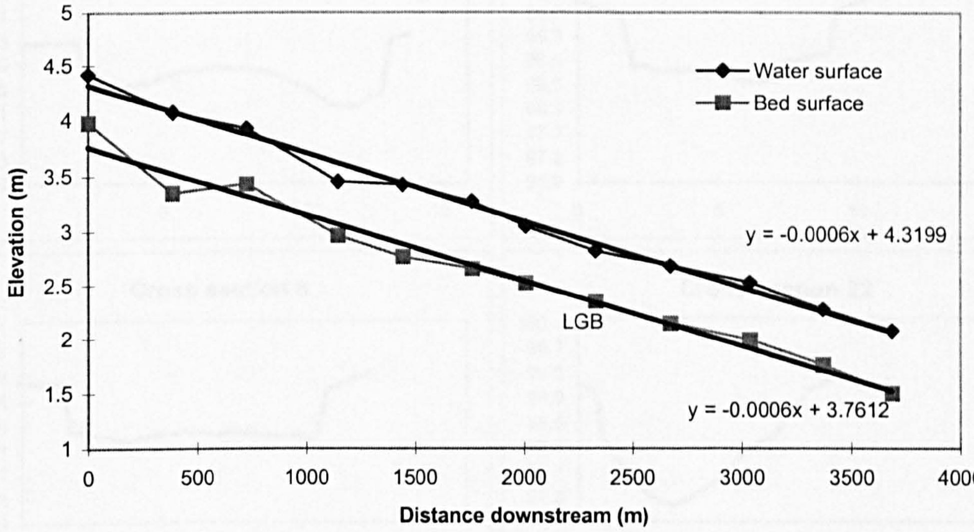


Figure 4.6: Vedder Canal water and bed surface profiles across the gravel-sand transition. Position of the last gravel bar (LGB) is shown.

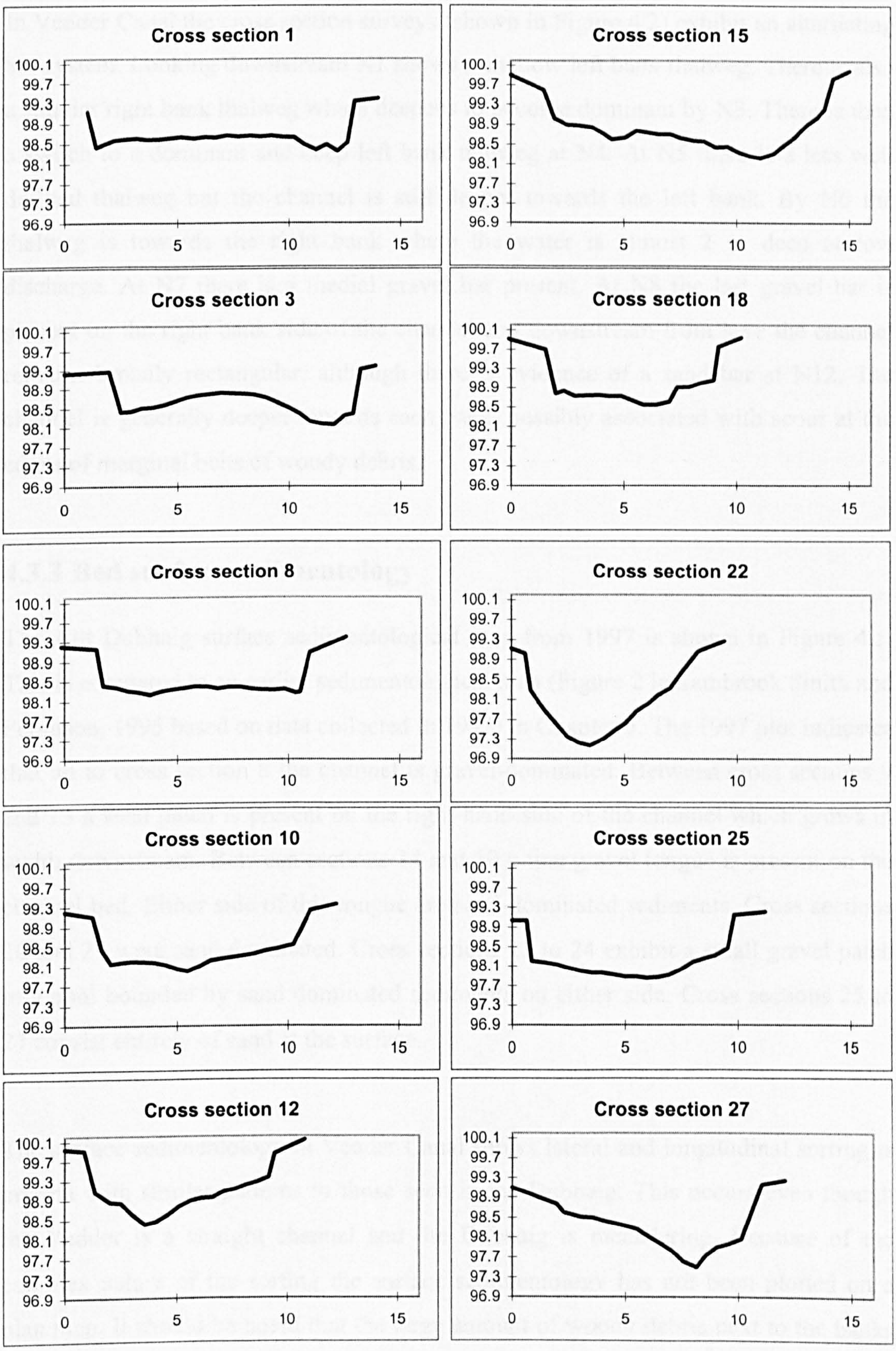


Figure 4.7: Allt Dubhaig cross section surveys mapped in 1997 (see Figure 4.1 for locations). All axes distances are in metres.

In Vedder Canal the cross section surveys (shown in Figure 4.2) exhibit an alternating bar system. Looking downstream N1 shows a shallow left bank thalweg. There is also a smaller right bank thalweg which deepens to become dominant by N3. There is then a switch to a dominant and deep left bank thalweg at N4. At N5 there is a less well defined thalweg but the channel is still deeper towards the left bank. By N6 the thalweg is towards the right bank where the water is almost 2 m deep at low discharge. At N7 there is a medial gravel bar present. At N8 the last gravel bar is present on the right bank side of the channel and downstream from here the channel remains broadly rectangular, although there is evidence of a sand bar at N12. The channel is generally deeper towards each bank, possibly associated with scour at the edges of marginal belts of woody debris.

4.3.3 Bed surface sedimentology

The Allt Dubhaig surface sedimentological map from 1997 is shown in Figure 4.1. This is compared to an earlier sedimentological map (Figure 2 in Sambrook Smith and Ferguson, 1995 based on data collected in 1992) in Chapter 5. The 1997 plot indicates that up to cross section 8 the channel is gravel-dominated. Between cross sections 9 and 13 a sand patch is present on the right-hand side of the channel which grows in width downstream. Between sections 14 and 19 a fine gravel tongue is present on the channel bed. Either side of this tongue are sand-dominated sediments. Cross sections 20 and 21 were sand dominated. Cross sections 22 to 24 exhibit a small gravel patch in a pool bounded by sand dominated sediments on either side. Cross sections 25 to 27 consist entirely of sand at the surface.

The surface sedimentology in Vedder Canal shows lateral and longitudinal sorting is present with similar patterns to those seen in the Dubhaig. This occurs even though the Vedder is a straight channel and the Dubhaig is meandering. Because of the complex nature of the sorting the surface sedimentology has not been plotted on a plan map. It should be noted that the large amount of woody debris next to the banks of Vedder Canal often tends to trap finer sediments which are not present in the more central parts of the channel where the flow velocity was greater.

Figure 4.8 shows a plot of the percentage of the channel width of Vedder Canal that was occupied by the different sedimentological units. According to the Kodama-type survey N1 is largely dominated by gravel containing less than 10% sand. At N2, N3 and N4 the dominant bed type is a gravel framework, which supports some interstitial sand. The same was true at N5 although the proportion of sand increased towards the banks. Near the right bank this is due to an exposed bar, from which the fine sediments are not winnowed. N6, N7 and N8 all have a dominantly gravel framework bed containing sand. The proportion of sand matrix-supported bed at these sections increases downstream. The bed at N9 contains sand matrix sediments with some gravel together with sands containing very little gravel. The surface sedimentology indicates that, apart from a small area towards the left bank of N11, none of the sediments downstream of N9 contains more than 10% gravel. Throughout the transition reach downstream of N8 the degree of lateral variability in bed surface sedimentology increases greatly, and this is reflected by the GSDs taken from these positions and presented in the following section.

4.3.4 Bed GSDs

Tables 4.6, 4.7 and 4.8 show details of the GSD samples collected in the Dubhaig and Vedder. The bimodality index (B*) is taken from Sambrook Smith *et al* (1997). The index is calculated as follows:

$$B^* = |\phi_2 - \phi_1| (F_2 / F_1) \qquad \text{Equation 4.1}$$

where ϕ_1 and ϕ_2 are the grain sizes (in half phi fractions) of the dominant and secondary modes respectively and F_1 and F_2 are the proportions of the bed sediment in the dominant and secondary modes respectively (in half phi fractions). Sambrook Smith *et al* (1997, p.1180) state that "the critical value of B* that defines whether a sediment is unimodal or bimodal lies in the range 1.5 - 2.0".

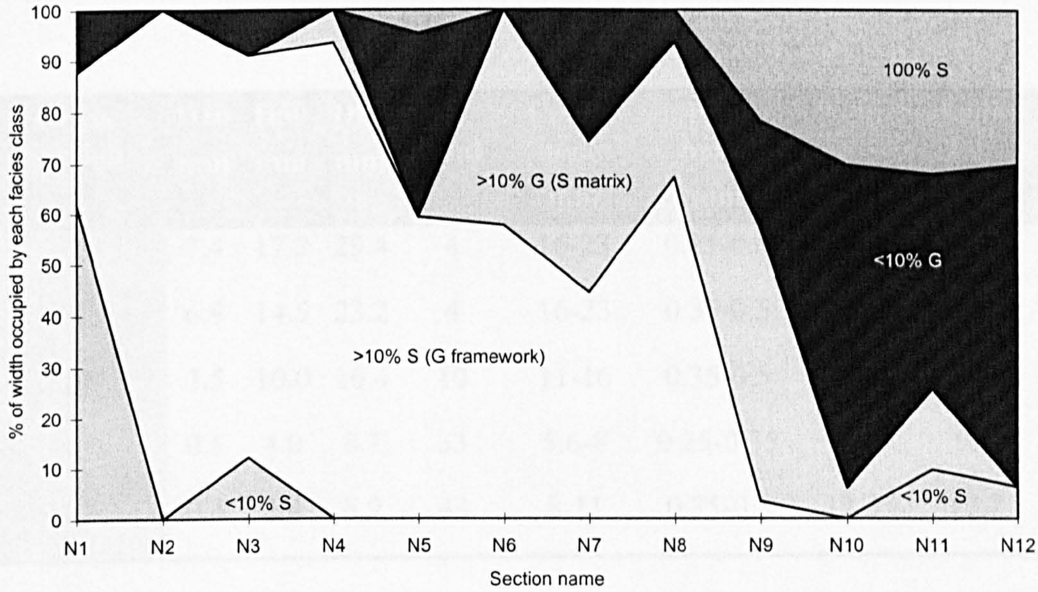


Figure 4.8: Proportion of the channel width of Vedder Canal occupied by different sedimentary facies. The GST occurs between N8 and N9.

Table 4.6: Details of the GSD samples collected from Allt Dubhaig in 1997. All samples were collected using a width-integrated technique. Coarse and fine mode refer to the half-phi sizes in which most gravel or sand sediments are found respectively.

Cross Section	D16 mm	D50 mm	D84 mm	% sand	Coarse mode mm	Fine mode mm	% in coarse mode	% in fine mode	B*
1	7.4	17.2	29.4	4	16-23	0.35-0.5	22.6	0.7	0.2
5	6.9	14.5	23.2	4	16-23	0.35-0.5	26.1	0.7	0.1
10	3.5	10.0	16.4	10	11-16	0.35-0.5	26.7	1.8	0.3
14	0.5	4.0	8.7	33	5.6-8	0.25-0.35	16.2	9.1	2.5
17	0.4	2.4	8.9	48	8-11	0.35-0.5	12.3	17.7	3.1

In Allt Dubhaig the character of the GSDs shows that the bed is gravel-dominated in the upstream samples, becoming bimodal further downstream and then sand dominated. The details of the samples collected in 1997 and 1999 are shown in Tables 4.6 and 4.7 respectively. Cumulative plots of the GSDs of the individual samples are shown in Figure 4.9 (1997) and Figure 4.10 (1999)

Table 4.7: Details of the GSD samples collected from Allt Dubhaig in 1999. The unshaded samples were collected using a width-integrated technique. Patch samples and the positions across the channel from which they were sampled (in m) are shaded. RB and LB are right bank and left bank respectively. Coarse and fine mode refer to the half-phi sizes in which most gravel or sand sediments are found respectively.

Cross Section	D16 mm	D50 mm	D84 mm	% sand	Coarse mode mm	Fine mode mm	% in coarse mode	% in fine mode	B*
1	5.8	14.2	24.1	7	16-23	0.25-0.35	23.7	1.3	0.3
5	5.1	13.5	22.5	8	16-23	0.25-0.35	23.7	1.8	0.5
10 LB-7	2.5	8.8	15.9	14	11-16	0.25-0.35	20.9	4.3	1.1
10 7-RB	0.3	0.4	0.6	99	None	0.25-0.35	N/A	48.4	N/A
10	0.3	4.3	13.9	43	11-16	0.25-0.35	13.9	19.1	4
14 LB-5	0.3	0.5	12.5	69	11-16	0.25-0.35	5.9	25	1.3
14 5-6	0.4	4.8	11.9	32	8-11	0.25-0.35	14.0	12.3	4.4
14 6-RB	0.3	0.3	0.6	90	2.8-4	0.25-0.35	2.8	68.1	0.1
14	0.3	0.4	6.7	72	8-11	0.25-0.35	4.9	43.8	0.6
17 LB-5	0.3	0.4	0.6	100	None	0.25-0.35	N/A	39.0	N/A
17 5-8	0.7	3.9	8.0	29	4-5.6	0.25-0.35	17.9	8.0	1.8
17	0.3	0.7	5.9	61	4-5.6	0.25-0.35	9.9	21.9	1.8
19	0.3	0.3	0.5	100	1.4-2	0.25-0.35	0.8	58.8	0.0
23	0.3	0.5	1.1	92	None	0.35-0.5	N/A	26.7	N/A
27	0.3	0.4	0.7	98	None	0.25-0.35	N/A	39.9	N/A

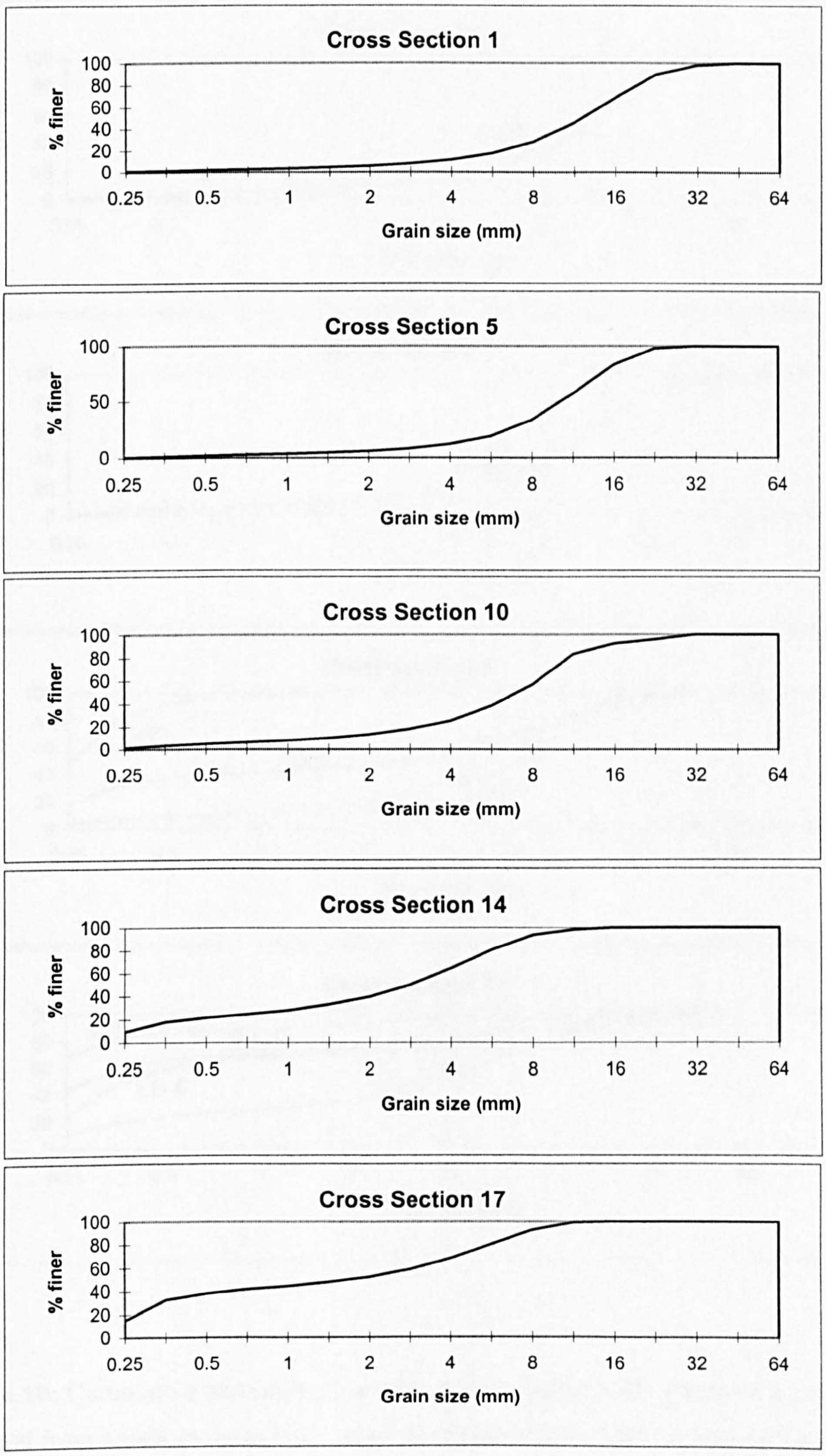


Figure 4.9: Cumulative bed GSD plots from Allt Dubhaig, 1997.

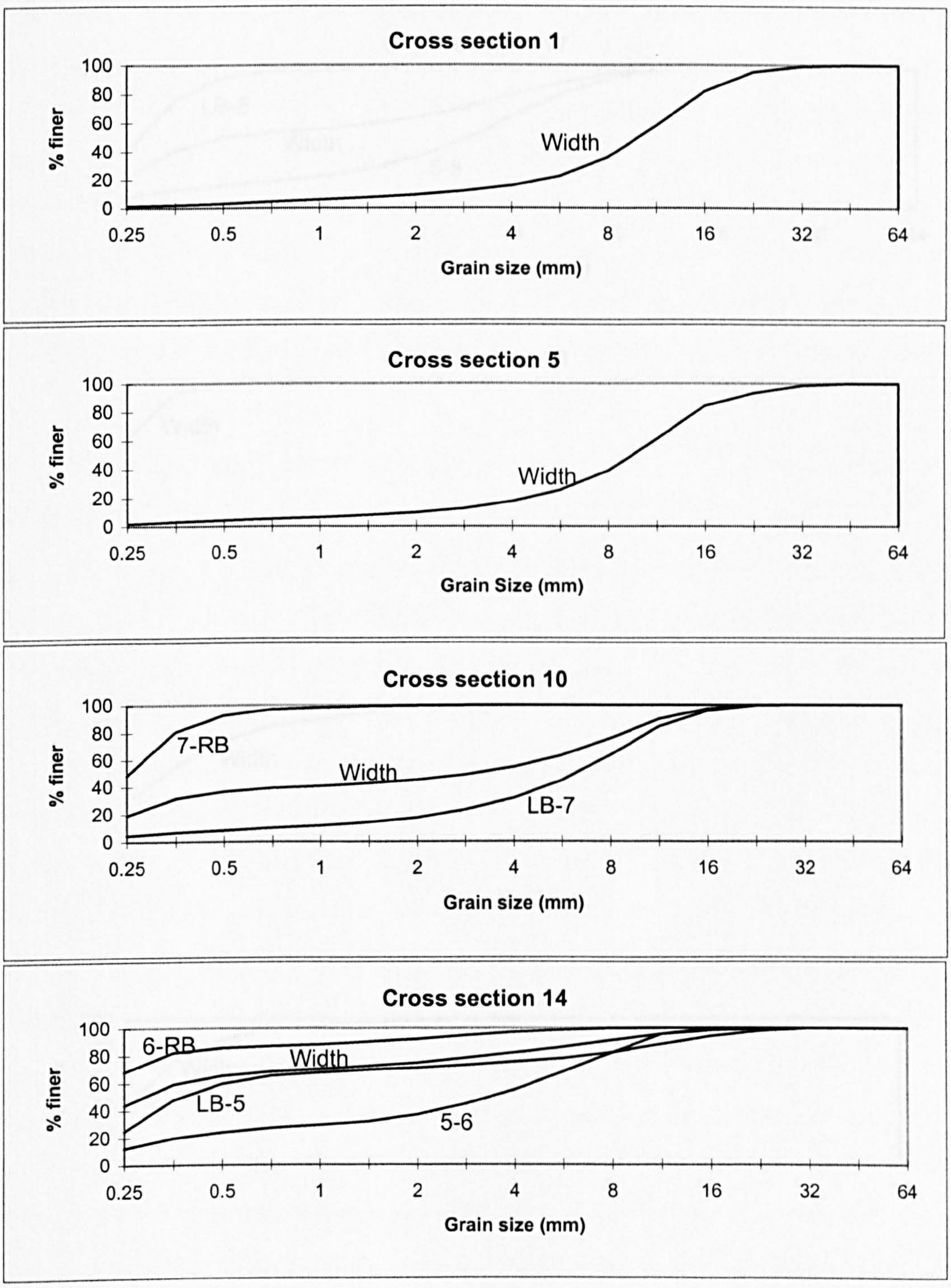


Fig 4.10: Cumulative bed GSD plots from Allt Dubhaig, 1999. Distance across the channel from which samples were taken are shown (in m). "Width" denotes a width-integrated sample. LB and RB are left bank and right bank respectively.

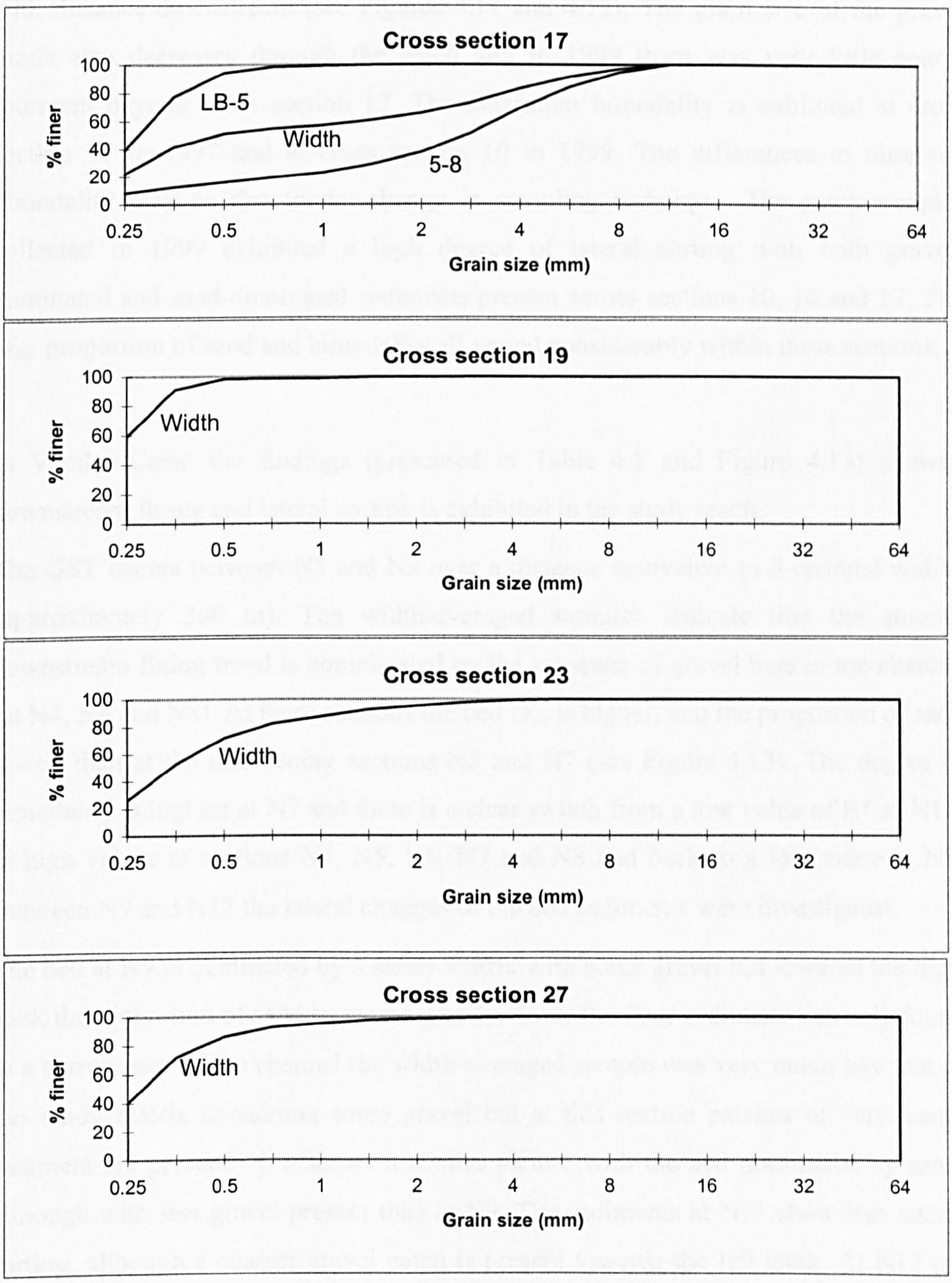


Fig 4.10: Cumulative bed GSD plots from Allt Dubhaig, 1999. Distance across the channel from which samples were taken are shown (in m). "Width" denotes a width-integrated sample. LB and RB are left bank and right bank respectively.

In the width-averaged samples the proportion of sand increases and bed D_{50} decreases with distance downstream (see Figures 4.11 and 4.12). The grain size of the gravel mode also decreases through the reach and in 1999 there was very little coarse sediment beyond cross section 17. The maximum bimodality is exhibited at cross section 17 in 1997 and at cross section 10 in 1999. The differences in observed bimodality may be due to the change in sampling technique. The patch samples collected in 1999 exhibited a high degree of lateral sorting with both gravel-dominated and sand-dominated sediments present across sections 10, 14 and 17. The D_{50}, proportion of sand and bimodality all varied considerably within these sections.

In Vedder Canal the findings (presented in Table 4.8 and Figure 4.13) showed downstream fining and lateral sorting is exhibited in the study reach.

The GST occurs between N8 and N9 over a distance equivalent to 3 channel widths (approximately 300 m). The width-averaged samples indicate that the smooth downstream fining trend is complicated by the presence of gravel bars in the channel (at N4, N6 and N8). At these sections the bed D_{50} is higher, and the proportion of sand lower, than at the intervening sections N5 and N7 (see Figure 4.13). The degree of bimodality is highest at N7 and there is a clear switch from a low value of B* at N1.5 to high values at sections N4, N5, N6, N7 and N8 and back to a low value at N9. Between N9 and N12 the lateral changes of the bed sediments were investigated.

The bed at N9 is dominated by a sandy matrix with some gravel but towards the right bank the proportion of sand increases greatly. Since the finer sediment was only found in a narrow part of the channel the width-averaged sample was very much like that of the sandy matrix containing some gravel but at this section patches of very sandy sediment are present. N10 shows a similar picture with the bed dominated by sand, although with less gravel present than at N9. The sediments at N11 show less lateral sorting, although a coarser gravel patch is present towards the left bank. At N12 the bed is dominated over the majority of the width by the sandy sediment that is present at N9 and N10, although the proportion of sand varied at different points across the channel.

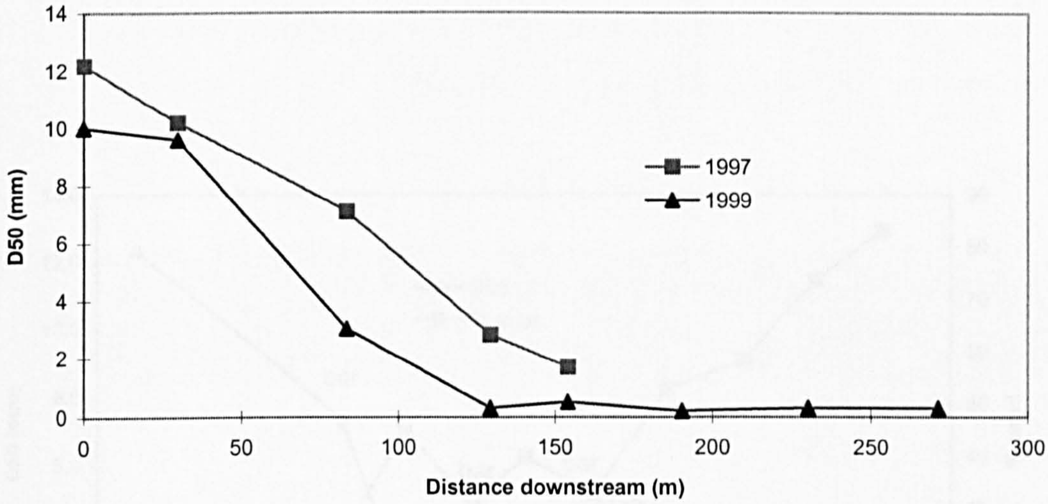


Figure 4.11: Width-averaged bed surface D_{50} from Allt Dubhaig in 1997 and 1999.

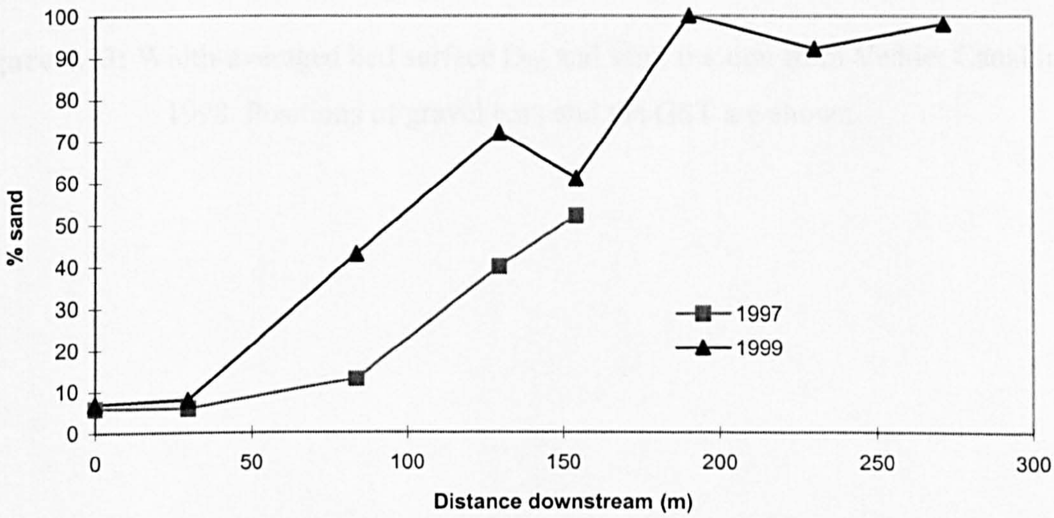


Figure 4.12: Width-averaged bed surface sand fraction from Allt Dubhaig in 1997 and 1999.

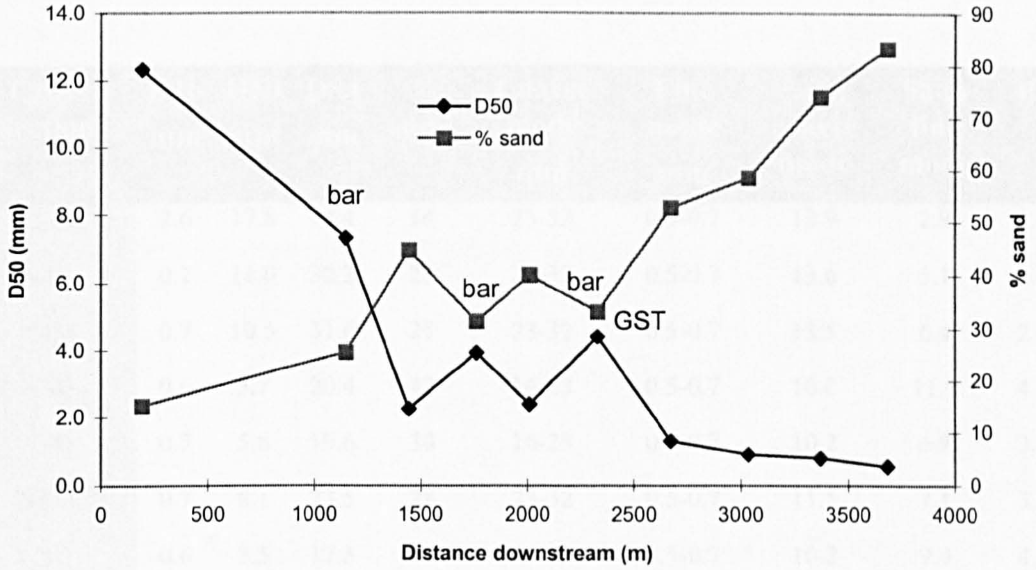


Figure 4.13: Width-averaged bed surface D_{50} and sand fraction from Vedder Canal in 1998. Positions of gravel bars and the GST are shown.

Table 4.8: Details of the GSD samples collected from Vedder Canal in 1998. "D" indicates that samples were collected using a width-integrated dredge technique. "M" indicates that the samples were collected at barhead locations using the technique of Martin (1992). Patch samples (collected using the dredge) and the positions across the channel from which they were sampled (in m) are shaded. RB and LB are right bank and left bank respectively.

Cross section	D16 mm	D50 mm	D84 mm	% sand	Coarse mode mm	Fine mode mm	% in coarse mode	% in fine mode	B*
N1.5D	2.6	17.8	44.4	14	23-32	0.5-0.7	13.9	2.9	1.1
N4D	0.7	11.0	30.2	23	23-32	0.5-0.7	13.6	5.1	2.1
N4M	0.7	10.5	31.6	25	23-32	0.5-0.7	13.5	6.4	2.6
N5D	0.6	3.7	20.4	42	16-23	0.5-0.7	10.0	11.1	4.5
N6D	0.7	5.8	19.6	30	16-23	0.5-0.7	10.2	6.9	3.4
N6M	0.7	6.1	23.5	28	23-32	0.5-0.7	11.5	7.1	3.4
N7D	0.6	3.5	17.3	49	16-23	0.5-0.7	10.2	9.4	4.6
N8D	0.6	6.5	21.2	32	16-23	0.5-0.7	13.2	7.1	2.7
N8M	0.5	5.9	18.8	30	16-23	0.5-0.7	12.3	7.3	3.0
N9 LB-65	0.6	2.0	7.1	50	2-2.8	0.5-0.7	11.8	10.2	1.7
N9 70-RB	0.5	0.9	2.3	81	None	0.5-0.7	N/A	22.9	N/A
N9D	0.6	1.9	6.6	52	1.4-2	0.5-0.7	11.7	11.1	1.4
N10 LB-18	0.7	2.8	10.6	41	2.8-4	0.5-0.7	10.6	8.5	2
N10 23-43	0.4	0.6	1.9	85	16-23	0.35-0.5	1.3	27.1	0.3
N10 48-RB	0.5	2.1	9.9	49	8-11	0.5-0.7	9.2	14.1	2.6
N10D	0.4	1.3	8.0	58	2.8-4	0.5-0.7	7.4	15.4	1.2
N11 LB-17	0.8	5.9	18.4	36	16-23	0.7-1	14.3	11.0	3.5
N11 23-43	0.4	1.1	3.1	75	1-1.4	0.5-0.7	14.1	14.9	0.9
N11 56-RB	0.5	1.1	2.2	81	1.4-2	0.5-0.7	16.9	14.8	1.3
N11D	0.5	1.1	3.0	74	1-1.4	0.5-0.7	14.6	14.3	1.0
N12 LB-64	0.4	0.8	1.8	87	None	0.5-0.7	N/A	22.4	N/A
N12 71-RB	0.5	1.1	5.5	67	5.6-8	0.5-0.7	7.2	17.4	1.4
N12D	0.4	0.8	2.3	82	5.6-8	0.5-0.7	3.1	21.2	0.5

A comparison between the samples collected using the dredge technique and those collected using the barhead sampling method of Martin (1992) shows similar results. At all three positions where both techniques were employed (N4, N6 and N8), Table 4.8 shows that the results are almost identical. This finding implies that for gravel-dominated sediments, such as those found at N4, N6 and N8, a barhead sample may characterise the bed sedimentology sufficiently. As noted above, however, these barhead samples are coarser than the samples collected between bars, and in gravel-sand mixtures there is a high degree of lateral sorting. In these situations facies samples may be more appropriate to characterise the bed sedimentology.

4.4 Summary

The results of an investigation into the form of two contemporary GSTs indicates both similarities and differences. In Allt Dubhaig the GST is associated with an order of magnitude break of slope and occurs in a meandering channel. In Vedder Canal there is no break of slope in the region of the GST. Through the transition zone the channel cross sections are dominantly rectangular, although there is evidence of an attenuated alternating bar morphology.

Bed surface sedimentology surveys indicate that there is a high degree of lateral sorting present in the transition zone of both rivers. In the Dubhaig this is controlled by channel morphology with gravel present in the deepest parts of the channel but in the Vedder the sorting is more complex and therefore more difficult to characterise.

Bed surface samples support the hypothesis of lateral sorting of different sedimentary facies units in the united gravel-sand zone. Samples from different patches indicate that at the same distance downstream both gravel-dominated and sand-dominated sediments can be present. Width-integrated samples collected from both rivers show the expected trend of downstream fining and associated increase in the proportion of bed surface sand. The strength of fining in Allt Dubhaig, however, is not quite as rapid as that indicated by previous work on this river. The apparent abruptness of the GST in both rivers is due to the change in visual bed surface facies rather than the rapid change downstream of the width averaged bulk GSD. The degree of bimodality increases in the transition zone, and the grain size and magnitude of the coarse mode decreases with distance downstream in both rivers. The relevance of these findings to the numerical modelling and the study of gravel-sand sediment mixtures and GSTs is discussed in Chapter 9.

Chapter 5. Channel change and GST evolution

Field evidence for morphological and sedimentological changes occurring over time in the GSTs of Allt Dubhaig and Vedder Canal is presented in this chapter. Data gathered from these streams for the current study is compared with that gathered at these sites by researchers previously. In addition, new methods to elucidate the changes occurring in the GST reaches over time were undertaken.

The form of the GST zones of the Dubhaig and the Vedder was investigated in detail for the current research and considerable information was available regarding the morphological and sedimentological characteristics of these reaches in the past. In Allt Dubhaig, Sambrook Smith (1994) surveyed channel cross sections and long profiles in 1992. He also collected bed sediment samples for GSD analysis from the last gravel bar, through the united gravel-sand reach, into the sand dominated reach. As noted in the previous chapter, however, these samples were not collected on a width integrated or facies-specific basis. In Vedder Canal, channel long profiles and cross section 1/49 (at the head of the Canal) were surveyed by engineers of the Canadian Water Management Branch in 1990. Martin (1992) also collected four bed sediment samples at gravel barheads in the Canal for GSD analysis in 1991.

The aim of the comparison, and new methods of investigation, was to understand the rates of aggradation, progradation of the gravel front, and changes in bed morphology and sedimentology operating in distal gravel and united gravel-sand reaches of the two rivers. In addition, bank erosion rates are analysed to assess the importance of these as the source of the fine sediment required to initiate a GST.

The techniques used to gather the data for the current study, and previous research, are outlined initially, where these differ from that discussed in the previous chapter. This is followed by a description of the changes indicated by the various sources of evidence. In all cases techniques and data from Allt Dubhaig are presented first, followed by Vedder River. At the end of the chapter a summary draws together, interprets and compares the available data from the two rivers. The degree to which rates of change experienced in the distal gravel and united gravel-sand reach can be ascertained from the information available is also discussed.

5.1 Methods of measurement

The techniques employed by the current research to collect the channel form, bed surface sedimentology and bed GSD data were outlined in Chapter 4. The methods used to gather data concerning fine gravel movement in the GST zone of Allt Dubhaig and coring in the GST reach of this river are explained in detail here.

5.1.1 Channel form

Using the information available a number of aspects of morphological change in the GST region of the Dubhaig and the Vedder were investigated. The evolution of channel form, including the extent of aggradation or degradation, was highlighted by channel long profile surveys.

The degree to which lateral migration of the channel occurs in the GST reach of the Dubhaig was also explored. This was achieved by comparing the 27 cross section surveys undertaken for the present research (the findings of which were outlined in the previous chapter) with those mapped by Sambrook Smith in 1992 (and reported in Sambrook Smith, 1994). The two surveys gave an indication of the importance of lateral fine sediment input from bank collapse. In the Dubhaig, samples were also collected to investigate the GSD of the bank sediments, for comparison with the fine mode in the bed sediments, to further constrain the possibility of bank sediments as the source of the increasing proportion of fine sediment on the channel bed. Bank sediment samples were collected from actively eroding banks upstream of, and in the region of the GST. These were sieved at half phi intervals to 0.25 mm.

5.1.2 Bed surface sedimentology

An investigation into the evolution of the GST reach of Allt Dubhaig was undertaken to assess the rate of progradation of the gravel front and the dynamics of the discrete grain size patches in the united gravel-sand reach. Change over time was assessed by comparing maps of the bed surface sedimentology, collected in 1992 by Sambrook Smith and in 1997 for the current study (see Section 4.3.3 and Figure 4.1). As noted in the previous chapter, a qualitative technique was used to classify the bed sediments to improve the understanding of the types and forms of lateral and longitudinal sorting occurring in gravel-sand mixtures. Sambrook Smith was present during the collection of data for the 1997 map to ensure that, as far as possible, similar sediments were classified in the same way for both sedimentological maps.

5.1.3 Bed GSDs

This aspect of the study was carried out to understand the recent sediment sorting processes occurring in the GST reach of Allt Dubhaig and Vedder River in a quantitative manner. In the Dubhaig, grain size samples were collected in 1992 by Sambrook Smith (and presented in Sambrook Smith, 1994) and in 1997 and 1999 for the current study. Similar techniques were used to collect the samples so that the spatial and temporal changes at these sites could be considered. Details of the method used to collect the sediment samples in each year were presented in Chapter 4. A list of the positions of the samples collected was also presented in Table 4.3.

In Vedder Canal three samples were collected at barheads in the distal gravel reach using the same method as Martin (1992) to allow comparison with these samples. This method (documented in Church *et al*, 1987) was outlined in the previous chapter. These samples were collected at N4, N6 and N8. As noted in Chapter 4 the samples were excluded from the analysis in Martin (1992) and Martin and Church (1995) because the rate of fining increased in the Canal when compared to the trend of the data for the gravel reach. Martin also collected a sample at N2 but, as noted in Chapter 4, when the fieldwork for the current research was carried out no gravel bar was present in this region. This site could not, therefore, be resampled using the barhead technique of Church *et al* (1987).

5.1.4 Fine gravel tracer pebble experiment

To understand the degree to which selective transport is operating in, and rate of progradation of, the distal gravel reach, a fine gravel tracer experiment was undertaken on Allt Dubhaig. The influence that the increase in areal extent of surficial sand in the united gravel-sand reach had on the gravel sediment sorting processes (suggested by Kuhnle, 1993a,b; Ferguson *et al*, 1989) was also investigated using this technique. As reported in Section 2.3.2, the decreased importance of hiding and the occurrence of gravel overpassing on a sand-dominated bed were thought to become important in the GST zone. It was noted, by Lisle (1995) and Paola and Seal (1995), that where sand and gravel sediments are present together on a river bed they tend to organise themselves into patches of the two distinct GSDs (see Section 2.3.3) and this process also required further investigation.

To fulfil these requirements two separate tracer experiments were undertaken. The first involved seeding tracers of different sizes on the last gravel bar upstream of the united gravel-sand reach. A second experiment was carried out in the united gravel-sand reach, downstream of the last gravel bar tracers. Tracers of different sizes were seeded on both gravel and sand patches in the GST zone and an analysis of the bed sediment type on which they were found when remapped allowed an assessment of the importance of bed sedimentology for fine gravel mobility in this zone.

To carry out the necessary experiments fine gravel sediments were required to be representative of the bed GSD in the field (discussed in Chapter 4). If the tracers were too coarse the results could have been unrealistic as the transport processes occurring would be unlike those experienced in the natural system. Where the tracers were seeded on the last gravel bar the width-averaged D_{50} was 14.2 mm (in 1999). In the united gravel-sand reach this was 4.3 mm (in 1999), with a D_{50} of 8.8 mm on the gravel patch and 0.4 mm on the sand. It was decided that artificial pebbles would be easier to use as inserting the magnets, necessary to allow the relocation of buried tracers, into natural fine gravels often resulted in splitting the grains (Ferguson, 1997, pers comm). The mildly elliptical tracers (A-axis 2 mm larger than B-axis) were made of 2-Ton epoxy resin shaped in an aluminium mould. Disc-shaped neodymium boron iron magnets with a diameter of 6 mm and width of 3 mm were inserted into the tracers during manufacture. To achieve the correct density (assumed to be 2650 kg m^{-3}

for the sediments found in Allt Dubhaig) fishing tackle weights were used to increase the tracer mass.

A number of different populations of tracers of different sizes were necessary to fulfil the aims of the experiment. Forty tracers of each size were seeded at each starting point. This number was chosen to as it is the same as the figure used in a long-term gravel mobility study in the Dubhaig. Analyses during this study have shown that forty tracers of each size allow statistical inferences to be drawn from the dataset (Ferguson, 1998, pers comm). Upstream, on the last gravel bar, 3 sizes of tracer were used with diameters of 10, 14 and 20 mm. A total of 120 tracers were therefore seeded here. These were inserted at cross section 2 (see Fig 4.1). Downstream, in the united gravel-sand reach, 2 tracer sizes were used (10 and 14 mm). A population consisting of each size was inserted on both the gravel and the sand bed. A total of 160 tracers were therefore seeded here. These were inserted at cross section 9 (see Fig 4.1). Tracers of different size and seeding point were painted different colours to assist in assessing their provenance during remapping.

Both sets of tracers were seeded in March 1998. The upstream tracers were resurveyed in October 1998 and April 1999. The downstream tracers were resurveyed in August 1999. The tracers were not remapped on the same occasion as the flow conditions in October 1998 and April 1999 prohibited access to the downstream tracers where the bed slope was lower than upstream and therefore the flow was deeper. The tracers were located using a Schonstedt magnetic detector and remapped using tapes, triangulating their position from cross section end points of known co-ordinates. It is estimated that errors in the surveyed distance travelled associated with this technique are of the order of +/- 0.05 m. No discharge data were available for the period over which the tracers were in the stream. However, since this aspect of the current research was undertaken to elucidate the types of sorting processes occurring, the rates at which these operated was of secondary importance.

5.1.5 Probing in the GST reach

Probing investigations were carried out to determine the recent depositional history of the GST reach of Allt Dubhaig. A description of the channel subsurface sedimentology was used to gain a better understanding of the dynamics of the gravel front and the united gravel-sand reach. The investigation also elucidated the influence that the construction of the diversion dam (discussed in Chapter 3) had on the depositional environment in the last 70 years. Coring in the floodplain by Ferguson and Hoey (unpublished study) suggested that the amount of lateral migration the lower reaches of the channel, where the GST occurs, had undergone recently was less than 15 m. It can therefore be assumed that the subsurface sediments were deposited by a channel that was in a similar location and form to that present today.

The probing was undertaken using a screw auger that was 105 cm from the tip to the handle. Several probes were taken at each of the 27 cross sections through the transition zone to gain a representative picture of subsurface sediment for a given distance downstream. The positions of these sections can be seen in Figure 4.1. Further probing was carried out using an extendable post-hole auger the length of which could be increased in 1 m intervals using extension rods.

The results of the probing investigations were purely qualitative, noting if the sediments felt like gravel, sand or a mixture of these sediments, and in some cases whether the gravel sediments were fine or coarse. This was judged by ease and degree of smoothness experienced when turning the auger. The harder it was to screw, and the rougher the rotation, the coarser the subsurface sediments. A small number of sediment samples retrieved during the probing were used to qualitatively validate and calibrate the technique.

5.2 Results of analysis

5.2.1 Channel form

As reported in Chapter 4, the water surface profile of Allt Dubhaig exhibited a break of slope in the region of the last gravel bar. The gradients upstream and downstream of the break of slope were 0.0020 and 0.0002 respectively. Sambrook Smith and Ferguson (1995) reported that in 1992 the water surface slope decreased from 0.0022 to 0.0002 at the last gravel bar. The different slope upstream of the last gravel bar may be the result of the surveys being carried out when the discharge was not identical, but fall within the survey error associated with the methodology (+/- 5 cm for each point).

The actual position of the break of slope, however, did not change between the surveys.

The overall form of the Allt Dubhaig cross sections did not change although there are minor differences on all sections. Figure 5.1 shows the cross sections upstream of the united gravel-sand reach which exhibit the maximum and minimum degree of change between 1992 and 1997. It is from these sections upstream of the united gravel-sand reach where eroded sediments are most likely derived if these cause GST initiation. A selection of representative overlain Dubhaig section surveys from 1992 and 1997 are shown in Figure 5.2.

Details regarding the grain size of the bank samples collected from Allt Dubhaig upstream of, and in the region of the GST are shown in Table 5.1.

Table 5.1: Positions and details of samples collected from actively eroding banks in the region of the GST in Allt Dubhaig. Upstream is represented in the table by U/S. RB, LB and XS are right bank, left bank and cross section respectively.

Position	% < 0.25 mm	% < 0.35 mm
LB c.300 m U/S of LGB	89.8	97.7
RB at XS1	94.7	100
RB at XS9	100	100

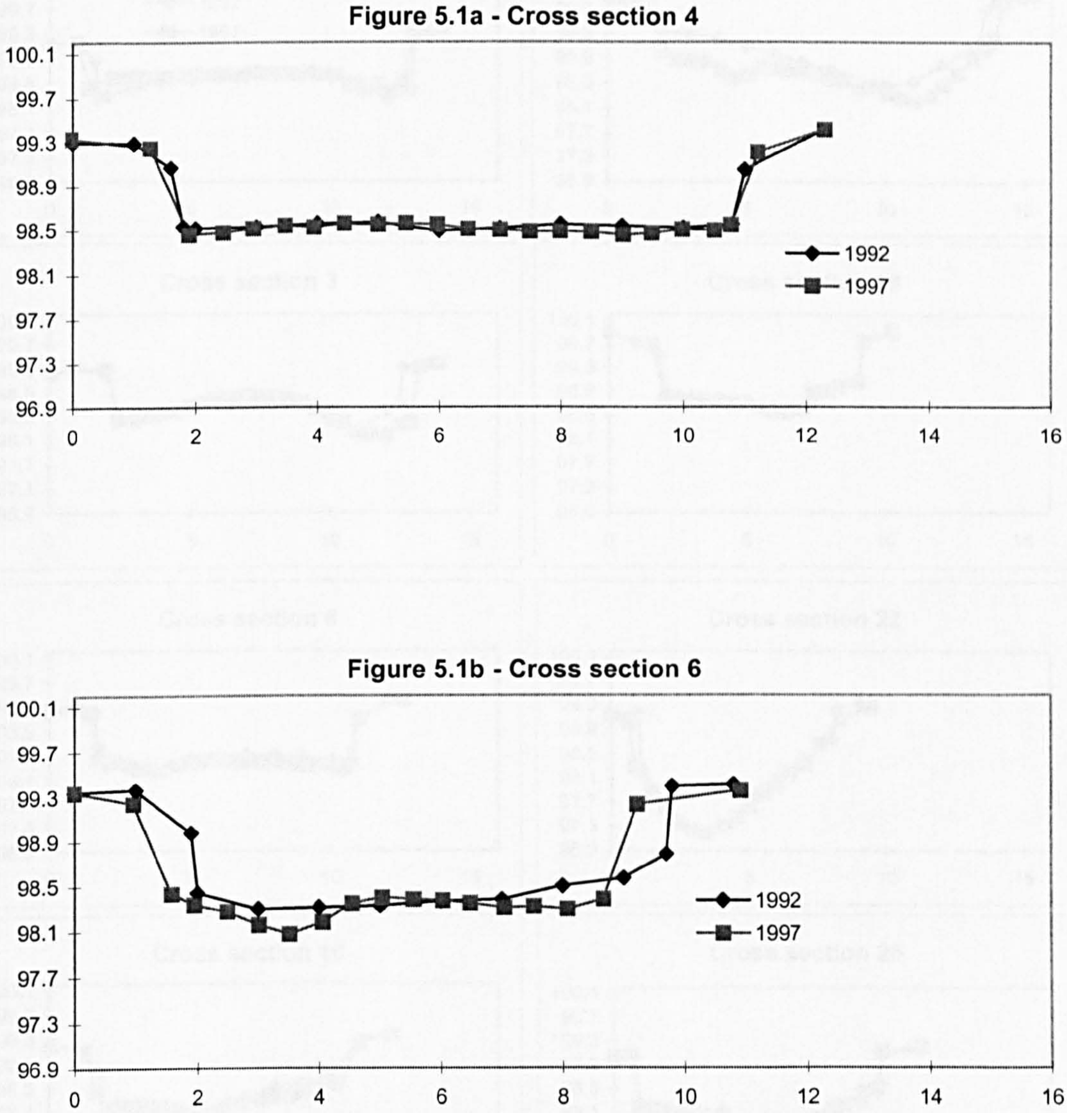


Figure 5.1: Allt Dubhaig cross sections upstream of the GST showing (a) minimum and (b) maximum change between 1992 and 1997. All axes distances are in metres.

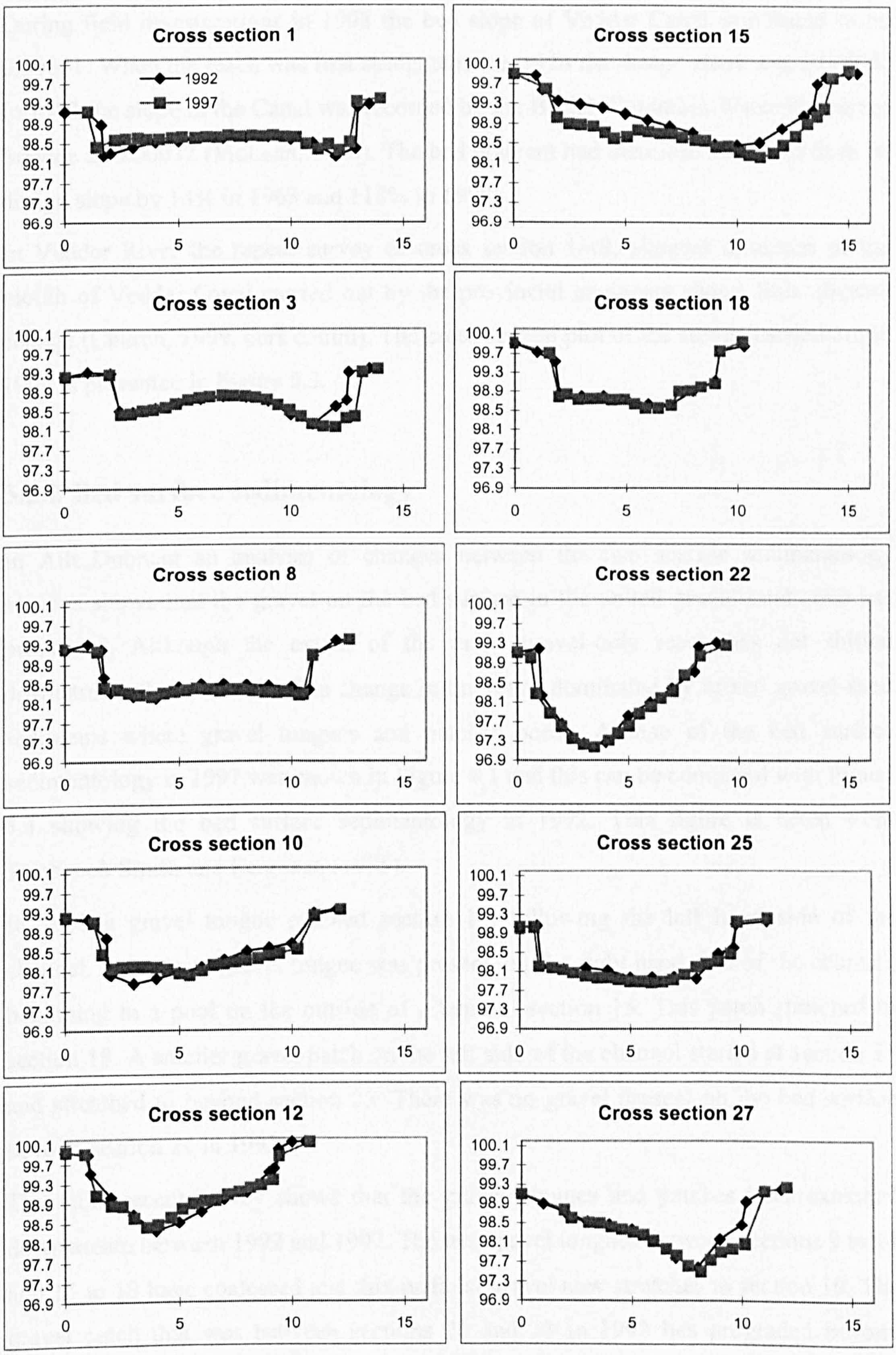


Figure 5.2: Allt Dubhaig cross sections mapped in 1992 and 1997, overlain for comparison (see Figure 4.1 for locations). All axes distances are in metres.

During field investigations in 1998 the bed slope of Vedder Canal was found to be 0.00061. When the reach was first straightened in 1924 the design slope was 0.00028. In 1963 the slope in the Canal was recorded by the British Columbia Water Resources Service as 0.00032 (McLean, 1980). The bed gradient had therefore increased from its design slope by 14% in 1963 and 118% in 1998.

In Vedder River the repeat survey of cross section 1/49, situated upstream of the mouth of Vedder Canal carried out by the provincial engineers shows little channel change (Church, 1999, pers comm). The cross section plot of the survey carried out in 1998 is presented in Figure 5.3.

5.2.2 Bed surface sedimentology

In Allt Dubhaig an analysis of changes between the two surface sedimentology surveys shows that the gravel on the bed surface in the united gravel-sand zone has prograded. Although the extent of the distal gravel-only reach has not shifted downstream there is noticeable change in the reach dominated by mixed gravel-sand sediments where gravel tongues and patches occur. A map of the bed surface sedimentology in 1997 was shown in Figure 4.1 and this can be compared with Figure 5.4 showing the bed surface sedimentology in 1992. This figure is taken from Sambrook Smith and Ferguson (1995).

In 1992 a gravel tongue reached section 14, following the left hand side of the channel. A separate gravel tongue was present, on the right hand side of the channel, beginning in a pool on the outside of a bend at section 15. This patch stretched to section 18. A smaller gravel patch on the left side of the channel started at section 21 and stretched to beyond section 23. There was no gravel present on the bed surface beyond section 24 in 1992.

The more recent survey shows that the gravel tongues and patches have extended downstream between 1992 and 1997. The two gravel tongues between sections 9 to 14 and 15 to 18 have coalesced and this surficial gravel now stretches to section 19. The gravel patch that was between sections 21 and 23 in 1992 has prograded beyond section 24 by 1997.

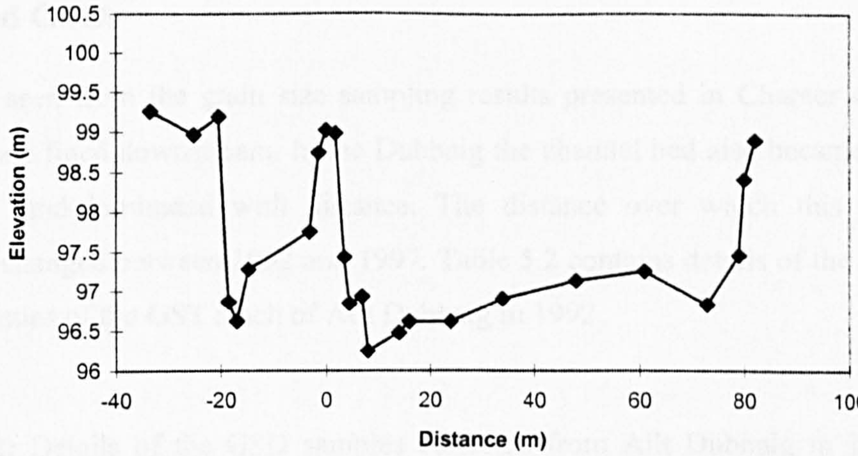


Figure 5.3: Cross section 1/49 of at the head of Vedder Canal, mapped in 1998.

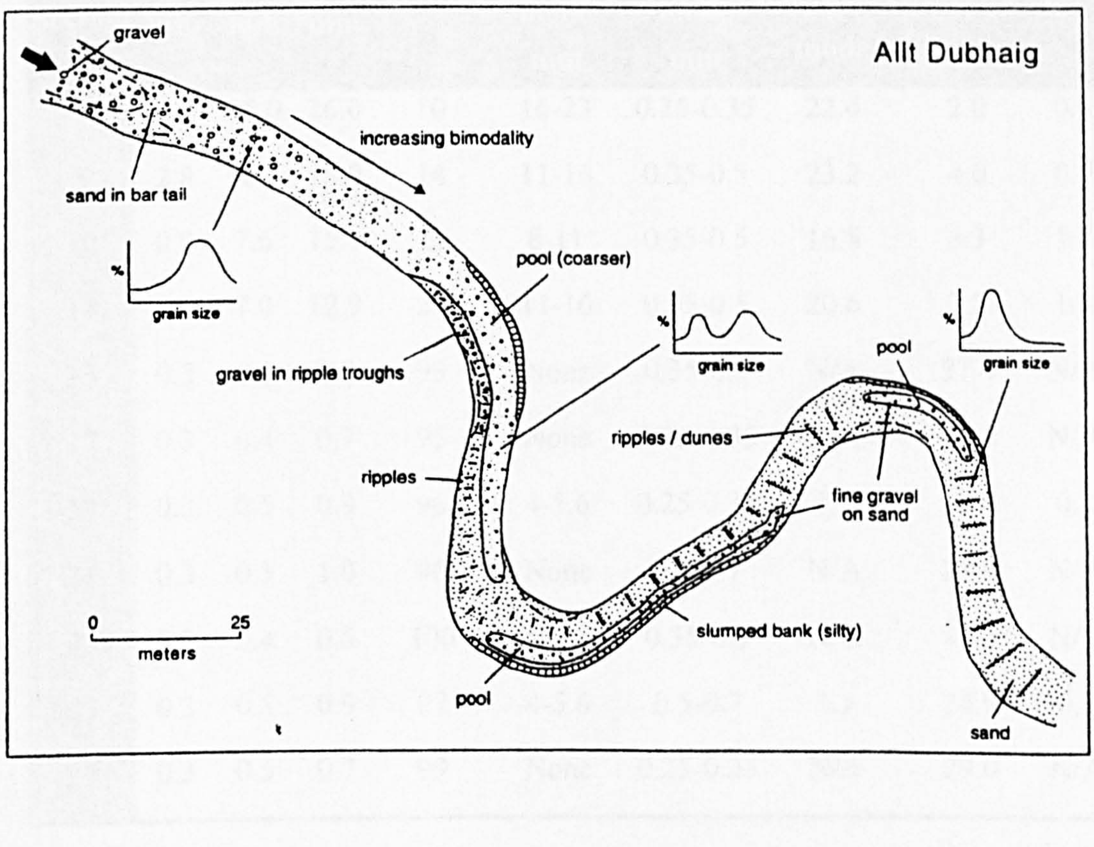


Figure 5.4: Main features of the transition zone of Allt Dubhaig (from Sambrook Smith and Ferguson, 1995), for comparison with Figure 4.1. Grain size curves are schematic only, to show downstream change in bed texture and GSD.

5.2.3 Bed GSDs

It can be seen from the grain size sampling results presented in Chapter 4 that the gravel mode fined downstream. In the Dubhaig the channel bed also became bimodal and then sand-dominated with distance. The distance over which this occurred, however, changed between 1992 and 1997. Table 5.2 contains details of the bed GSD characteristics of the GST reach of Allt Dubhaig in 1992.

Table 5.2: Details of the GSD samples collected from Allt Dubhaig in 1992. (for further details see Sambrook Smith, 1994). B* is the bimodality index defined in Equation 4.1.

Cross Section	D16 mm	D50 mm	D84 mm	% sand	Coarse mode mm	Fine mode mm	% in coarse mode	% in fine mode	B*
1	4.6	15.0	26.0	10	16-23	0.25-0.35	22.4	2.0	0.5
5	2.8	10.1	18.0	14	11-16	0.35-0.5	23.2	4.0	0.9
10	0.8	7.6	15.9	23	8-11	0.35-0.5	16.8	6.3	1.7
14	0.8	7.0	12.9	25	11-16	0.35-0.5	20.6	7.2	1.8
15	0.3	0.5	1.0	93	None	0.35-0.5	N/A	27.4	N/A
17	0.3	0.4	0.7	95	None	0.25-0.35	N/A	25.5	N/A
19	0.3	0.5	0.9	96	4-5.6	0.25-0.35	1.3	30.0	0.2
21	0.3	0.5	1.0	96	None	0.5-0.7	N/A	28.4	N/A
23	0.3	0.4	0.5	100	None	0.35-0.5	N/A	47.6	N/A
25	0.3	0.5	0.9	97	4-5.6	0.5-0.7	1.1	26.9	0.1
27	0.3	0.5	0.7	99	None	0.25-0.35	N/A	29.0	N/A

The data included in Table 5.2 shows that Allt Dubhaig exhibited a spatially rapid reduction in grain size and an increase in the proportion of sand present in the bed in 1992. The gravel mode also tended to fine in a downstream direction. The pattern of changing bed GSD characteristics with distance downstream is therefore similar to that present in this reach of the stream in 1997 and 1999. A comparison of the bed D_{50} in 1992 with that present in 1999 is shown in Figure 5.5. This plot indicates that the spatially rapid fining that occurs takes place over a shorter distance in 1992 than in 1999. In 1992 the switch from a gravel framework bed to a sand matrix bed took place over a distance of 7 m whereas in 1999 it takes approximately 100 m. The rate of change of the proportion of bed sand was similarly increased between the two surveys (see Table 4.7 for comparison).

In Vedder Canal the three barhead samples collected for the present research can be compared to those collected by Martin in 1991 (and presented in Martin, 1992; Martin and Church, 1995). Details of the samples collected in 1991 are shown in Table 5.3.

Table 5.3: Details of barhead subsurface bulk sediment samples collected from Vedder Canal in 1991 (Church, 1997, pers comm). Further details can be found in Martin (1992).

Cross Section	D50 (mm)	D84 (mm)	% sand
N2	7.6	14.2	32
N4	16.0	31.1	31
N6	3.9	7.2	73
N8	3.2	4.0	95

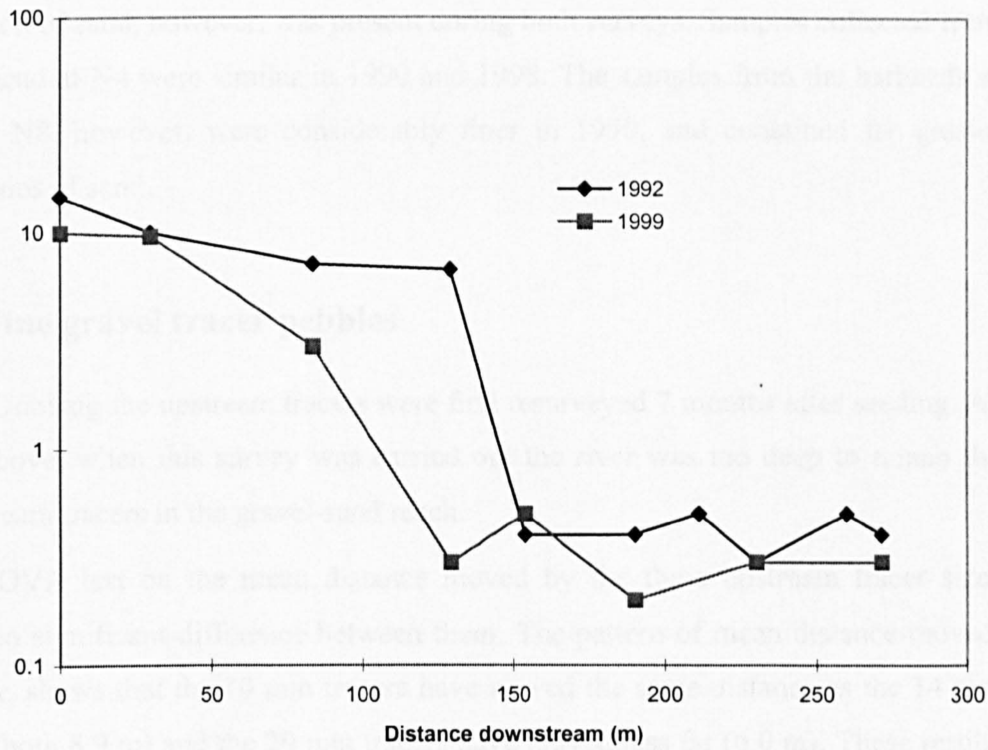


Figure 5.5: Width-averaged bed surface D_{50} from Allt Dubhaig in 1992 and 1999 (from cross section 1).

These results can be compared to the barhead samples, presented in Table 4.8 in the previous chapter, that were collected for the present research using the same method. As noted above, there was no bar present at N2 in 1998 from which to retrieve a sample. It can be seen that between 1990 and 1998 significant change had occurred through the reach. The overall pattern of downstream fining and an increasing proportion of sand, however, was present during both surveys. Samples collected from the barhead at N4 were similar in 1990 and 1998. The samples from the barheads at N6 and N8, however, were considerably finer in 1990, and contained far greater proportions of sand.

5.2.4 Fine gravel tracer pebbles

In Allt Dubhaig the upstream tracers were first resurveyed 7 months after seeding. As noted above, when this survey was carried out the river was too deep to remap the downstream tracers in the gravel-sand reach.

An ANOVA test on the mean distance moved by the three upstream tracer sizes shows no significant difference between them. The pattern of mean distance moved, however, shows that the 10 mm tracers have moved the same distance as the 14 mm tracers (both 8.9 m) and the 20 mm tracers have moved less far (6.0 m). These results indicate that selective transport may occur, although the processes had not had sufficient time to disperse the tracers in a statistically significant size selective manner. Details of later surveys of the upstream (13 months after seeding) and downstream (17 months after seeding) tracers are shown in Table 5.4.

The mean transport distances of the upstream tracers after 13 months show a similar pattern to that found after 7 months. The 20 mm tracers have, on average, moved less far than either the 14 mm or the 10 mm tracers. The maximum distance travelled by a 20 mm tracer is also lower. The 14 mm tracers have, however, moved further on average than the 10 mm tracers. This indicates that selective transport is not acting on these tracers. The proportion of the 20 mm tracers buried when remapped is half that of the 14 and 10 mm tracers.

Table 5.4: Recovery rates and general details of fine gravel tracer pebbles in Allt Dubhaig. U/S and D/S indicate the upstream and downstream tracers respectively. The bed sediments that the downstream tracers were seeded on is shown in brackets in the 'Tracer type' column. Max / Min refers to the maximum and minimum distance moved by each tracer type, respectively.

Tracer type	Recovery rate (%)	Mean dist moved (m)	Max / Min (m)	Buried (%)	Found on? Sand / Gravel (%)
U/S 10 mm	95	8.9	43 / 0	65	N/A
U/S 14 mm	95	11.7	52.4 / 0.7	63	N/A
U/S 20 mm	93	7.6	33.4 / 0.3	33	N/A
D/S 10 mm (sand)	90	5.8	29.5 / 0.3	87	92 / 8
D/S 14 mm (sand)	85	2.8	24.3 / 0	88	76 / 24
D/S 10 mm (gravel)	93	6.7	33.8 / 0	92	22 / 78
D/S 14 mm (gravel)	98	4.8	23.6 / 0	95	13 / 87

When compared to the upstream tracers, a larger proportion of the downstream tracers are buried. The evidence presented in Table 5.4 also indicates that the downstream tracers tend to remain on the bed type on which they were seeded. Selective transport by size is operating for both those seeded on sand and on gravel although those seeded on gravel had moved further on average.

Various statistical tests were carried out to investigate patterns of movement of tracers and burial of different sizes. The importance of the bed sedimentology upon which the downstream tracers were seeded and found was also explored. Tests were carried out on both the absolute and ranked distances travelled of the tracers. Details can be found in Tables 5.5, 5.6 and 5.7.

The results presented in Tables 5.5, 5.6 and 5.7 show that only a limited number of the tests carried out on the tracer data are statistically significant. Transport of the upstream tracers after 13 months (shown in Table 5.5) is not size selective. The ranked distance travelled, however, is related to whether or not the tracers are buried. Tracers that are found on the surface move further than those which are buried.

Table 5.5: Details of statistical analyses carried out on the upstream tracers. P-values significant at the 5% confidence level are shown in **bold**. +ve/-ve indicate whether the relationship between the response and predictors was positive/negative for significant regression tests (if applicable).

Test	Response	Predictors	p value	+ve/-ve
Multiple regression	Actual distance	Size	0.355	N/A
		Buried?	0.275	N/A
Multiple regression	Ranked distance	Size	0.180	N/A
		Buried?	**0.039**	-ve
ANOVA (oneway)	Actual distance	Size	0.304	N/A
ANOVA (oneway)	Ranked distance	Size	0.666	N/A
ANOVA (oneway)	Actual distance	Buried?	0.388	N/A
ANOVA (oneway)	Ranked distance	Buried?	0.079	N/A

Table 5.6: Details of statistical analyses carried out on the entire set of downstream tracers. P-values significant at the 5% confidence level are shown in **bold**. +ve/-ve indicate whether the relationship between the response and predictors was positive/negative for significant regression tests (if applicable).

Test	Response	Predictors	p value	+ve/-ve
Multiple regression	Actual distance	Size	0.065	N/A
		Buried?	0.236	N/A
Multiple regression	Ranked distance	Size.	**0.004**	-ve
		Buried?	0.592	N/A
Multiple regression	Ranked distance	Size	**0.004**	-ve
		Seeding	0.681	N/A
		Buried?	0.597	N/A
ANOVA (oneway)	Actual distance	Size	0.058	N/A
				N/A
ANOVA (oneway)	Ranked distance	Size	**0.004**	N/A
ANOVA (oneway)	Actual distance	Buried?	0.207	N/A
ANOVA (oneway)	Ranked distance	Buried?	0.696	N/A
ANOVA (oneway)	Actual distance	Seeding	0.283	N/A
ANOVA (oneway)	Ranked distance	Seeding	0.626	N/A
ANOVA (oneway)	Actual distance	Found on?	0.119	N/A
ANOVA (oneway)	Ranked distance	Found on?	0.788	N/A
ANOVA (oneway)	Found on?	Seeding	**0.000**	N/A
ANOVA (oneway)	Buried?	Size	0.577	N/A
ANOVA (oneway)	Buried?	Seeding	0.873	N/A

Table 5.7: Details of statistical analyses carried out on the downstream tracers. The grains were sorted according to their size and the type of bed surface sedimentology (gravel or sand) on which they were seeded. P-values significant at the 5% confidence level are shown in **bold**. +ve/-ve indicate whether the relationship between the response and predictors was positive/negative for significant regression tests (if applicable).

Test	Response	Predictors	p value	+ve/-ve
ANOVA (oneway)	Ranked distance (10 mm)	Seeding (10 mm)	0.472	N/A
ANOVA (oneway)	Ranked distance (14 mm)	Seeding (14 mm)	0.928	N/A
ANOVA (oneway)	Ranked distance (sand seeding)	Size (sand seeding)	**0.007**	N/A
ANOVA (oneway)	Ranked distance (gravel seeding)	Size (gravel seeding)	0.128	N/A

The downstream tracers are transported on a size selective basis, unlike the upstream tracers. Table 5.6 shows that, when considering all the downstream tracers, the ranked distance travelled decreases with increasing tracer size in a statistically significant manner, although the absolute distance travelled did not. These tracers also tend to remain on the bed type on which they are seeded. Table 5.7 shows that if the downstream tracers are sorted depending on the bed sediment type on which they are seeded, however, only the sand-seeded tracers are transported on a statistically significant size selective basis.

5.2.5 Probing in the GST reach

A simplified one-dimensional diagram of the probing information collected from Allt Dubhaig is presented in Figure 5.6. The data indicates that the first 2 sections are all gravel below the surface. Sections 3 to 8 have some sand below the surface, although this is only present in thin lenses at depth.

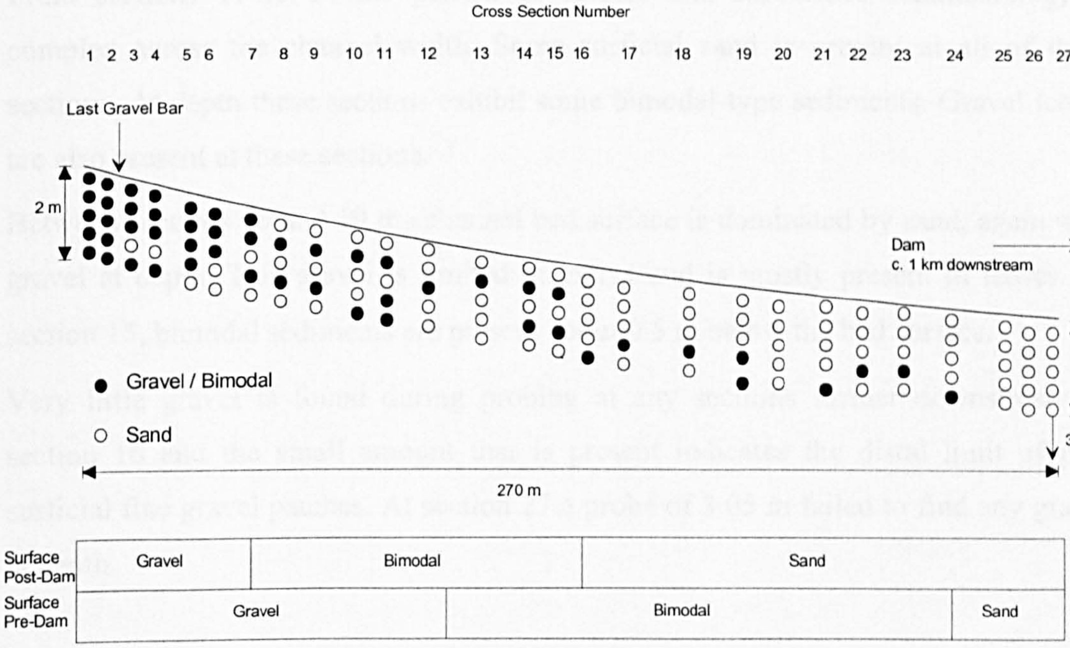


Figure 5.6: Simplified one-dimensional diagram of subsurface sedimentology from probing information gathered in the GST reach of Allt Dubhaig. Inferences are made regarding the bed surface sedimentology before and after dam construction.

Section 8 shows bimodality to some depth and sections 9 and 10 shows that the near-surface of the bed has become predominantly sandy today. This sand is approximately 0.4 m deep. The gravel tongues and patches present on the surface from section 8, and further downstream, are less than 5 cm thick and therefore easy to probe through.

From sections 11 to 14 the pattern of surface and subsurface sedimentology is complex across the channel width. Some surficial sand is present at all of these sections. At depth these sections exhibit some bimodal-type sediments. Gravel lenses are also present at these sections.

Between sections 15 and 19 the channel bed surface is dominated by sand, again with gravel at depth. This gravel is limited in extent and is mostly present in lenses. At section 15, bimodal sediments are present about 0.5 m below the bed surface.

Very little gravel is found during probing at any sections further downstream of section 16 and the small amount that is present indicates the distal limit of past surficial fine gravel patches. At section 27 a probe of 3.05 m failed to find any gravel at depth.

5.3 Summary and preliminary interpretation

The Allt Dubhaig water surface profiles surveyed in 1992 and 1997 are similar and suggest that aggradation in the distal gravel reach of this stream is progressing at a rate that was undetectable over this timescale. This may be due to discharge not being equal in the stream during the two surveys. The position of the break of slope, however, did not vary between 1992 and 1997 indicating that any distal gravel progradation is also occurring at a slow rate.

The cross section surveys on the Dubhaig show lateral change of only limited extent supporting the floodplain coring evidence of Hoey and Ferguson (unpublished study). The results of the grain size investigations of the Dubhaig bank samples indicate that these sediments are considerably finer than the fine mode of the bed sediments, which is typically between 0.35 and 0.7 mm. This fact, combined with the low rate of lateral channel migration, suggests that sediments supplied from bank erosion are unlikely to be the primary cause of GST initiation in this case.

Field investigations undertaken for the current research show both the bed and water surface slope of Vedder Canal is 0.00061. The channel has, therefore, steepened considerably since 1924 when the Canal was constructed with a bed slope of 0.00028. It is likely that there has been progradation of gravel from upstream in the Vedder River at the proximal end of the Canal and this has caused aggradation through the whole reach. There is no evidence of a 0.00028 bed slope towards the distal end of the Canal. It is extremely unlikely that degradation at the lower end of the Canal has been the cause of the increase in bed gradient. A backwater zone was created in the distal part of the Canal where the Vedder was joined by the Sumas River and as the stream joined the mainstem of the Fraser. The flow velocity would therefore be lower in this zone, impeding entrainment and erosion at high flows. It is known that some aggradation had occurred by 1963, when the Canal slope was recorded as 0.00032 (McLean, 1980).

The evidence available from the survey at cross section 1/49 indicates that little morphological change is occurring at the mouth of Vedder Canal. This, combined with stable channel banks in the Canal, suggests that, as with Allt Dubhaig, sediments derived from channel lateral migration and bank erosion are unlikely to be the source of fines necessary to generate a GST. There is no evidence available regarding the change over time of the stream morphology in Vedder Canal where the GST occurs.

The surface sedimentological maps of Allt Dubhaig collected in 1992 and 1997 indicate that the form of the sand and gravel patches in the united gravel-sand reach has changed. The findings provide evidence that the surficial gravel in the united gravel-sand reach is prograding and patches on the bed surface in the transition zone are merging. The main gravel patches have increased in length by between 11 and 15 m during the 5 year gap between the surveys. The gravel front has prograded less far, however. This may have been due to the fact that no gravel overpassing can occur in that part of the reach. Gravel grains present in the tails of the coarse patches in the united gravel-sand reach may be transported rapidly to the head of the next coarse patch downstream due to their increased exposure. For the gravel front to prograde, however, coarse sediment must be transported over the entire channel width, rather than just in the thalweg where the gravel patches are present in the united gravel-sand reach.

During all bed surface sediment sampling the expected trend of downstream fining and an increase in the proportion of sand occurs in both Allt Dubhaig and Vedder Canal. The strength of this fining with distance varies over time, however. In Allt Dubhaig in 1992 the spatially rapid fining between a gravel and a sand bed occurred over a much shorter distance than in 1997. The order of magnitude difference was likely to be a result of the alternative sampling strategies employed to collect the sediment samples. It is therefore clear that the sampling strategy used to collect sediment samples in gravel-sand rivers is crucial to characterise the stream sedimentology.

In Vedder Canal the barhead samples indicate that there has been considerable coarsening of the stream bed at N6 and N8 between 1991 and 1998, although there was no bed coarsening at N4. This evidence supports the change in bed surface profile over time by indicating a prograding gravel front.

The recovery rates of artificial fine gravel tracers from Allt Dubhaig are high. This is a result of the relatively low degree of activity in the reach in which they were seeded and careful, methodical searching during resurveys together with the fact that the grains had not been in the stream for a long period. The average distance moved by the upstream tracers did not increase by a large amount between the two surveys indicating that these tracers are reaching the limit of rapid transport as they approach the gravel front. Their mobility may also be impeded as they become buried and therefore transported less often. The average distance moved by the 10 mm upstream tracers is the same after 13 months in the stream as it was after 7. The upstream tracers have not undergone size selective transport. This may be due to a limited number of flows capable of entraining the grains. If the tracers had remained in the stream longer the results may have indicated that size selective transport was operating. The high recovery rates support this hypothesis, indicating that few of the grains are deeply buried. Unfortunately, no discharge data is available for the period over which the tracers were in the stream and therefore this hypothesis could not be tested.

The downstream tracers, however, have experienced size selective transport and this is strongest for those seeded downstream on sand. These tracers show no evidence of gravel overpassing on a sand bed. Conversely, the average distances moved indicate that tracers seeded on gravel patches are more mobile than those seeded on sand. Tracers seeded on gravel may have moved further because the part of the reach in which they were seeded was more dynamic (the thalweg) causing sediment to be transported more rapidly through this part of the stream. This fact may account for the apparent lack of gravel overpassing as the sand patches are less dynamic. As many sand patches are found in areas of shallower flow, for example on the insides of bends, shear stress in these zones was lower, limiting gravel mobility on fine bed sediments.

The probing evidence can be interpreted as showing that the backwater effect from the diversion dam has had an influence on the depositional regime of the GST reach of Allt Dubhaig. This can be seen in Figure 5.6, the simplified one-dimensional diagram of subsurface sedimentology. Prior to dam construction the bed was gravel-dominated to section 12. The united gravel-sand reach extended to section 23, with sand downstream. Following the construction of the dam, the base-level was raised by up to 1 m (Sambrook Smith, 1994), altering the hydraulics in the GST reach. The three zones of differing bed sediments (gravel, united gravel-sand, sand) were shifted further upstream as a result of the backwater. The gravel-dominated reach stretched to section 3, although this has now prograded over the old gravel sediments to section 8. The distance between sections 3 and 8 is approximately 50 m. This gives an average progradation rate of the gravel front of 0.8 m per year since the dam was built in 1930. Surficial bimodal sediments now reach section 15, with sand beyond this. Only a little surficial fine gravel is present in pools or small patches beyond section 15.

Gravel is found at depth up to section 24 around 1 m below the surface. The deep probing at section 27 did not find any gravel indicating that this section was always sand bedded, even prior to dam construction, and therefore the GST has been present for over 70 years and is not a result of the dam. It should be noted, however, that there is very little gravel in the bed, even at depth, downstream of the bend beyond section

16. This morphological feature may have exerted some controlling influence on the location of the transition by causing a decrease in the flow velocity or change in near bed hydraulics. Parker (1998) noted that there was evidence to suggest that sufficient bend sinuosity can stabilise bars in place and prevent their further migration downstream. A similar process may be responsible for the paucity of gravel beyond the bend after section 16 in the Dubhaig. The gravel reach might extend further downstream had the bend not been present.

The GST reaches of both Allt Dubhaig and Vedder River, therefore, experienced little lateral migration in the recent past and exhibit a prograding gravel front beyond the last gravel bar, with mixed gravel-sand sediments immediately downstream, followed by a sand-dominated bed.

Chapter 6. Numerical models of downstream fining and the GST

This chapter details the main numerical models that have been developed to simulate downstream fining. As these models have been used for this purpose it follows that they have the potential to simulate a GST even if they were not explicitly put to this use originally. A review of the general structure and operation of downstream fining models is followed by a consideration of some prominent models. This discussion will provide the rationale for the decision of which model to use for the current research. The reasons for choosing the selected model over the other possibilities are outlined.

The model required for this study must fulfil a number of criteria. The aim is to attempt to create a GST using selective bedload sorting on a width-averaged basis as the only process generating downstream fining. Factors that could complicate the pattern of downstream fining, such as lateral inputs of water or sediment, dropout from suspension, abrasion of gravels into finer fractions or lateral sorting of the bed surface into different grain-size patches, will not be investigated. If a GST cannot be formed using selective bedload sorting alone it can be assumed that one or more of these additional processes is essential in generating a GST. Information gained from the field investigations will aid in identifying which is potentially most crucial.

6.1 Numerical modelling of fluvial systems: limitations

Numerical models are a simplification of reality and for this reason they cannot be used on their own to predict relationships between isolated parts of the natural system to which they are being applied. Caution must always be exercised when applying models and interpreting their predictions.

The accuracy of a numerical model can be limited by a number of factors. Some examples related to fluvial systems are shown below (adapted from Naden, 1988):

1) The errors associated with the flow and sediment transport equations chosen will affect the predictions of the model. The accuracy of sediment transport equations beyond the range of hydraulic conditions and sediment characteristics from which the functions were formulated cannot be guaranteed (Gomez and Church, 1989). Batalla (1997), for example, found variations between data produced by bedload transport equations and observations (with a discrepancy ratio between 0.5 and 2) of between 25 and 68%. The type of flow conditions, the transport rate and bed structuring all have an effect on the choice of bedload transport equation.

2. The type and method of collection of field or laboratory data which are used as an input for the model, or for calibration, has implications for the reliability of model output. These data may include a long profile specification, cross section form information, and bed grain-size characteristics. The way in which the necessary data were collected and then manipulated into a form suitable for use in the model is important information to have available when analysing model results. Bed GSDs derived from sediment samples collected at barheads, for example, may be used as data to set initial conditions of a model run or to test predictions. In the case of models outlined here, however, these data are not ideal since it is assumed that the specified GSD is representative of the sediments spread over the whole channel width rather than just one point (such as a barhead) at a cross section.

3. The significance and importance of factors which are not accounted for in the model may lead to erroneous results. For example, bed armouring or structuring, abrasion of bed sediments, lateral sorting into grain-size patches and a change in base level can all have effects on the rate and size of sediment in transport.

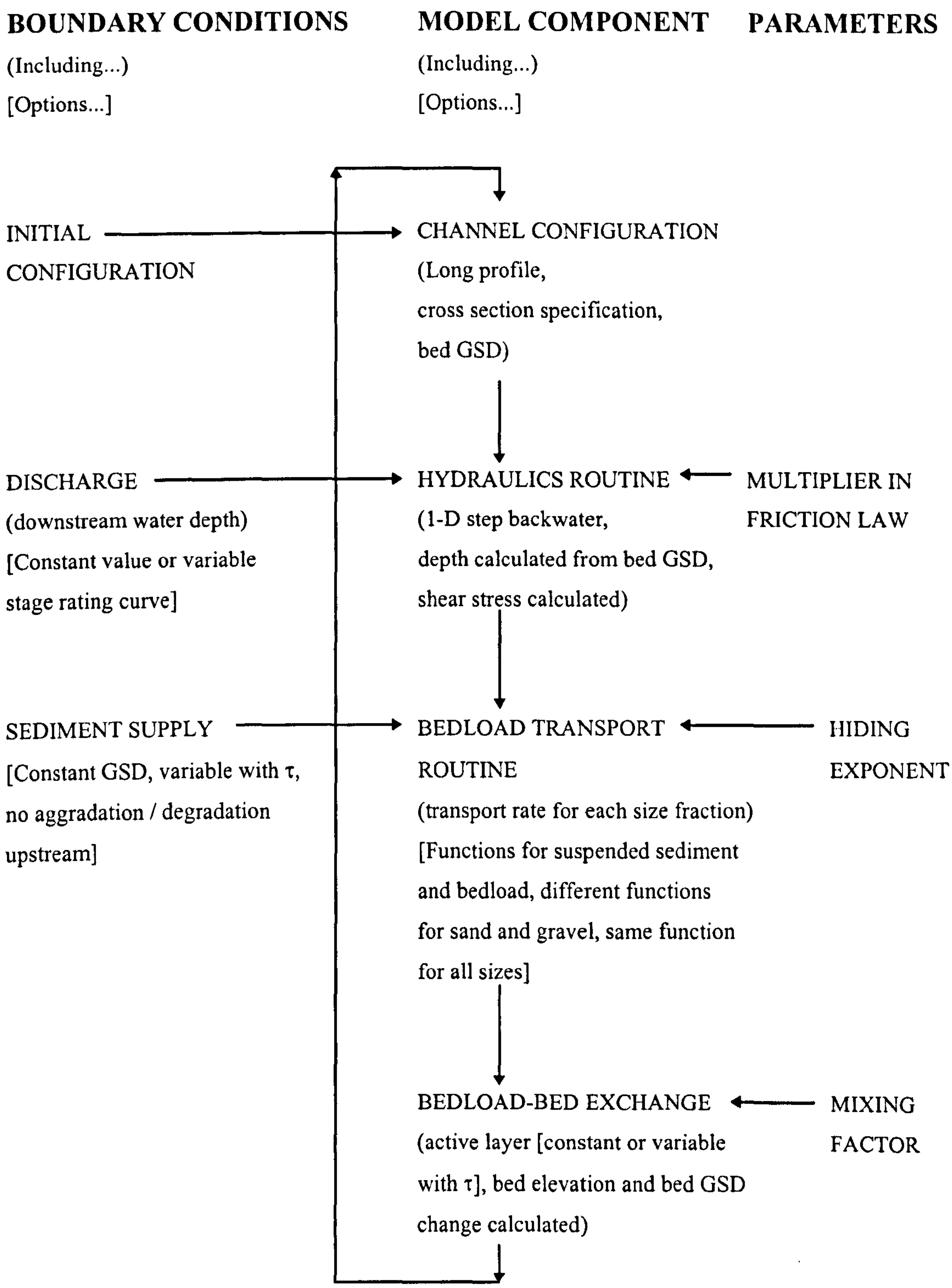
6.2 Numerical simulation of downstream fining

To create downstream fining a given model must have the capacity to simulate the behaviour of mixed-size sediments. Although a number of studies have attempted to model sediment transport and downstream fining in rivers (see, for example, Parker, 1991a,b; Van Niekerk *et al*, 1992; Hoey and Ferguson, 1994; Cui *et al*, 1996), most concentrate on the movement of mixed-size gravel sediments, rather than gravel-sand mixtures. At present it is unclear whether these models can be used to successfully predict the behaviour of gravel-sand mixtures or if the models break down as particle size decreases and additional processes, such as overwhelming of a gravel bed by sand, the lateral sorting of different grain-size patches or gravel overpassing on a sand-dominated bed, become important.

All the models considered here were constructed on the premise that there was some relationship between the rate and amount of sediment movement near the river bed and the specific hydraulic conditions and bed sedimentology in the stream. Geomorphological models based on sediment transport equations focus on the amount of sediment being moved through a channel and the associated erosion and deposition of the bed. Changes in the type and rate of these processes can lead to the evolution of the river's long profile over time.

Models of this type have several generic features: a specified initial long profile; a specified initial bed GSD; specific boundary conditions (for example upstream sediment input and downstream discharge or water level); a hydraulics routine to calculate the flow characteristics; a function calculating the size-specific bedload transport capacity of the simulated flow calculated in the hydraulics routine; a bedload-bed exchange function specifying whether grains are entrained from or deposited to the bed surface or subsurface; and an equation calculating the amount of erosion or deposition that has occurred at a particular point along the reach being modelled. Each of the features influences the overall aggradation or degradation and defines the GSD of the deposited or eroded sediment. A diagram containing these features of mixed-size sediment routing models, and how they interact, is shown in Figure 6.1 below.

Figure 6.1: Generic features of some grain size-specific sediment routing models (adapted from Ferguson *et al*, 1998).



As noted above, all of the models discussed in the current chapter function in a similar way following the structure of Figure 6.1. The simulated reach is divided into a number of cross sections whose characteristics are width-averaged. The importance of lateral variation in grain size and its effect on downstream fining is therefore not considered.

A hydraulics routine calculates the water surface profile using an iterative numerical solution of the flow formula (Equation 2.1, $Q = wdv$) and a friction law (such as the Darcy-Weisbach function, Equation 2.4) relating flow velocity to water depth, slope and bed grain-size. These calculations are carried out using a step-backwater approach. This method uses a specified water depth at the lowest cross section of the simulated reach as a starting point in the calculations. The water depth here is used in the calculation for the next cross section upstream, and so on up the reach.

The shear stress is then calculated using Equation 2.2 ($\tau = \rho gRS$). Using this information transport rates of each of the bedload size fractions (q_i) can be calculated. A function relating the shear stress, the amount of a particular size fraction available for transport (F_i), and its size with respect to the local bed GSD is used, of the form:

$$q_i = F_i \text{ x function of } (\tau, D_i, D_{50}) \qquad \text{Equation 6.1}$$

where D_i is the diameter of the ith grain size and D_{50} is the diameter of the median grain size of the bed sediments.

There are many alternative functions to use for this calculation, some of which were outlined in Chapter 2. In some cases different functions for the sand and gravel, or suspended and bedload fractions may be used.

In a mixed-size sediment hiding and protrusion effects must be taken into account when assessing a grain's mobility. This is often done using equation 2.14 where grain mobility is governed by a specified exponent. When the exponent is set at 0 this corresponds to equal mobility leading to no longitudinal sediment sorting (Parker *et al*, 1982). Parker (1990) suggests that the exponent should be 0.0951, leading to a small degree of size-selective transport of finer materials on the bed surface. This exponent was enough to cause downstream fining during aggradational conditions in the model of Hoey and Ferguson (1994, see Chapter 2, Section 2.1.2, for more details).

If the sediment transport rate does not equal the sediment supply from the cross section immediately upstream the bed will aggrade or degrade and fine or coarsen according to the overall sediment continuity equation (Equation 2.23) and the fractional continuity equation of Parker and Sutherland (1990) (Equation 2.24). These functions allow the channel long profile and GSD characteristics to be updated during a model run and ensure sediment continuity for each fraction through the modelled reach. The process of selective transport conserves the total mass of each grain-size range, but redistributes each size differently along the reach. In the case of abrasion, a transfer of mass from coarser to finer sizes takes place. It is therefore important to account for sediment conservation on a grain-size specific basis if downstream fining is being modelled. The solution to this sediment routing proceeds from the upstream end of the simulated reach in a streamwise direction, as opposed to the step-backwater approach used to calculate the hydraulics through the reach.

A bed active layer must be defined from which sediment is entrained and may be deposited. The thickness of this layer is usually specified as a fixed function of the bed grain-size (for example some multiple of the bed D_{84}) although in some models this thickness varies with shear stress. The way in which bedload interacts with the bed sediment must also be specified. This is done in a bedload-bed exchange function based on a generalisation of the fractional continuity equation. A generalised mixing model (such as that from Hoey and Ferguson, 1994) can be used of the form:

$$E_i = cF_i + (1 - c)\, p_i \qquad \text{Equation 6.2}$$

where E_i is the volume of material of the ith size class in the exchange size, F_i is the volume of material of the ith size class in the active layer, p_i is the volume of material in the ith size class in the bedload and c is the exchange parameter. The exchange parameter (c) lies in the range 0 (all sediment deposited in the subsurface) to 1 (all sediment deposited in the active layer). Each grain-size can have a different specified exchange parameter. Different grain-sizes can be specified to be deposited in or below the active layer in varying proportions.

6.3 Relevant numerical models

The first attempt to model downstream fining by selective sorting was carried out by Rana *et al* (1973) using predictive equations for both suspended and bedload transport in sand bed streams. For a channel with an exponential long profile an exponentially decreasing grain size was produced along the channel and the degree of downstream fining was shown to be dependent on water discharge and sediment concentration.

Deigaard (1982) used the size-specific bedload function of Engelund and Fredsoe (1976) as the basis for a numerical model. At the start of each model run a standardised GSD was specified along the entire channel length and during a run the channel width and discharge were kept constant along the river. The results of model runs showed similar patterns of decreasing grain size to that of Rana *et al* (1973) along the exponential long profile of a gravel-bed river. It was noted that grain size sorting occurred far more quickly than the long profile evolved. Consequently the mean grain diameter along the river always corresponded to the channel long profile shape at any given time (Deigaard and Fredsoe, 1978; Deigaard, 1982). Specific details of other mixed-size sediment routing models necessary to decide which is best suited to the requirements of the current research are outlined below.

6.3.1 Parker (1991) - ACRONYM

Parker (1991a,b) simulated both selective transport and abrasion to model downstream fining in gravel-bed rivers. Sand-sized sediments were not considered. His papers presented a first attempt at providing a framework for the prediction of the effects of selective transport and abrasion. Parker simulated a dynamic equilibrium rather than developing an all purpose model for transient evolution of a gravel-bed river. The channel long profile maintained constant concavity throughout each run but aggraded so that in effect the specified profile moved downstream.

Selective sorting through transport was considered only in a downstream direction, rather than laterally across the channel. The modelling of abrasion was restricted to collision of naturally rounded bedload particles with the bed and each other. Only abrasion to silt was considered and this was subsequently treated as wash load. A surface layer, where sediment could be entrained from or deposited to, was specified to be $1D_{90}$ of the bed surface thick. The bedload function of Parker (1990) was employed by this model. In this function, the selective transport of finer sediments was controlled by hiding, with the coarser particles slightly less mobile than the finer grains. The exponent parameter of this function was calibrated using field data from Oak Creek, USA (Milhous, 1973).

6.3.2 Van Niekerk (1992) - MIDAS

Van Niekerk *et al* (1992) developed a one-dimensional sediment routing model for mixed-size and mixed-density sediments. The model simulated erosion, transport and deposition of various bed material grain-sizes within one straight channel. Bedload transport for each size-density fraction was calculated using the modified Bagnold (1973) equation of Bridge and Dominic (1984) and Vogel *et al* (1992). Critical shear stress for entrainment was derived in the model using the functions of Komar (1989), Egiazaroff (1967) and James (1990). Suspended sediment transport was incorporated using a convection-diffusion sediment continuity equation. A bed continuity equation, solved for each size-density fraction in the active layer, was used to quantify the interaction of the transported load with the bed. The active layer thickness was variable, increasing with higher shear stresses.

Robinson and Slingerland (1998) employed the Van Niekerk *et al* (1992) model to test the sensitivity of downstream fining witnessed in ancient fluvial sediments to a number of variables. These included subsidence rate, sediment flux, water discharge, and hydraulic geometry. Their results demonstrated that subsidence and sediment feed were the most important variables controlling the rate of downstream fining with distance.

6.3.3 Hoey and Ferguson (1994) - SEDROUT

In this model sediment transport was predicted by employing the function of Parker (1990) which specified a low degree of size selectivity. This bedload equation was originally derived as a gravel-only function and was not applied to sand by Hoey and Ferguson (1994). The active layer thickness was defined as a constant function of the bed surface grain-size ($2D_{84}$ of the bed surface). For each individual grain size a different exchange value could be specified defining the proportion of grains of that size deposited on the bed surface (in the active layer) or in the bed subsurface (below the active layer). The long profile shape, which was specified at the start of each run, was free to evolve in association with aggradation or degradation at each cross section.

This one-dimensional sediment routing model was shown to produce reasonable simulations of the downstream fining of gravels for its prototype stream (Allt Dubhaig). The predicted bed grain sizes for the distal part of the prototype were, however, finer than those observed at the field site. The strength of fining in the distal reach predicted by the model was thought likely to be related, to a certain extent, to the unrealistic choice of initial grain size conditions along the channel and also the idealised channel long profile (Hoey and Ferguson, 1994). Prior to the current research there were no published investigations regarding the use of SEDROUT to simulate gravel-sand mixtures.

6.3.4 Cui *et al* (1996) - ACRONYM 2

Cui *et al* (1996) developed a one-dimensional numerical model of downstream fining and used a series of flume investigations as a verification. The results of the flume experiments can be found in Paola *et al* (1992) and Seal *et al* (1997). The transport function employed in the model was taken from the surface based bedload transport relation of Parker (1990). The model used a three-layer system for sediment conservation, containing a bedload, surface (or active) and a subsurface layer. The system for simulating bed surface and bedload sediment exchange (detailed in Toro Escobar *et al*, 1996) was identical to that used by Hoey and Ferguson (1994). The active layer was set to approximately $1D_{90}$ of the bed surface.

When testing the model against the flume runs, the material in the feed finer than 2.0 mm was excluded, in order to fit in with the Parker (1990) bedload function. The agreement between the model predictions for a heterogeneous sediment mix and the experimental findings was generally good indicating that the model successfully described downstream fining with a prograding gravel front (Cui *et al*, 1996).

6.3.5 Parker and Cui (1998) and Cui and Parker (1998) - ACRONYM 5

In these studies Parker and Cui were attempting to simulate a downstream fining profile that was in equilibrium with a stationary gravel front and sand downstream of this. Their model assumed that two processes could cause a GST of this type: abrasion of gravel, or basin subsidence. As a result of the first assumption, once a gravel grain was reduced to a particular size in the model it spontaneously broke down into sand, forcing a GST. Parker and Cui (1998) analysed a simplified model and Cui and Parker (1998) presented a numerical solution of a more complex version of this model. Both were concerned with the equilibrium situation featuring an arrested gravel front.

The model was initially developed on the assumption of two grain sizes, gravel and sand (Parker and Cui, 1998). The analysis was then generalised in the paper to consider continuous GSDs which exhibit a paucity in the fine gravel sizes. For simplicity the paucity of these grains was approximated as a complete absence of such grain-sizes (Cui and Parker, 1998). Also for simplicity, a constant subsidence rate was assumed in the model. In the absence of subsidence or abrasion, the continued supply of gravel sized sediment from upstream should cause the gravel to prograde, shifting the gravel front downstream until base level is reached by the coarser sediment, and a delta is formed.

6.3.6 Gasparini *et al* (1999) - GOLEM

Gasparini *et al* (1999) simulated downstream fining through selective transport of two grain sizes (sand and gravel) for an entire channel network in a river basin to investigate the importance of lateral inputs. Using the model of Tucker and Slingerland (1997), downstream fining emerged as a natural dynamic adjustment to the variables simulated even under conditions of uniform GSD in the sediment flux. Sediment deposition and storage within the basin were not simulated. Each reach within the network was treated one-dimensionally, removing the possibility of lateral sorting of sediments. Different functions for calculating the sediment transport rates of sand and gravel fractions were specified, derived from the relations developed by Wilcock (1997).

6.4 Summary: choice of model

Referring back to the modelling aim specified in Chapter 1, the current research is investigating whether a GST can be formed by selective sorting alone on a width-averaged basis along a single, straight reach. To meet this aim a model is required that simulates size-specific bedload sediment transport using the same function for sand and gravel sizes. To achieve the clearest signal of the importance of size selective bedload sorting the model must not simulate lateral inputs of water or sediment, lateral sediment sorting, abrasion, dropout from suspension or changing bed hydraulic or sediment sorting parameters as the proportion of sand increases. If a GST is not generated by the chosen model then some other processes must be important and these may be elucidated by detailed field investigations at Allt Dubhaig and Vedder River. These streams were chosen for further study because they have no lateral inputs of water or sediment and the GST occurs over a distance sufficiently short to make abrasion of negligible importance.

In the case of rivers exhibiting GSTs in sediments with geologically stable grains, the numerical treatment of abrasion by Parker (1991a,b) seemed an unnecessary complication. Abrasion could be set to zero for all grain sizes but it does not make the Parker (1991a,b) model stronger than the other possibilities. The spatial rate of downstream fining associated with the GSTs in Allt Dubhaig and Vedder River occurred over such a short distance that abrasion could be only of very limited importance in these streams. The abrasion investigations from Allt Dubhaig gravels (reported in Chapter 3) support this hypothesis. Another limitation of the Parker (1990a,b) model was that a GST could not be formed since sand-sized sediments were not present at the start of a model run, or produced during a simulation. All abraded grains created silt-sized sediments rather than the sand sizes necessary to generate a GST.

Van Niekerk *et al* (1992) separated sediment transport into two components, bedload and suspended load. This treatment could introduce an artificial discontinuity into the relationship between the amount of a particular grain size's transport rate, and its abundance in the bed, for a given shear stress. Even if this treatment was realistic it does not meet the requirements necessary for the current research because suspended sediment transport is included. The impact of sediments deposited to the bed as shear stress falls, in association with the reducing bed slope downstream, would blur the importance of selective bedload transport in forming a GST.

A 2 mm lower grain size limit for bedload transport was employed by Cui *et al* (1996). Any additional sand beyond that which could be held in the interstices of the gravel bed was treated as throughput. This fact prohibited the formation of a GST. It is conceptually difficult to visualise a situation where sand is deposited until the interstices in the gravel bed are filled, but after this has occurred, all sand remains in transport as suspended load and is not deposited immediately downstream. If this situation were to take place then a river would perpetually be in a state where the gravel beyond a certain distance downstream would be filled with sand. There would never be a switch to a fully sand-bedded channel, however, even with a consistently decreasing shear stress downstream.

The bedload-bed exchange function of the Cui *et al* (1996) model was calibrated to a flume data set which was also used to test its accuracy (see Toro Escobar *et al*, 1996 for more details). It seems unlikely that the predictions of this function would be as accurate if it was to be tested on field or experimental data collected from elsewhere and therefore transferring this model to another prototype may lead to errors associated with this function. Prior to calibration the function is essentially the same as that used in the SEDROUT model of Hoey and Ferguson (1994).

The model of Parker and Cui (1998) created a GST through the spontaneous breakdown of the fine gravels. For a GST to be formed, however, it was assumed that all particles abraded at the same rate as they travel downstream. It was also assumed that bed material was constantly deposited causing the river to aggrade. This aggradation balanced the subsidence experienced by the river. For many streams these simplifications do not seem reasonable. It is unclear on what data the assumption of the spontaneous breakdown of gravel of a particular size was based. The inclusion of this assumption forced the formation of a GST since gravel was not allowed to exist once it reached a set lower grain size. Gravel was also assumed to abrade to silt, with no sand produced, until this lower grain size was reached (as in Parker, 1991a,b).

The investigations of both Parker and Cui (1998) and Cui and Parker (1998) were based on the assumption that a GST was a non-migrating phenomenon. No timescale over which this assumption might hold was specified, however. It is possible that this situation is occurring in some degrading natural channels since the progradation of an active gravel front would only occur to any great extent during aggradational conditions. If these aggradational conditions were prevalent then channel avulsion within the river valley becomes likely. This switch in channel position would have the effect of altering the bed GSD, "resetting" the progradation of the gravel front. Parker and Cui attempted to compensate for this by assuming that any deposited sediment was distributed across the whole of the active floodplain. This seems to simplify the natural conditions since, with channel avulsion the sedimentary processes would occur at different rates associated with, for example, a change in bed slope. Avulsions may also occur when the gravel front was at a different distance down the river valley. After an avulsion the stream may also have a new sediment source available, if it undercuts a valley side, for example. This new source of sediment may influence the grain size texture of the bed and alter the position of the GST. It is simple and meets the requirements of the current research to simulate a gravel-sand mixture that is contained within a non-migrating, fixed-width channel. However, in investigations related to long-term channel and river valley evolution and dynamics, this simplification may not be valid.

The model of Gasparini *et al* (1999) did create a GST, although the sharp reduction in dominant bed grain size over a short distance associated with natural GSTs was not simulated. Transitions only occurred in simulations where the channel bed was eroded and a sand-dominated subsurface sediment had been specified at the start of the run. The model also used different functions to calculate the sediment transport rates of the sand and gravel fractions. The importance of selective sorting in creating a GST may be blurred since the formation of a GST could an artefact of the different treatment given to the sand and gravel sizes by the model. The overall approach of Gasparini *et al* (1999) also introduces complications into identifying the importance of selective bedload transport by simulating an entire drainage basin stream network rather than a single channel.

As a result of the limitations associated with the models discussed above the SEDROUT model of Hoey and Ferguson (1994) was chosen for the current research. The model simulates only width-averaged size-selective bedload transport as the cause of downstream fining, therefore meeting the requirement of this study. The model was initially developed using Allt Dubhaig as a prototype, one of the two streams investigated in detail for the current research. Since the predictions for this stream were reasonably accurate it is sensible to use and extend this model. The model of Cui *et al* (1996) is very similar to SEDROUT and either model could have been utilised for the current investigations. The reasons for choosing SEDROUT over the Cui *et al* (1996) model are related to the expert advice and assistance available on the use of SEDROUT.

Other reasons for choosing SEDROUT are that model simulations for the current research are carried out on relatively small rivers over short timescales in tectonically stable areas making the inclusion of subsidence an unnecessary complicating factor. SEDROUT treats sand-sized sediments in the same way as gravel rather than using different functions as employed by Van Niekerk *et al* (1992) and Gasparini *et al* (1999). The importance of the role played by size selective bedload sorting in generating downstream fining is therefore clear.

The gravel front is allowed to prograde during SEDROUT model runs, unlike the simulations of Parker and Cui (1998) and Cui and Parker (1998) and results presented in the previous chapter indicate that this progradation occurs in nature. Hoey and Ferguson (1994) also showed that SEDROUT gave better predictions of downstream fining along the gravel reach of Allt Dubhaig than the model of Parker (1991a,b). SEDROUT simulates the behaviour of sediment mixtures in a single channel with no migration (as in the Parker and Cui approach) or lateral inputs of water or sediment (featured in Gasparini *et al*, 1999). Dropout of sediment carried in suspension, which occurred in the model of Van Niekerk *et al* (1992), was not part of the structure of SEDROUT. The lack of these complications make the importance of selective bedload transport easier to identify.

Details of the model structure can be found in Hoey and Ferguson (1994) and are outlined in Chapter 7. It was discovered during model runs simulating the behaviour of gravel-sand mixtures that SEDROUT became unstable. Some development was therefore required to simulate these sediments with a large grain-size range. The enhancement of SEDROUT necessary to simulate gravel-sand mixtures, rather than gravel-only sediments the model has simulated previously, is also explained in the following Chapter. A discussion of the output of the enhanced model is featured in Chapter 8, identifying whether it is capable of generating a GST.

Chapter 7. Modification and testing of SEDROUT for gravel-sand mixtures

The modification of the one-dimensional sediment routing model of Hoey and Ferguson (1994), SEDROUT, to allow the simulation of sorting processes occurring in gravel-sand mixtures in fluvial systems, is outlined in this chapter. This model was introduced in the previous chapter. The results of a structured sensitivity analysis of the influence of sand on the model predictions are then reported. The results from a series of model runs using SEDROUT to simulate gravel-sand mixtures are presented in Chapter 8.

The overall aim of this research project is to elucidate the forms and processes occurring as a river switches from a gravel-dominated to a sand-dominated bed. SEDROUT helps to fulfil this aim because it simulates size selective sorting of bedload which has been postulated as one of the main controlling processes. If the model does not generate a GST then an important factor is missing from the simulations. It could therefore be inferred that this process missing from SEDROUT is crucial in generating a GST, and field investigations are used to define more clearly the role of this process. It is not the aim of the modelling approach to simulate the development of specific a GST and therefore SEDROUT should not be seen as a predictive tool.

7.1 Background

As noted in the previous chapter, many studies have attempted to model sediment transport and downstream fining in rivers. Most investigations concentrated on the movement of gravel-size sediments rather than sand-gravel mixtures. In most cases it is unlikely that the models used in these past investigations can be applied successfully to predict the movement of gravel-sand mixtures. This is because, as outlined in Chapter 2 (Section 2.3), even small proportions of sand (<30% of the width-averaged bulk bed GSD) can greatly alter the dominant bed surface grain size and near-bed hydraulics. SEDROUT was developed by Hoey and Ferguson (1994) for gravel-only sediments and the results of simulations carried out on the gravel-dominated reach of the Allt Dubhaig are reported in Hoey and Ferguson (1994;1997). In these studies it was shown that the downstream fining observed in the prototype was "closely matched by the model predictions" (Hoey and Ferguson, 1994, p2251).

Prior to this research SEDROUT simulated only the sorting processes occurring in river gravels (> 2 mm diameter). In Allt Dubhaig, field investigations have shown that the proportion of sand (<2 mm) in the bed active layer rises from less than 0.2% proximally, to 10% at the distal end of the modelled reach (Hoey and Ferguson, 1994; Ferguson *et al*, 1996). This proportion then rises rapidly, becoming greater than 90% some 300 m further downstream (Sambrook Smith and Ferguson, 1995; Chapter 4 of the current study). To model the gravel-sand transition this 300 m reach must be included in model runs and it is necessary that SEDROUT is able to simulate the behaviour of the gravel-sand and sand sediments found here. Sand has various possible effects on sediment transport which were not considered in the gravel-only version of SEDROUT, including: a modification of the degree of size selectivity in sediment transport as there is evidence of equal mobility within the sand size range (Wathen *et al*, 1995 and Church *et al*, 1991); the bedload transport equation used for gravel sizes may not apply successfully to sand as the bed becomes more poorly sorted; the assumed process and parameter values for bedload-bed sediment exchange are known to vary with grain size (Peloutier *et al*, 1997); the presence of significant amounts of sand on the surface of a gravel bed causes changes in hydraulic roughness, bedform regime and ultimately water surface slope (Sambrook Smith and Ferguson, 1996). Some of these factors, however, are accounted for by SEDROUT for example the bedload-bed exchange parameter can be varied for each half phi fraction being simulated. As the proportion of sand in the bed increases, however, the differences in mobility between gravel and sand may vary (Ikeda and Iseya, 1987; Wilcock, 1993;1998) and the rate of infiltration of fines into a gravel bed may also change (Sambrook Smith *et al*, 1997). Evidence from laboratory experiments suggests that the exchange of sand between the bedload, bed active layer and sub-surface is distinctly different from that of gravel (Peloutier *et al*, 1997). Variations in hydraulic roughness associated with increasing proportions of sand on the bed are taken into account implicitly in SEDROUT the roughness parameter which is calculated using a function of the D_{84} of the bed material (described in Chapter 3).

7.2 Simple debugging

A number of difficulties were encountered whilst attempting to run SEDROUT with a gravel-sand sediment mix and it was found to be much more complicated than anticipated to incorporate sand sized sediments into the model. The problems discussed below were often recurring, in different evolutions of the model, and with varying run conditions. The changes implemented to allow the model runs to be carried out were discussed with Ferguson and Hoey, and the raw code of SEDROUT was altered by Hoey.

7.2.1 Input parameters

SEDROUT is set up so that the specified grain sizes are read as negative phi (or psi) values from the initiation file, and these were then used elsewhere in the model. This led to a problem because, since finer sediments were being used it required a value to be raised to a negative non-integer power for some of the finest grain-sizes, (for example -1.5 and -0.5), causing the model to crash. A line in the code was therefore introduced to overcome this problem in the subroutine which calculates the bedload.

7.2.2 Execution failures

The way that SEDROUT calculates the D_{50} of the bedload was found to contain a flaw. If more than 50% of the bedload was finer than the smallest specified size fraction during a run the model crashed. This is because the model calculates the D_{50} and D_{84} of the bedload sediments by considering size ranges in the classes 1-2, 2-3,....., 15-16. If more than 50% of the bedload is below the finest size fraction the model is unable to calculate a D_{50}. Altering the code in the subroutine which deals with the grain size parameters solved this.

7.3 Parker (1990) bedload transport function

The main difficulty associated with the gravel-only version of SEDROUT was the discovery of some limitations to the bedload transport submodel (outlined below), namely the surface-based Oak Creek model of Parker (1990). The bedload transport function was therefore altered (by Hoey) to overcome this difficulty. This went beyond the realm of a simple debugging procedure and an explanation of the development follows (from Ferguson and Hoey, 1997, pers comm).

To calculate the bedload transport rate of the various size fractions being modelled SEDROUT uses the Parker (1990) surface-based bedload transport relation for gravel bed rivers. The equation is developed from transport rates of gravel-only sizes in Oak Creek (USA) and this has been used to produce a relationship based on surface grain size. Sand sizes were excluded from the analysis even though some sand was present at the site (Milhous, 1973). As noted above, it was discovered that the original Parker (1990) function was not suitable for cases where sediments with a wide range of gravel and sand sizes were present, and modification was therefore required.

7.3.1 Structure of the Parker (1990) bedload function

The sediment routing subroutine of SEDROUT is based around the Parker (1990) procedure, and as such, any difficulties in the application of these equations to gravel-sand sediment mixtures will have implications for simulating a GST successfully and producing accurate model output. To understand the behaviour of the function when routing gravel-sand mixtures reference is made to various parts of the function which are critical in this instance.

This account alters the notation from that used in Parker (1990). The model calculates q_i, the volumetric transport rate per unit width of each half-phi size fraction with representative diameter D_i, from the applied shear stress τ and the GSD of the bed surface, as given by the fraction F_i of sediment in each size class, from which can be calculated the geometric mean diameter D_m and phi standard deviation σ of the bed.

Transport rates are made dimensionless (w^*) using:

$$w^* = q_i Rg/F_i(\tau/\rho)^{1.5}$$ Equation 7.1

in which R is the submerged specific gravity of the sediment, g is the acceleration due to gravity and ρ is the density of water.

The applied stress is made dimensionless using:

$$\tau^* = \tau/R\rho g D_m$$ Equation 7.2

and then used to form a transport stage ratio (φ):

$$\varphi = \tau^*/\tau^*_r$$ Equation 7.3

where τ^*_r is a threshold-type reference stress for the mean diameter and was set to 0.0386, as a best fit to the Oak Creek data. Differences in the mobility of the various fractions are incorporated by multiplying φ by the hiding function $(D_i/D_m)^{-\beta}$ with β = 0.0951 (again fitted to the Oak Creek data); the result is called φ\`. In effect, this makes the reference stress for size $D_i$ become slightly size-dependent: $\tau/R\rho g D_m^{1-\beta} D_i^{\beta}$. In ACRONYM, $\varphi$ is further multiplied by a straining parameter $\omega$ which is explained below; this converts $\varphi$\` to φ\`\`.

The adjusted transport stage φ\` is converted to a transport rate for the ith grain size using:

$$w^*_i = 0.00218G(\varphi``)$$ Equation 7.4

The function G(x) is in three parts: $G = x^{14.2}$ for $x<1$, $G = \exp[14.2(x-1)-9.28(x-1)^2]$ for $1 \leq x \leq 1.59$, and $G = 5474(1-0.853/x)^{4.5}$ for $x>1.59$, where x is a dummy variable. The values of the coefficients ensure smooth matching of G and dG/dx. The shape of this function is shown in Figure 7.1. The transport rate therefore depends on three functions: the straining function (described below); a stress ratio (φ), and a hiding function evaluated for each size fraction in turn. Of these, the first two are crucial in understanding the limitations of the bedload transport relation in this case. The hiding function returns constant results for any given size distribution of material and cannot, therefore, be the cause of the observed problems.

The straining function, ω, is required because the bedload function is a development of the earlier model of Parker *et al* (1982) which used the bulk subsurface GSD. Converting this to a surface-based model requires allowance for the anticipated change in surface GSD with changing applied stress: less coarse and less well sorted as stress rises (Parker, 1990). For Oak Creek this tendency is quantified in two empirical curves which appear in Figure 5 of Parker (1990) and as lookup tables in the ACRONYM software and SEDROUT; they are reproduced as Figure 7.2 here. The phi standard deviation of the bed (σ) is assumed to be steady at rather over 0.8 for low stresses and transport stages ($\varphi <0.9$), then increase gradually to over 1.3 at $\varphi = 3$. The straining parameter ω is close to 1 at low stresses but decreases beyond $\varphi \approx 1$ to about 0.6 at $\varphi = 3$. The straining function is an attempt to generalise the curves in Figure 7.2 to other channels with surface GSDs which have a different degree of sorting than that found in Oak Creek.

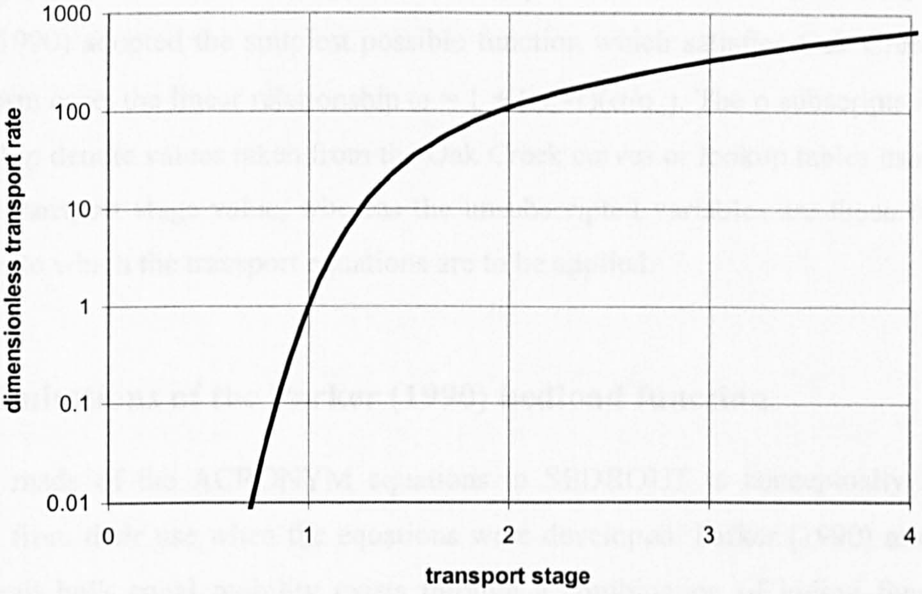


Figure 7.1: Surface based Oak Creek bedload transport function, G(x), as in Parker (1990). Plotted for mean bed diameter (no hiding). Transport stage denotes φ` and dimensionless transport rate is w^*_i.

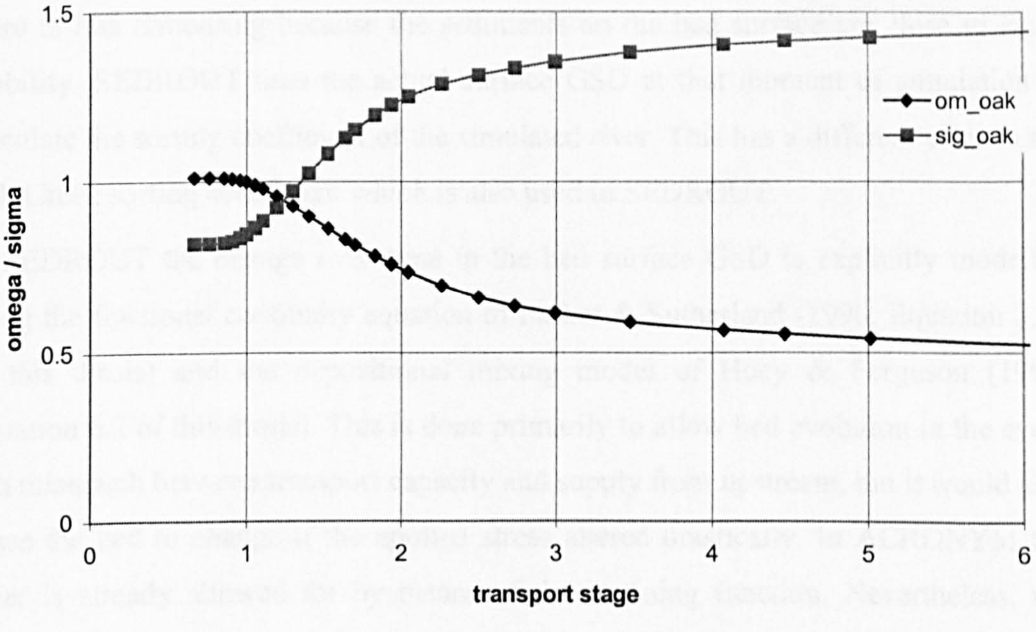


Figure 7.2: The Oak Creek straining functions (after Parker, 1990). Transport stage denotes φ`. A value of 1 corresponds to the quasi-threshold reference stress for mean diameter. Sig_oak denotes σ_o, and om_oak denotes ω_o.

It is assumed that uniform sediment requires no straining since its GSD cannot alter with stage; ω should therefore tends to 1 as the phi standard deviation σ tends to 0. Parker (1990) adopted the simplest possible function which satisfies Oak Creek and the uniform case: the linear relationship $\omega = 1 + (\omega_0-1)(\sigma/\sigma_0)$. The o subscripts in this relationship denote values taken from the Oak Creek curves or lookup tables using the required transport stage value, whereas the unsubscripted variables are those for the new river to which the transport equations are to be applied.

7.3.2 Limitations of the Parker (1990) bedload function

The use made of the ACRONYM equations in SEDROUT is conceptually rather different from their use when the equations were developed. Parker (1990) assumed that overall bulk equal mobility exists through a combination of hiding functions (micro and macro hiding of Parker and Klingeman, 1982) and armouring. The bed surface GSD is therefore a function of stress, rather than a function of the history of the channel. This explains the fact that the sorting coefficient (σ) is a function of applied stress, rather than a constant. At high applied stress the equilibrium surface layer is considerably finer and more poorly sorted (Parker, 1990). This means that there is less armouring because the sediments on the bed surface are close to equal mobility. SEDROUT uses the actual surface GSD at that moment of simulation to calculate the sorting coefficient of the simulated river. This has a different basis to the Oak Creek sorting coefficient which is also used in SEDROUT.

In SEDROUT the change over time in the bed surface GSD is explicitly modelled using the fractional continuity equation of Parker & Sutherland (1990, Equation 2.24 of this thesis) and the depositional mixing model of Hoey & Ferguson (1994, Equation 6.2 of this thesis). This is done primarily to allow bed evolution in the event of a mismatch between transport capacity and supply from upstream, but it would also cause the bed to change if the applied stress altered drastically. In ACRONYM the latter is already allowed for by means of the straining function. Nevertheless, the initial work on downstream fining using SEDROUT (see Hoey and Ferguson, 1994) was carried out using the full ACRONYM equations including the straining function. When work began on extending SEDROUT to gravel-sand mixtures it was decided to investigate how ACRONYM would work when applied to sizes <2 mm. There seemed to be no physical reason why these functions would be inaccurate and no clear difference in the behaviour of sand and gravel bedload fractions had been detected in the analyses of fractional transport rates in several gravel-bed and mixed-bed fieldsites (see, for example, Ashworth and Ferguson 1989; Ashworth *et al*, 1992).

After carrying out runs with the full ACRONYM functions it was found that for a particular channel long profile, the model produced highly unlikely output values for the slope and sediment transport rates. It seemed that the fine grained sediment in the bed caused the sediment routing to become unstable. Before the model crashed the long profile of the channel levelled off and the flow became super-critical. Sediment transport (including those for very fine grain sizes) dropped to zero even at shear stresses theoretically high enough to entrain the bed sediments. Part of the bedload transport algorithm became unstable after the introduction of fine sediment into the bed and feed, as the standard deviation of the bed material increased. In some cases, even if only a small percentage of bed sand was included, SEDROUT appeared to stall altogether. This stalling was found to be associated with a shortening of the computational time step, which is automatically varied in the SEDROUT forward finite difference scheme. After extensive investigation the reason for the shortened time step was discovered. Major differences in transport rate were occurring from one section to the next, causing rapid local aggradation and breaks of slope at locations and times where the channel exhibited a poorly sorted bed. This, in turn, was found to be the result of implausible output from the transport submodel: low, sometimes near-zero, transport rates at high shear stresses where the bed was poorly sorted. This resulted in the stalling of bedload, yet the resultant steepening of the bed slope and thus increase in shear stress which would normally lead to enhanced transport and remove the "bump", now caused enhanced aggradation due to the decreased transport rate.

The problem is illustrated in Figure 7.3 by calculations using the ACRONYM equations without a hiding function, hence for the mean diameter of the bed GSD, and for different values of the phi standard deviation (σ). For Oak Creek, with $\sigma < 1$, transport increases monotonically with stress.

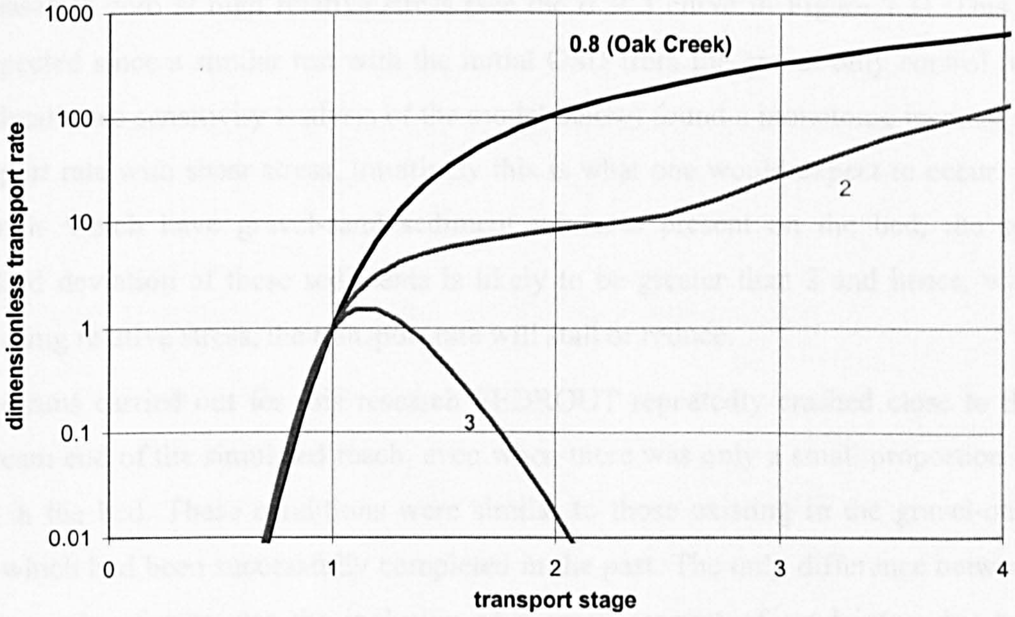


Figure 7.3: Behaviour of the transport function G(φ``) for the mean diameter in a poorly sorted bed (phi standard deviation ,σ, of 2 or 3) using the Parker (1990) straining function. The original Oak Creek relationship is also plotted. Transport stage denotes φ` and dimensionless transport rate is w^*_i.

For σ beyond approximately 1.6, however, the curve develops an inflection, with only a slow rise in transport for a big increase in stress around a transport stage of 2 (see the $\sigma = 2$ curve in Figure 7.3). Beyond $\sigma \approx 2.1$ the inflexion becomes a turning point and transport reduces with increasing stress, though remaining positive over the plausible range of stress. In even less well sorted beds, beyond $\sigma \approx 2.6$, transport decreases to zero at high relative stress (see the $\sigma = 3$ curve in Figure 7.3). This is unexpected since a similar test with the initial GSD from the gravel-only control run (for details see sensitivity analysis of the model below) found a monotonic increase in transport rate with shear stress. Intuitively this is what one would expect to occur. In channels which have gravel-sand sediment mixtures present on the bed, the phi standard deviation of these sediments is likely to be greater than 2 and hence, with increasing relative stress, the transport rate will stall or reduce.

In the runs carried out for this research SEDROUT repeatedly crashed close to the upstream end of the simulated reach, even when there was only a small proportion of sand in the bed. These conditions were similar to those existing in the gravel-only runs which had been successfully completed in the past. The only difference between the two sets of runs was the inclusion of a small amount of sand into what had previously been a gravel-only run. From this it can be inferred that the cause of the problems was the increase in the proportion of fine sediments on the bed of the simulated river. As noted above, the reason for these failures, and unexpected results prior to crashing, appeared to be an instability in the transport relation under certain conditions, causing alternating phases of aggradation and degradation. The important controlling factor on whether a model run crashed, or not, seemed to be the degree of sorting of the particular bed GSD and sediment feed material which is specified in the start-up files before a model run is undertaken.

Why this occurs mathematically is apparent on consideration of the linear form of the Parker straining function:

$$\omega = 1 + (\omega_o - 1)(\sigma/\sigma_o) \qquad \text{Equation 7.5}$$

together with the values of the curves in Figure 7.2. For transport stages <1 $\omega_0 \approx 1$ so $\omega \approx 1$ also, irrespective of the value of σ. But for high transport stages $\omega_0 < 1$ (asymptotically <0.5) so a multiple of σ/σ_0 is subtracted on the right-hand-side of the equation and ω can thereby become <<1 or even negative. The value of G, which is normally very high at high transport stages, is thereby depressed.

Using numerical, and experimental evidence it can therefore be inferred that the problem is associated with the straining function of Parker's (1990) function which does not accurately simulate very poorly sorted surface GSDs, for example those found in a river with a gravel-sand sediment mix on the bed.

7.3.3 Development of the Parker (1990) bedload function

An alternative straining function, therefore, has to be introduced which also fits the data presented by Milhous (1973) and utilised by Parker (1990). Parker has two known results from which he has developed the straining function: (1) uniform sediment for which ω=1, and; (2) Oak Creek, with a surface phi standard deviation of 1.011, where ω declines as a function of shear stress. Parker's straining function assumes that there is a straight line which can be extrapolated between these two results, and that this is a function of the surface standard deviation of the bed sediment. This is unrealistic and it is more likely that the amount of straining should not increase as rapidly as the standard deviation increases and therefore the power that the standard deviation of the bed GSD is raised to should be less than 1. As a short-term fix, for the purposes of this research, Hoey developed a modified straining function which use nonlinear interpolation between, and extrapolation beyond, the uniform-bed and Oak creek cases. The modified function is:

$$\omega = 1 + (\omega_0 - 1)\sigma^{0.3}/\sigma_0 \qquad \text{Equation 7.6}$$

Figure 7.4 compares the transport-rate predictions using the Parker and modified Parker straining functions in a poorly-sorted case (σ = 3)

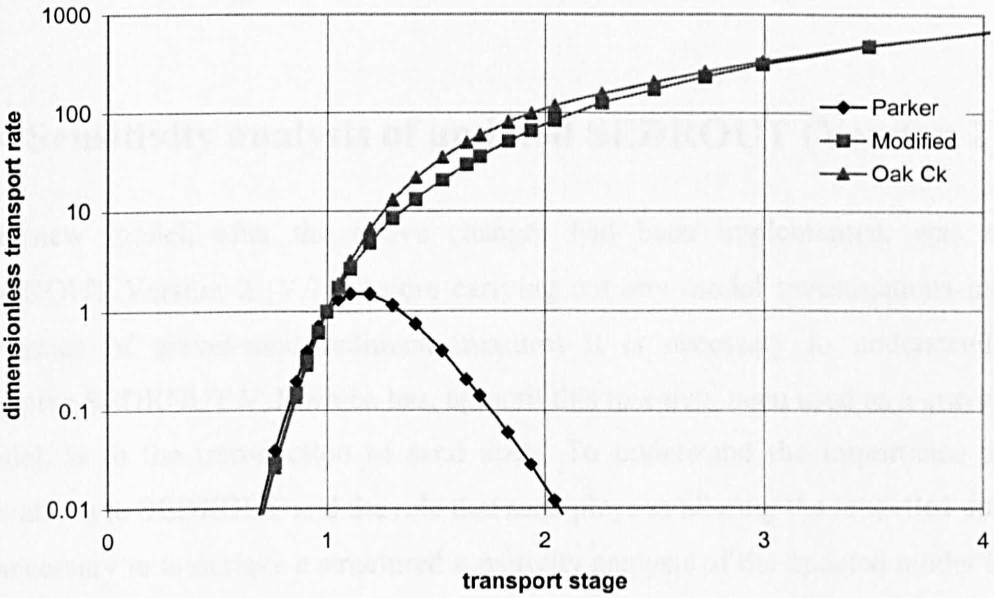


Figure 7.4: Behaviour of the modified straining function for mean diameter in a poorly sorted bed (phi standard deviation, σ, of 3) compared with the Oak Creek relationship and the Parker (1990) straining function.

As before, the plotted curves are for the fraction containing the mean bed diameter. It can be seen that the modified straining gives results much closer to those for Oak Creek, without the anomalous shoulder in the transport curve. Results for even more poorly-sorted cases are also robust, as can be seen in the model runs presented elsewhere in this thesis.

7.4 Sensitivity analysis of updated SEDROUT (Version 2)

The new model, after the above changes had been implemented, was named SEDROUT Version 2 (V.2). Before carrying out any model investigations into the dynamics of gravel-sand sediment mixtures it is necessary to understand how sensitive SEDROUT V.2, which has, up until this research, been used as a gravel-only model, is to the introduction of sand sizes. To understand the importance of any alterations to SEDROUT and the role that sand plays in altering the modelled output it is necessary to undertake a structured sensitivity analysis of the updated model for the current research.

As noted above, work carried out by Hoey and Ferguson (1994) tested the gravel-only version of the model on a field prototype in Scotland (the Allt Dubhaig). Results of a sensitivity analysis on SEDROUT were reported by Hoey and Ferguson (1997). While attempting to understand the role of sand it is sensible to build on this knowledge and that gained during my field investigations, and continue to use the Allt Dubhaig as the prototype on which to test the model. The initial runs carried out on the Allt Dubhaig (by Hoey and Ferguson, 1994) were undertaken without any calibration and gave an acceptable match to the observed downstream fining profile (Hoey and Ferguson, 1994). SEDROUT, however, contained several parameters which had either been derived empirically in the field or laboratory, or were hypothetical. In the sensitivity analysis of SEDROUT, presented in Hoey and Ferguson (1997) a single parameter was varied at a time from its value in a control run (details below) in order to identify the influence each exerted on the development of a downstream fining profile. This technique was employed for the investigations presented here.

7.4.1 Comparison of SEDROUT and SEDROUT V.2

To investigate the influence that altering the Parker (1990) bedload transport relation present in SEDROUT V.2 has had, the gravel-only control run simulated by SEDROUT and discussed in Hoey and Ferguson (1994), was re-run for a 3500 m reach. The run is called CONT3500. The output from CONT3500 is compared to a run which was identical, apart from the use of the new transport relation. This run is called STR_3500. These model runs show that although the alterations to the function do exert some influence over the model predictions, these variations are small, certainly compared to those seen when altering other variables in the model during the sensitivity analysis. At T_w (the first time the active layer D_{50} at 3 km was below that at any cross section upstream) the D_{50} at 1 km was predicted by the original function to be 41.0 mm, whereas the modified function gives a value of 45.0 mm. At 3 km these values are 7.3 mm and 7.2 mm respectively. The fining wave moves more slowly when using the original function than with the modified straining function.

7.4.2 Default model run

The same long profile and initial bed GSD were used in the sensitivity analysis of SEDROUT V.2 as those employed by Hoey and Ferguson (1994;1997). The simulated reach was extended from 2800 m to 3500 m to include the stretch of river that, at Allt Dubhaig, exhibits the GST. A small proportion of sand was included in the bed material of the control run, as was originally found in the field when the data was collected. This sand was excluded from earlier investigations. The modified Parker (1990) bedload function was employed when carrying out the default run. The initial conditions of the gravel-sand control run, to which model output will be compared, are as follows. The run started with an exponential long profile fitted to the surveyed prototype, with a best fit concavity of 0.895 km^{-1}. As noted above, the reach was lengthened, from that investigated by Hoey and Ferguson (1994), to 3500 m, minimising end effects. The reach was defined by 36 rectangular cross sections, 10 m wide and 100 m apart, and deep enough so that the flow did not go overbank. The measured bed surface bulk GSD at a distance (x) of 0.22 km was assumed to extend along the entire reach at the start of each run (time, t = 0 minutes). This GSD has a D_{50} and D_{84} of 85 and 178 mm respectively, was specified at half-phi fractions, was truncated at 0.25 mm, and contained 1.53% sand. Discharge was held constant throughout the entire run at 5 m^3s^{-1}, which is a high near bankfull flow at most places along the prototype reach and was exceeded 2% of the time through the reach between 1990 and 1993 (Ferguson and Wathen, 1998). The upstream boundary condition used in the control run was that of no aggradation or degradation at x = 0 (the first cross section). The supply of each sediment size fraction was varied to match its capacity at x = 0. A bedrock sill in the prototype at x = 0 supports the use of this boundary condition (see Chapter 3).

7.4.3 Sensitivity analysis model runs

The parameters varied In the sensitivity analysis are shown in Table 7.1, along with their control run values.

Table 7.1: Parameter values varied in the sensitivity analysis model runs.

Parameter	Units	Control-run value	Alternative values
Discharge, (Q)	m^3s^{-1}	5	3, 8, 20
Hiding parameter, β (sand)		0.0951	0, 0.2
Hiding parameter, β (gravel)		0.0951	0, 0.2
Exchange parameter, c (sand)		1	0.5, 0
Active layer thickness (*D_{84}), k	m	2	1, 4
Porosity, λ		0.3	0.1, 0.5
Roughness coefficient, a		1.1	0.5, 1.5
Concavity, b	km^{-1}	0.895	0.6, 1.2

The modified Parker (1990) bedload function has two crucial parameters, the dimensionless reference shear stress (τ^*_{r50}) at which median sized bed material is transported at a low dimensionless rate, and the hiding factor (β) which indicates the extent to which the threshold stress for size fraction *i* is dependent on the relative size

D_i/D_{50}. Equal mobility occurs when $\beta = 0$, and independent Shields-type behaviour for each size in the bed holds when $\beta = 1$. Values of β were both increased and decreased in the sensitivity analysis (runs SENSI_8-12 and SENSI_17-19, respectively), and τ^* was investigated indirectly by altering the discharge (SENSI_6-7, and SENSI_20).

The exchange parameter (c) was varied for sand sizes from its default value of 1 (all deposited in the active layer) in the control run (SENSI_1), to 0.5 (SENSI_15), and 0 (all deposited in the subsurface, SENSI_16). The thickness of the active layer itself was varied from $2D_{84}$ in the control run to $1D_{84}$(SENSI_3) and $4D_{84}$ (SENSI_2), as was the porosity of the bed (SENSI_4-5). The degree of concavity was both increased and decreased from the value given in the control run (SENSI_21-22).

The terminology of Hoey and Ferguson (1997) is employed to describe the two aspects of fining that are being investigated: its strength, meaning the downstream change in grain size, and its rate, meaning the change over time of the grain size and fining profile.

7.4.4 Method of comparison

In terms of the formation of a GST, the main area of interest is the GSD of the active layer, which should fine more rapidly with distance if a GST is to be formed. Hoey and Ferguson (1997) used the variation in D_{84} of the active layer to define surface grain size. As this study is more concerned with the active layer GSD as a whole, rather than the coarsest grains, the D_{50} is investigated. The percentage of sand in the bed surface is also considered. It was found that each run has two distinct phases. Firstly, a wave of fine sediment progrades through the reach, associated with fining and rapid aggradation at each cross sectional node as it passes. This is then followed by a lower rate of aggradation and coarsening towards an equilibrium. A reliable measure that distinguishes between the two phases is the time taken (T_w) for the fine wave to prograde through the reach of interest, and is defined as the first time the active layer D_{50} at 3.0 km falls below the size of any cross section upstream. The strength of fining developed by time T_w is given by the D_{50} at $x = 1.0$ and 3.0 km, and the distance, L_h, from $x = 0$ for the active layer D_{50} to halve (cf Hoey and Ferguson, 1997). The model results were dumped to output files every 20000 minutes.

7.4.5 Summary of model simulations

Various model runs were carried out during the sensitivity analysis and any parameter which could be influenced by the inclusion of sand into the model runs was investigated. The model output of the sensitivity analysis runs is shown Table 7.2.

As stated above, the model is not being used to simulate the field conditions present in Allt Dubhaig, rather as a tool to investigate the processes that may be important in causing a GST. As such, a comparison of the modelled results with the field data will not be undertaken. The sensitivity of overall fining should be compared to that in the gravel-sand control run, rather than the conditions witnessed in the prototype.

A comparison of how the runs carried out for the sensitivity analysis vary in D_{50} at time T_w can be found in Figure 7.5. Figure 7.6 shows the contrast of the downstream fining profiles at time T_w, giving an indication of L_h, the half-distance of the bed D_{50}.

7.4.6 Gravel-sand control run simulation

A short analysis of the predictions of the control run, to which other model runs carried out during the sensitivity analysis are to be compared, shows the following: at T_w (180000 minutes of model time) the D_{50} at 1 km is 43.2 mm, and the bed at this point contains 3.5% sand. The half-distance (L_h) is 1.1 km. At 3 km the D_{50} has fallen to 5.2 mm, and the sand content has risen to 30.2%.

Table 7.2: Model output from sensitivity analysis runs. T_w, x and L_h are defined in the text.

Run	T_w (model mins)	D_{50} (mm) and % sand (at T_w) at x=1 km	D_{50} (mm) and % sand (at T_w) at x=3 km	L_h (km)
SENSI_1 (G/S Control)	180000	43.2, 3.5	5.2, 30.2	1.1
CONT3500 (G control)	480000	41.0, 0	7.3, 0	1.0
STR_3500 (G cont, mod P1990)	360000	45.0, 0	7.2, 0	1.1
SENSI_2 (K=4)	360000	46.1, 3.3	6.2, 24.6	1.1
SENSI_3 (K=1)	120000	41.6, 3.7	4.1, 31.0	1.0
SENSI_4 (porosity=0.1)	220000	42.9, 3.6	5.7, 31.7	1.1
SENSI_5 (porosity=0.5)	140000	43.6, 3.5	5.1, 27.7	1.1
SENSI_6 (discharge=8)	120000	57.9, 2.7	4.8, 30.1	1.4
SENSI_7 (discharge=3)	340000	34.9, 4.4	3.8, 39.0	0.9
SENSI_8 (S=0 hiding)	220000	44.1, 3.4	6.2, 22.3	1.1
SENSI_9 (S/G=0 hiding)	∞	84.7, 1.5	84.7, 1.5	∞
SENSI_10 (S=-0.2 hiding)	160000	42.7, 3.6	4.6, 36.6	1.1
SENSI_11 (G=-0.2 hiding)	100000	28.0, 4.8	4.2, 27.5	0.7
SENSI_12 (G/S=-0.2 hiding)	100000	28.0, 4.8	4.0, 29.9	0.7
SENSI_13 (0.5 roughness)	160000	51.0, 2.9	5.0, 27.4	1.3
SENSI_14 (1.5 roughness)	220000	40.3, 3.8	4.8, 32.3	1.0
SENSI_15 (c=0.5 for S)	340000	46.2, 2.9	7.4, 12.3	1.1
SENSI_16 (c=0 for S)	420000	47.5, 2.5	7.9, 5.9	1.2
SENSI_17 (S=-0.05 hiding)	200000	43.7, 3.5	5.7, 25.6	1.1
SENSI_18 (G=-0.05 hiding)	440000	57.8, 2.8	7.3, 28.0	1.4
SENSI_19 (S/G=-0.05 hiding)	500000	58.6, 2.7	8.2, 22.4	1.5
SENSI_20 (discharge=20)	40000	73.5, 2.0	8.3, 20.8	2.0
SENSI_21 (concavity=1.2)	240000	42.9, 3.6	2.8, 44.5	1.1
SENSI_22 (concavity=0.6)	200000	49.1, 3.0	10.2, 12.8	1.3

Fig 7.5

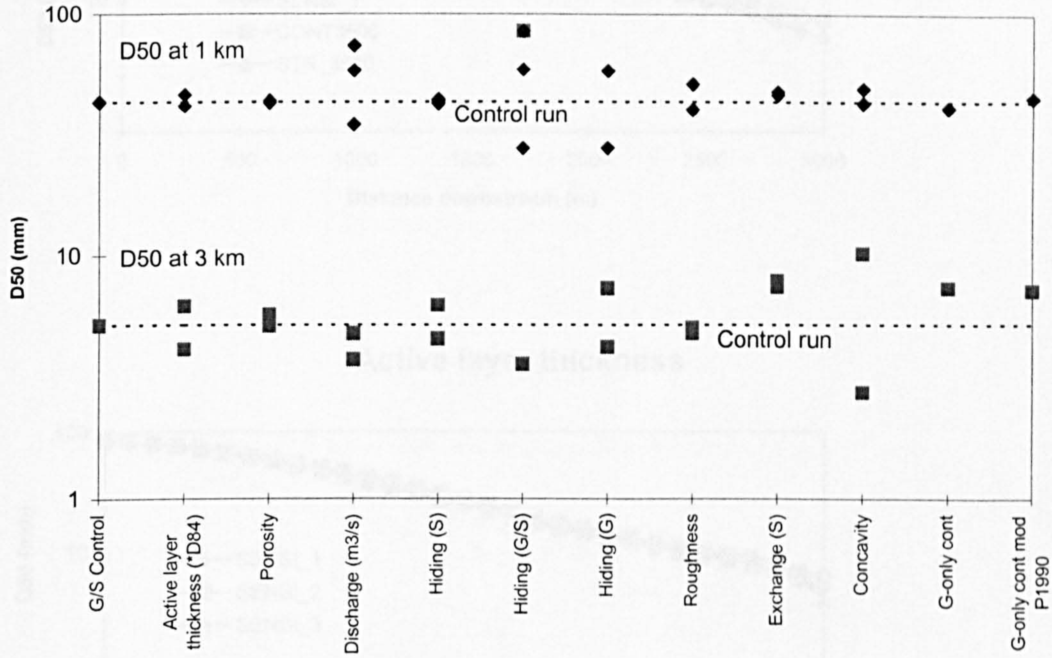
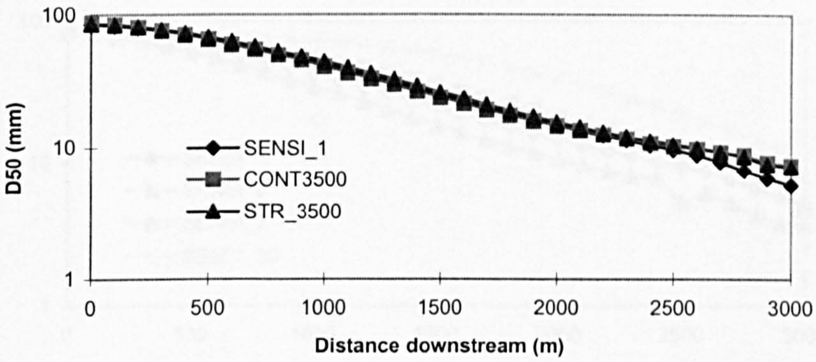
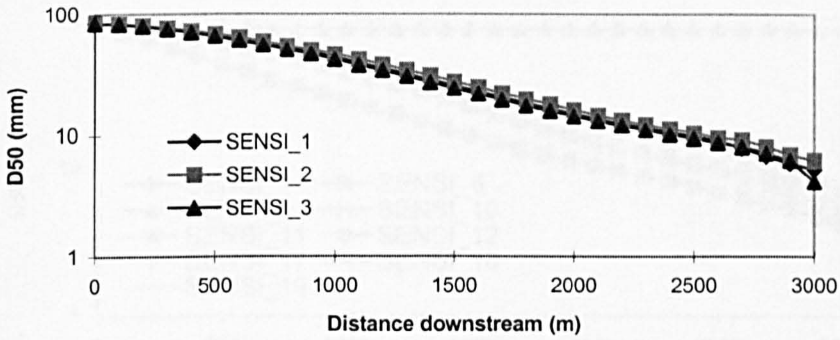


Figure 7.5: Output of SEDROUT V.2 sensitivity analysis model runs showing bed surface D_{50} at time T_W for various parameters. G and S indicate gravel and sand respectively P1990 denotes the original Parker (1990) bedload transport relation. The figure is best viewed in association with Table 7.2, showing parameter values.

Control Runs



Active layer thickness



Bed Porosity

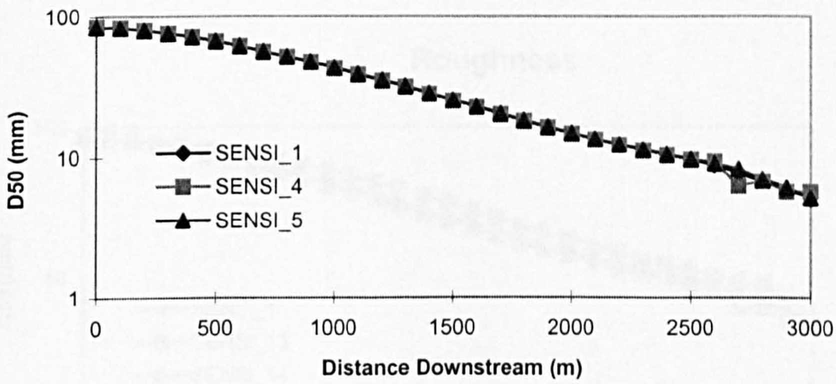
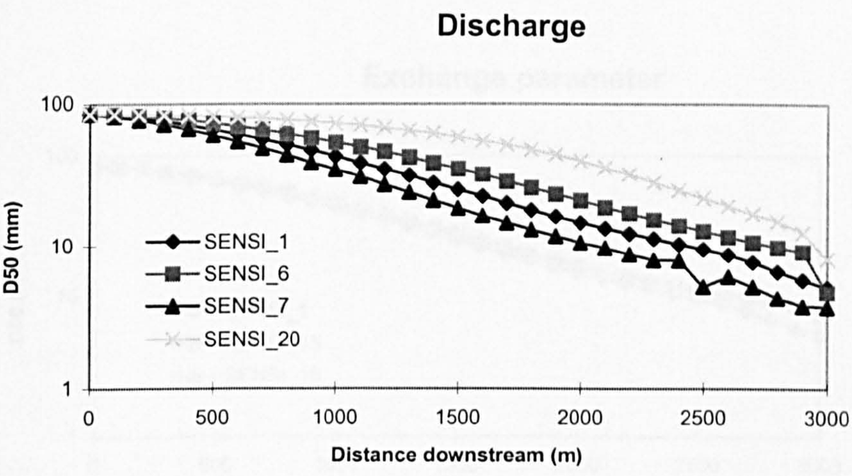
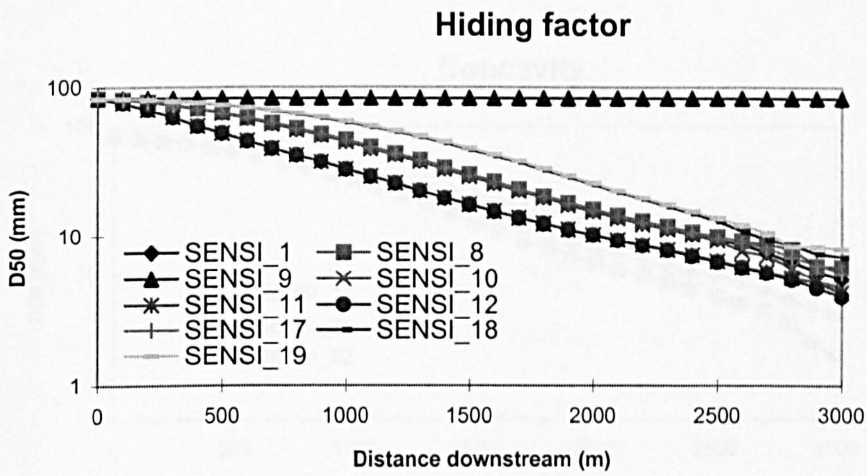


Figure 7.6: See page 23 for caption.





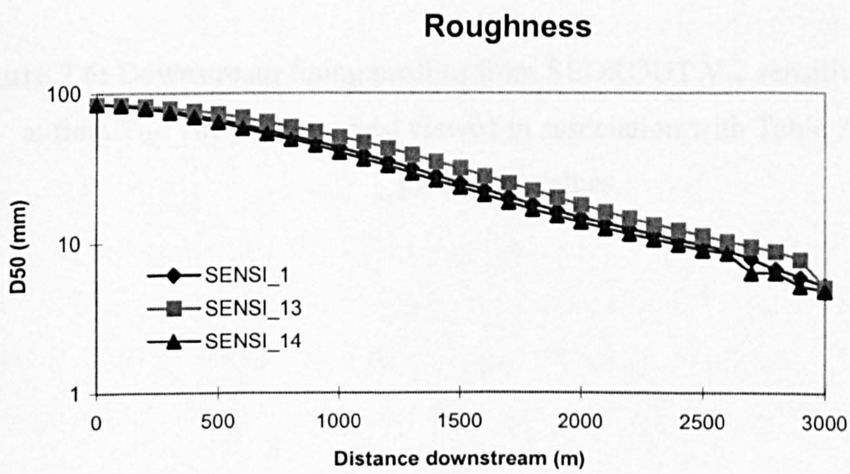


Figure 7.6: See page 23 for caption.

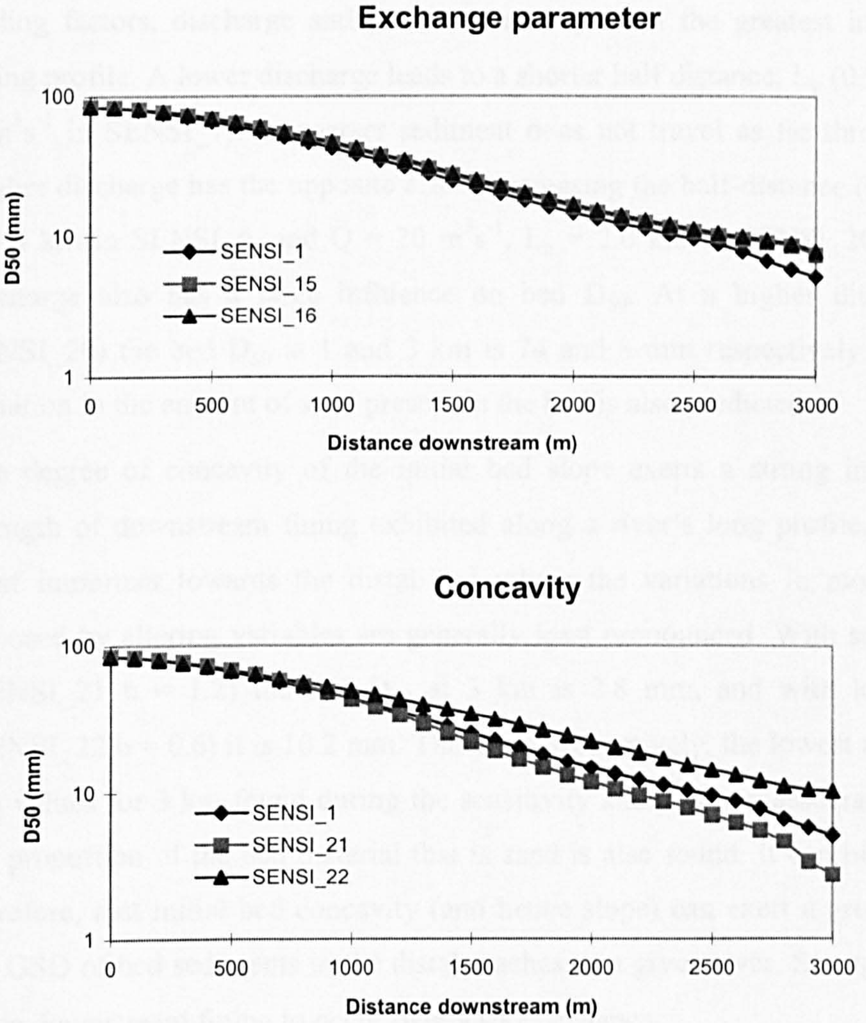


Figure 7.6: Downstream fining profiles from SEDROUT V.2 sensitivity analysis runs at time T_W. The plots are best viewed in association with Table 7.2, showing parameter values.

7.4.7 Strength of fining

Hiding factors, discharge and profile concavity exert the greatest influence on the fining profile. A lower discharge leads to a shorter half distance, L_h (0.9 km when Q = 3 m^3s^{-1} in SENSI_7), as coarser sediment does not travel as far through the reach. Higher discharge has the opposite effect, increasing the half-distance (Q = 8 m^3s^{-1}, L_h = 1.4 km in SENSI_6, and Q = 20 m^3s^{-1}, L_h = 2.0 km in SENSI_20). Altering the discharge also has a large influence on bed D_{50}. At a higher discharges (as in SENSI_20) the bed D_{50} at 1 and 3 km is 74 and 8 mm respectively. An associated variation in the amount of sand present in the bed is also predicted.

The degree of concavity of the initial bed slope exerts a strong influence on the strength of downstream fining exhibited along a river's long profile. This becomes most important towards the distal end where the variations in model predictions imposed by altering variables are generally least pronounced. With strong concavity (SENSI_21 b = 1.2) the bed D_{50} at 3 km is 2.8 mm, and with lower concavity (SENSI_22 b = 0.6) it is 10.2 mm. These are, respectively, the lowest and highest bed D_{50} values for 3 km found during the sensitivity analysis. An associated variation in the proportion of the bed material that is sand is also found. It can be clearly stated, therefore, that initial bed concavity (and hence slope) can exert a great influence on the GSD of bed sediments in the distal reaches of a given river. Strong concavity can force downstream fining to occur over a short distance.

It can be seen in Figure 7.5 that in the upstream part of the reach (x = 1 km) gravel hiding factors are important (SENSI_11, gravel hiding 0.2, D_{50} = 28.0 mm and SENSI_18, gravel hiding = 0.05, D_{50} = 57.8 mm). As the hiding factor is reduced, and the various sediment sizes approach equal mobility, the half distance of the D_{50} increases. The reverse is true for larger hiding factors. In the distal reach, where rapid downstream fining is expected, the mobility of gravel remains an important control on bed GSD but the degree to which sand sizes are selectively transported also becomes crucial. This is surprising as at the start of the run only 1.5% of the bed material is sand. At lower slopes, altering parameters associated with this initially small proportion of the bed material, however, can have appreciable results. Depending on the degree of size selectivity of the sand fraction, the proportion of sand in the bed at 3 km (where the GST occurs in the prototype) can vary from 22.3% (in SENSI_8, sand hiding 0), to 36.6% (in SENSI_10, sand hiding 0.2). This is significant as the two values are on either side of 30% which has been regarded (by Sambrook Smith *et al*, 1997) as the necessary amount of sand present in a bed to cause a switch from a gravel framework to a sand matrix (and therefore to initiate a GST). With equal mobility for sands (SENSI_8) the D_{50} of the bed material at 3 km is 6.2 mm, as opposed to 5.2 mm in the control run (SENSI_1, sand hiding 0.0951). Church *et al* (1991, p2951), suggested that equal mobility of sands "could describe a period average condition of the sediment transport", and since the model is run for a relatively long period, these are the conditions that we would expect in the simulations. With sand hiding set at 0.2 (SENSI_10) the D_{50} at 3 km is 4.6 mm. The influence of sand mobility at more proximal locations is less important. This is because gravel is more easily transported at these higher slopes and therefore the coarser sediments exert a greater influence over the bed D_{50}.

The thickness of the active layer from which sediment can be readily entrained during a model run also exerts an influence on the fining profile (SENSI_2 and 3). A thicker active layer leads to coarser bed surface. Active layer thickness exerts less influence at 1 km than 3 km (as with all other variables investigated except concavity). There is, however, still a 50% difference in the bed D_{50} from an active layer thickness of $1D_{84}$ (SENSI_3, D_{50} = 4.1 mm) and $4D_{84}$ (SENSI_2, D_{50} = 6.2 mm). The reason for the variation associated with active layer thickness is due to the fact that a different GSD of sediment available for transport will be calculated for each active layer thickness specified. The difference may also be due to the depth to which the deposited fine sediment is mixed. A thicker active layer (as in SENSI_2) will take longer to alter by a given amount for this reason.

Both the bed porosity (which affects the aggradation depth for a given rate of deposition) and the coefficient of the roughness equation (which affects flow depth and therefore shear stress) have only a limited effect on the strength of downstream fining. In proximal reaches, where coarser sediments are present on the surface, there is some deviation in bed D_{50}, particularly associated with low roughness coefficients (SENSI_13, D_{50} = 51.0 mm) and L_h is also increased (to 1.3 km).

The exchange variable was investigated for the sand sized material as it is likely that sand infiltration into a gravel bed is an important process in GST formation. The reasons for this are outlined in Chapter 2 (Section 2.3.3). The function does not exert a large influence on bed D_{50} at the 1 km point at T_w, but its importance becomes greater downstream. Here, larger proportions of sand are deposited. The D_{50} increases as the exchange parameter is reduced and therefore more fines are deposited below the active layer (D_{50} at 3 km increases to 7.9 mm when the exchange parameter = 0). Varying the exchange parameter, therefore, impacts greatly on the proportion of sand present in the bed surface. With all the sand deposited in the subsurface (SENSI_16) the amount present at 3 km at T_w is only 5.9%, almost an order of magnitude less than the amount required to initiate a GST. It is unlikely, however, in an aggrading gravel bed channel, that all the sand deposited on the bed would remain on the surface, as in the control run (SENSI_1). With the exchange parameter set so that half of the sand is deposited in the subsurface, and half on the surface (SENSI_15), the proportion of sand in the bed at 3 km is found to be 12.3%. This is still significantly less than that required to generate a GST.

7.4.8 Rate of downstream fining development

As discussed above, the time taken for the D_{50} at 3 km to fall below that at any point upstream (T_w) was taken to be an indication of the rate of development of the downstream fining profile. In the control run, T_w, was 180000 minutes of model time. Since the initial conditions of the run (with a standardised GSD present at each cross section simulated) are arbitrary, a detailed analysis of the rate of development of a downstream fining profile is unlikely to prove useful for the purposes of the current research. In all cases during the sensitivity analysis runs the fining wave had reached 3 km by 500000 minutes of model time. The most rapid progradation of the fining wave through the reach occurred with high discharge (SENSI_20, discharge = 20), when T_w was 40000 minutes of model time.

7.5 Summary

This chapter outlined the main objectives that need to be met in order for a numerical model to aid in the investigation of the main processes operating in gravel-sand sediment mixtures. While the majority of the problems were quite simple to correct, the failure of the bedload transport function was considerably more challenging. The cause of the failure was discovered and examined in detail. This allowed a successful fix to be implemented in SEDROUT. Simulation of the sorting processes occurring in gravel-sand mixtures was successfully carried out following the alterations made to the model. The updated model may, however, require testing against independent bedload transport data before it is used as a predictive tool.

Since SEDROUT V.2 had not previously been applied to gravel-sand mixtures prior to the current research it was felt prudent to carry out a sensitivity analysis. This investigation showed that there are only small differences between the model predictions of SEDROUT V.2 and the gravel-only version of the model. Variations in hiding factors, discharge and concavity had the strongest effect on the downstream fining profile produced by SEDROUT V.2. Investigations into the importance of the exchange parameter, which controls the amount of fine sediment infiltration into a gravel bed, provide the basis for further investigations into this parameter in Chapter 8.

Chapter 8. Simulating the behaviour of gravel-sand mixtures using SEDROUT

This chapter outlines the results of investigations into the effectiveness of SEDROUT in simulating the rapid downstream fining associated with a GST. The results build on the information presented in the previous chapter describing the enhancement of SEDROUT to deal with gravel-sand mixtures and the degree to which the model is sensitive to different parameter values. After setting out the general procedure for carrying out the model runs, the influence that the initial conditions and run parameterisations exert on model output is investigated. The simulations that ran successfully are then summarised and this is followed by a short discussion about those runs which failed. A more detailed discussion of the model in relation to field data will be undertaken in Chapter 9. Reasons why SEDROUT may not accurately represent the behaviour of gravel-sand mixtures will also be discussed in this later chapter.

Model crashes occurred on a number of occasions and were particularly associated with simulations of GST evolution using real GSD and slope information collected at the fieldsites as the initial conditions for a run. Reasons for these difficulties are briefly discussed and, where appropriate, the methods employed to overcome specific problems are outlined. Because of the difficulties associated with running simulations based on real field data, Allt Dubhaig and Vedder River were used as prototypes, and SEDROUT is used to investigate behaviour in gravel-sand mixtures in general

Initial investigations continued to use Allt Dubhaig as the prototype. This was due to the fact that the majority of the previous research using SEDROUT (including the sensitivity analysis of Chapter 7) was based on this stream. Once modelling investigations had been completed on Allt Dubhaig SEDROUT was applied to Vedder River to understand the transferability of the model. Causes of differences between the model output for the Dubhaig and the Vedder are discussed in the summary at the end of this chapter.

8.1 General procedures

As noted above, the aim of the model runs is to simulate accelerated and spatially extensive downstream fining between a gravel and a sand bed using SEDROUT. The initial runs are based on Allt Dubhaig. The general procedural form of the investigation is outlined in this section. Relevant details of the runs carried out on Allt Dubhaig can be found in Table 8.1.

In order to shorten the length of time each model run took, the simulated reach, which was 3500 m long in the sensitivity analysis runs, is shortened. The proximal sections are removed and only the last 2000 m (with a node spacing of 100 m) or 1100 m (with a node spacing of 50 m) are included in the simulations. These two series of runs are named DISTAL and BLT respectively. The upstream limit of these runs is in the region of the bedload trap in the prototype. Stream lengths shorter than these are not simulated to avoid forcing a GST by inputting the specified feed containing a large proportion of sand immediately upstream of the expected GST zone in the model run.

Simulations are also carried out using Vedder River as the prototype. These runs are aimed at testing the transferability of SEDROUT to a river an order of magnitude larger than Allt Dubhaig. The Vedder runs are based on simulations carried out by Ferguson and Church (unpublished) with the study reach extended from 11000 m to 14400 m to include the GST zone in the prototype. The causes of differences between Allt Dubhaig and Vedder River model predictions are discussed in the summary. Relevant details of runs carried out on Vedder River can be found in Table 8.2. The results of the runs carried out on Vedder River are outlined in Section 8.3.4 below.

In this chapter the following parameters and boundary conditions are varied: the initial slope and bed GSD; the exchange parameter; the upstream boundary condition (feed GSD); and the discharge. Of these, the specified initial slope and bed GSD of the runs are likely to be most critical, following the findings presented in the previous chapter. If a gravel bed is assumed throughout the reach then the infiltration and possible saturation of the gravel matrix with sand were the main processes that were to be simulated. Field observations (see Chapters 4 and 5), however, suggest downstream progradation of the distal end of the gravel bed over sand-dominated sediments is the process occurring at the study sites today. An experiment which exhibits downstream fining as an initial condition is undertaken to examine this process (see Section 8.2).

Table 8.1: Name and important details of SEDROUT model runs based on Allt Dubhaig discussed in this chapter. Y / N refers to yes / no, c is the bedload-bed exchange parameter. See text for further details.

Run name	Ran	Investigating?	Points to note
SENSI_1	Y	Baseline simulation for comparison	Basic sensitivity analysis run
DOGLEG_1	Y	Model stability	c=0 for sand, break of slope
DOGLEG_2	N	Model stability	c=1 for sand, break of slope
DISTAL_1	Y	c parameter	c=1 for sand. Break of slope. Variable channel widths
DISTAL_2	Y	Initial bed slope	Break of slope in initial long profile
DISTAL_4	Y	c parameter	c and hiding=0 for sand
DISTAL_5	Y	Sand input at upstream end	Sand in feed, c=0 for sand
DISTAL_6	Y	Sand input at upstream end	Sand in feed, c=1 for sand
DISTAL_13	Y	c parameter	>30% sand in subsurface during run. c=0 for sand
DISTAL_14	Y	c parameter	As DISTAL_13, run until >30% sand in subsurface
DISTAL_15	Y	c parameter	Uses NEWSTART files from DISTAL_14. c=1 for sand
DISTAL_16	Y	Initial long profile	As DISTAL_13, with smooth concave long profile
BLT_1	Y	Sediment feed	As DISTAL_13 but 1100 m long (not 1600 m). Smooth concave long profile
DISTAL_20	Y	Initial bed slope	Smooth concave initial long profile
BLT_2	Y	Initial bed GSD	As BLT_1 with initial downstream fining and a different feed GSD
BLT_11	Y	Initial bed GSD	As BLT_2 with no initial downstream fining
BLT_12	Y	Discharge	As BLT_11 with Q=7
BLT_13	Y	Discharge	As BLT_2 with Q=3
BLT_15	Y	Initial bed GSD	As BLT_1 with finer initial bed GSD
BULK_1A	N	Real field conditions	Real grain size, long profile and width data
BULK_1B	N	Real field conditions	Smoothing the rate of change of gradient at the break of slope
BULK_1C	N	Real field conditions	Flat initial long profile (slope = 0.002)
BULK_1D	N	Real field conditions	As BULK_1A with constant channel width of 10 m
BULK_6	N	Real field conditions	As BULK_1D with B parameter in roughness calculation =1.7

Table 8.2: Name and important details of SEDROUT model runs based on Vedder River. Y / N refers to yes / no. See text for further details.

Run name	Ran	Investigating?	Points to note
VEDGST_01	Y	Initial long profile	Smooth concave long profile. No initial downstream fining
VEDGST_02	Y	Initial long profile	As VEDGST_01 with a break of slope
VEDGST_03	Y	Initial bed GSD	As VEDGST_01. Coarse sediments up to cross section 50. Then fine
VEDGST_04	Y	Initial bed GSD	As VEDGST_02. Coarse sediments up to cross section 50. Then fine
VED_1	N	Real field conditions	Real grain size, long profile and width and sediment feed data

Data collected from the field sites can be used to run and test the predictions of SEDROUT. Detailed information is available for both field sites regarding bed GSDs, cross section geometry, slope, discharge and upstream boundary condition. This data was collected for the present research and previously for a period of time greater than five years. There was therefore the potential to test SEDROUTs ability to simulate real conditions. It should be noted, however, that simulating real GSTs is not the main aim of this thesis.

8.2 Influence of initial conditions on model output and stability

The impact that varying the specified parameter values has on model output was discussed in the previous chapter. While carrying out runs to simulate a GST the degree to which simulations are sensitive to initial conditions and specified parameter values also becomes apparent.

When modelling Allt Dubhaig the fining wave (discussed in the previous chapter) did not prograde through the entire simulated reach in some of the runs. Figure 8.1 is a plot of the bed D_{50} with distance downstream at time T_W, when the fine wave has reached 3 km, from gravel-sand control run (SENSI_1). This figure shows that the distal stream bed was coarse at the end of the run when compared to the rest of the channel upstream. This fact is associated with the general form of the initial conditions employed in the model simulations. All successful runs (except BLT_2) had the same coarse bed GSD specified throughout the reach at the start of the simulation. As a result of these start-up conditions, and long profile concavity, the coarse sediments at the distal nodes take longer to be transported past the end of the reach as the slope is lower. This causes the shear stress to decrease downstream. At the start of run SENSI_1, for example, the shear stress at the top of the reach is 77 N/m^2, compared to 7 N/m^2 at the distal limit of the simulation. By the end of the run there is still an order of magnitude difference in the shear stress between the proximal and distal ends of the simulated reach. Coarser sediments from upstream are therefore deposited and the coarse fractions from the cross sections in the distal part of the simulation are not removed at a fast rate. These coarser sediments are therefore left as a relict of the specified initial conditions at the end of the run or until aggradation increases the slope to the extent that they can be entrained. The specified length of time for a run to be carried out therefore greatly influences the GSD at the distal-most cross sections. While the fine wave prograded through most of the cross sections relatively rapidly, the lower slope at the distal cross sections increases the time taken for the fining wave to pass. For this reason it is prudent to remove the lowest few sections from any discussion of model output when the initial conditions were set so that all the cross sections simulated have the same GSD. In the case of SENSI_1 in Figure 8.1, for example, the last 5 cross sections would be excluded from the analysis of the downstream fining profile at time T_W. This procedure removes end effects that could otherwise influence the interpretation of the model output.

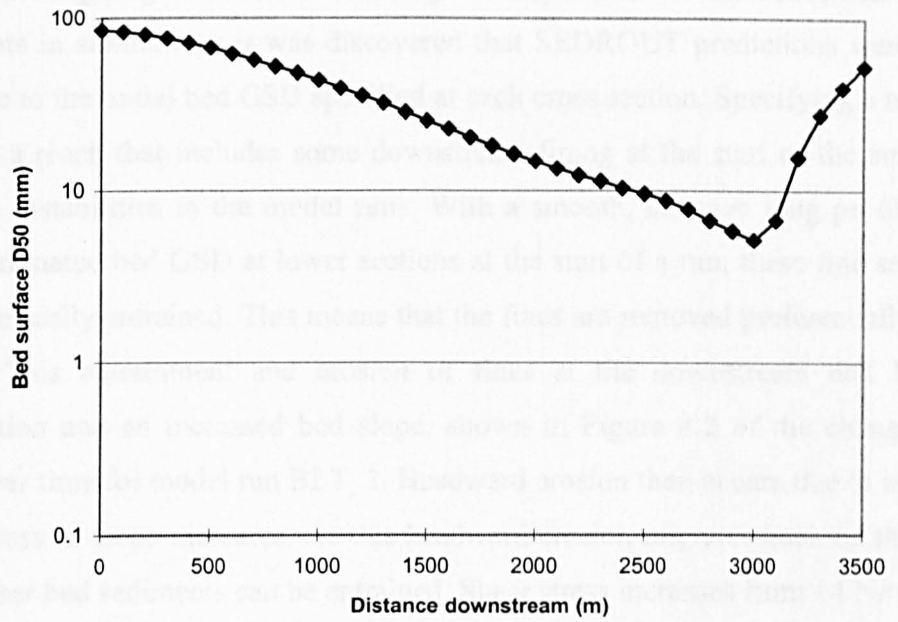


Figure 8.1: Bed surface D_{50} with distance downstream at time T_W from the gravel-sand control run (SENSI_1).

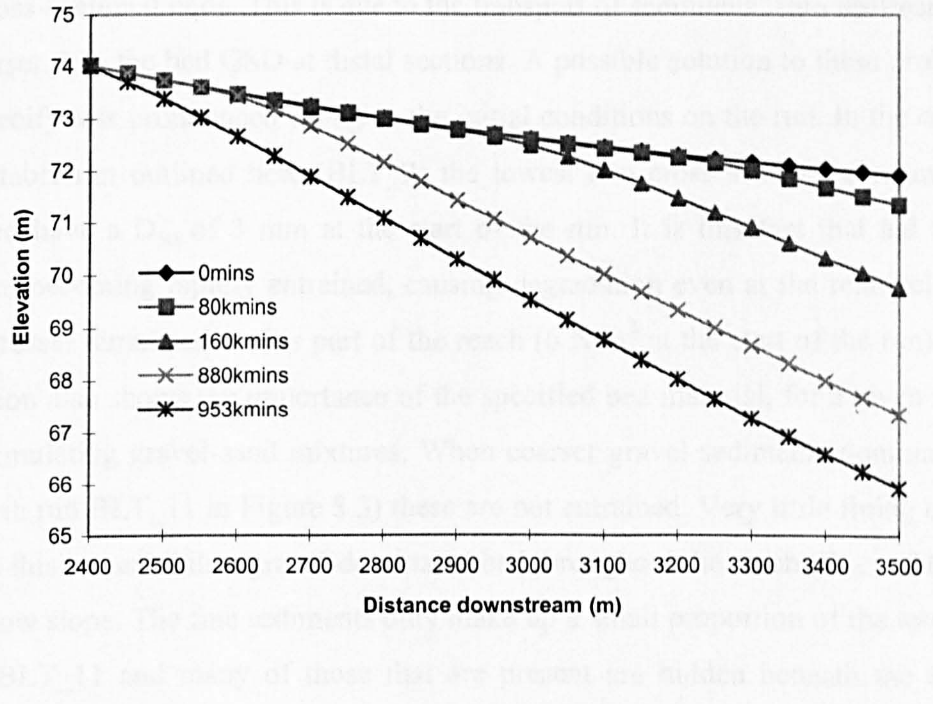


Figure 8.2: Changing bed slope over time from run BLT_2. The numbers in the legend are x10^3 minutes of model time.

While investigating methods of reducing the importance of the relict coarse distal sediments in simulations it was discovered that SEDROUT predictions were highly sensitive to the initial bed GSD specified at each cross section. Specifying a bed GSD through a reach that includes some downstream fining at the start of the simulation leads to instabilities in the model runs. With a smooth, concave long profile and a sand-dominated bed GSD at lower sections at the start of a run, these fine sediments are more easily entrained. This means that the fines are removed preferentially by the model. This entrainment and erosion of fines at the downstream end leads to degradation and an increased bed slope, shown in Figure 8.2 of the changing bed slope over time for model run BLT_2. Headward erosion then occurs due to increased shear stress as slope increases. As this headward erosion migrates back up the reach, the coarser bed sediments can be entrained. Shear stress increases from 14 N/m^2 at the head of the reach at the start of the run to 22 N/m^2 at the end of the simulation. At the end of the run the sediment that is present on the bed of the simulated reach is dominated by that provided from the specified feed material. If there is too much sand in the initial bed at some sections a coarsening wave also passes through part of the reach, rather than the fining wave witnessed in all runs with the same coarse GSD at each cross-sectional node. This is due to the transport of sediments from upstream that are coarser than the bed GSD at distal sections. A possible solution to these problems is to specify less pronounced fining in the initial conditions on the run. In the case of the unstable run outlined here (BLT_2) the lowest two cross sections contain 48% sand and have a D_{50} of 3 mm at the start of the run. It is this fact that led to the sediment becoming rapidly entrained, causing degradation even at the relatively low shear stresses simulated in this part of the reach (6 N/m^2 at the start of the run). This simulation also shows the importance of the specified bed material, for a given slope, when simulating gravel-sand mixtures. When coarser gravel sediments dominate the bed (as in run BLT_11 in Figure 8.3) these are not entrained. Very little fining occurs because this run exhibits a gravel-dominated bed throughout the reach (D_{50} = 31 mm) with a low slope. The fine sediments only make up a small proportion of the total bed in run BLT_11 and many of those that are present are hidden beneath the coarse grains because near equal mobility is specified in all runs. The supply of fine sediment required to initiate a GST is therefore not present in runs with a low distal slope and coarse initial bed GSD.

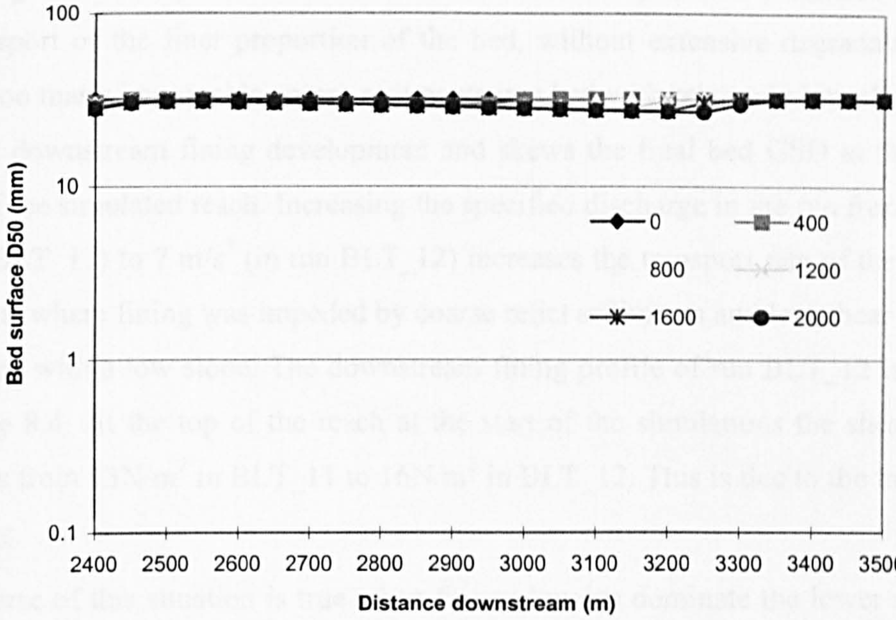


Figure 8.3: Change in bed surface D_{50} over time with distance downstream from run BLT_11. The numbers in the legend are x10^3 minutes of model time.

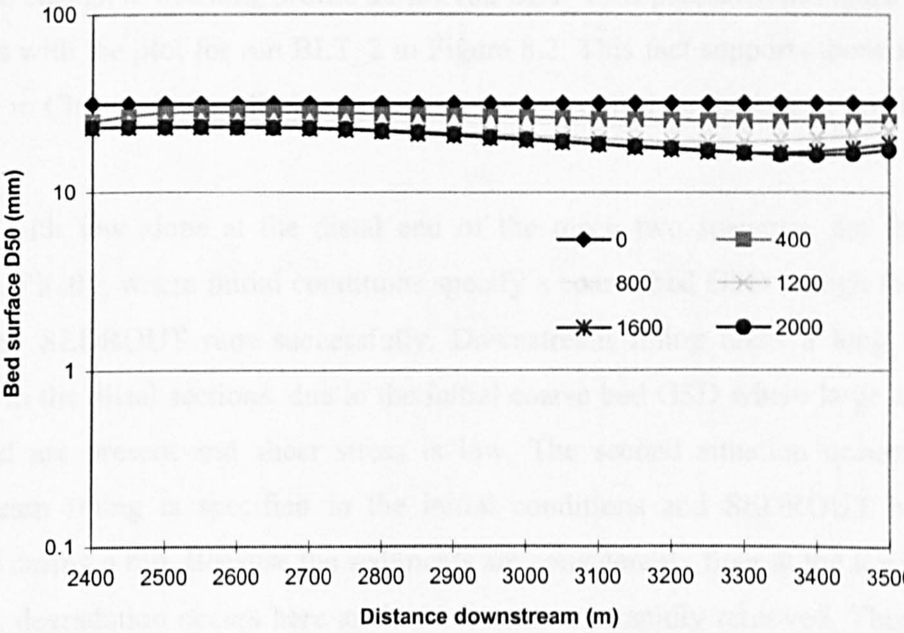


Figure 8.4: Change in bed surface D_{50} over time with distance downstream from run BLT_12. The numbers in the legend are x10^3 minutes of model time.

Following the investigations outlined above it is necessary to find a balance between the transport of the finer proportion of the bed, without extensive degradation, and having too many immovable coarse sediments in a bed with relatively low slope. This impedes downstream fining development and skews the final bed GSD at the lower nodes of the simulated reach. Increasing the specified discharge in the run from 5 m/s^3 (in run BLT_11) to 7 m/s^3 (in run BLT_12) increases the transport rate of the coarser sediments where fining was impeded by coarse relict sediments and low shear stresses associated with a low slope. The downstream fining profile of run BLT_12 is shown in Figure 8.4. At the top of the reach at the start of the simulations the shear stress increases from 13N/m^2 in BLT_11 to 16N/m^2 in BLT_12. This is due to the increased discharge.

The reverse of this situation is true when fine sediments dominate the lower sections of the modelled reach. If the run where headward erosion occurs (BLT_2) is repeated using a lower discharge of 3 m/s^3, instead of 5 m/s^3, (in run BLT_13) it is found that rapid degradation and headward erosion do not occur. The shear stresses predicted at the lower end of the reach are 6 N/m^2 (BLT_2) and 4 N/m^2 (BLT_13). These findings indicate that the model is highly sensitive to both initial bed GSD and discharge. The predicted change in bed long profile during run BLT_13 is presented in Figure 8.5 and contrasts with the plot for run BLT_2 in Figure 8.2. This fact supports those findings reported in Chapter 7, that discharge exerts a strong control on the rate of downstream fining.

In runs with low slope at the distal end of the reach two scenarios are therefore possible. Firstly, where initial conditions specify a coarse bed GSD though the entire reach and SEDROUT runs successfully. Downstream fining takes a long time to develop in the distal sections, due to the initial coarse bed GSD where large amounts of gravel are present and shear stress is low. The second situation occurs where downstream fining is specified in the initial conditions and SEDROUT becomes unstable during a run. Because the sediments are considerably finer at the lower cross sections, degradation occurs here as the sediments are rapidly removed. This causes an increased bed slope and, in turn, headward erosion. Once the headward erosion reaches the upstream limit of the simulation the coarse fraction of the bed and feed sediments can be transported at the new higher slope, therefore decreasing the bed D_{50}.

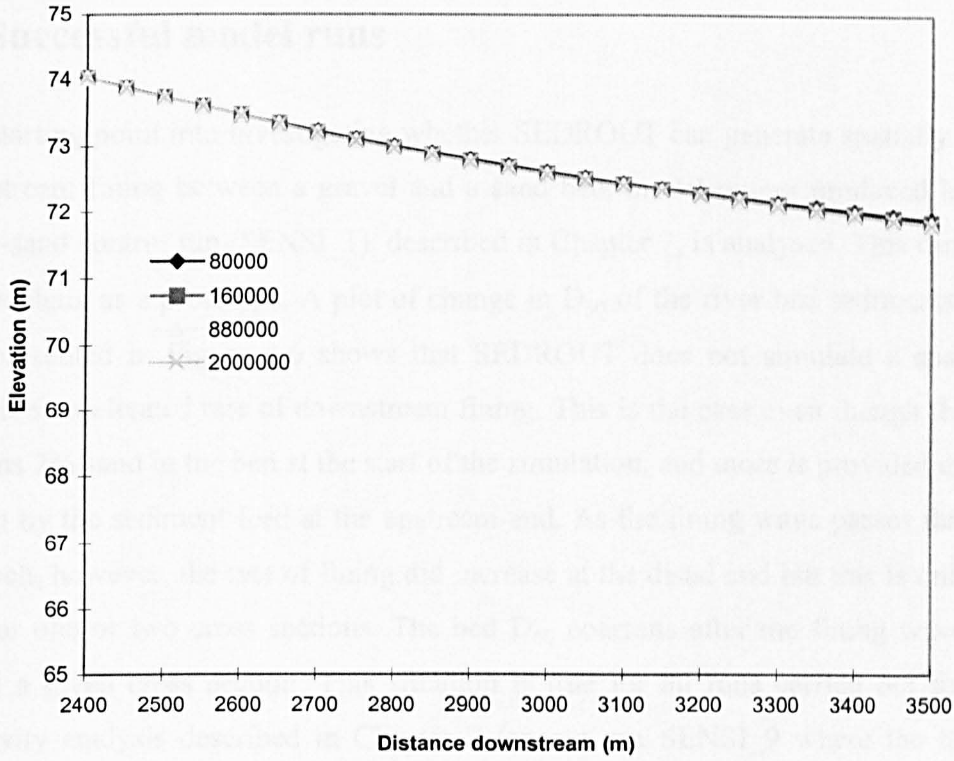


Figure 8.5: Changing bed slope over time from run BLT_13. The numbers in the legend are in minutes of model time.

An important finding of this research is, therefore, that a balance must be found between the proportion of fines specified in the initial bed GSD and the discharge, so that rapid headward erosion does not occur.

8.3 Successful model runs

As a starting point into investigating whether SEDROUT can generate spatially rapid downstream fining between a gravel and a sand bed, model output produced by the gravel-sand control run (SENSI_1), described in Chapter 7, is analysed. This run uses Allt Dubhaig as a prototype. A plot of change in D_{50} of the river bed sediments over time presented in Figure 8.6 shows that SEDROUT does not simulate a spatially extensive accelerated rate of downstream fining. This is the case even though the run contains 2% sand in the bed at the start of the simulation, and more is provided during the run by the sediment feed at the upstream end. As the fining wave passes through the reach, however, the rate of fining did increase at the distal end but this is only the case for one or two cross sections. The bed D_{50} coarsens after the fining wave has passed a given cross section. This situation is true for all runs carried out for the sensitivity analysis described in Chapter 7 (except run SENSI_9 where the hiding factor for gravel and sand was set to 0 and therefore all sizes were equally mobile). To generate what could be interpreted as a realistic GST it is necessary to increase the spatial extent of the fining wave during the run so it crosses several neighbouring sections or nodes at once. Simulation results such as these may manifest themselves as an accelerated rate of downstream fining, or a GST, in the field.

As SENSI_1 failed to generate a GST it is important to investigate which run conditions are necessary to allow SEDROUT to simulate an accelerated rate of downstream fining between the gravel and the sand reaches. Factors that can be varied to generate a GST include alternative initial or boundary conditions, parameter values, or changes to the parameter values while the model is running. The initial simulations concentrate on Allt Dubhaig. The methods are then applied to Vedder River, building on the work of Ferguson and Church (unpublished), to investigate differences in the model predictions for the two rivers, related to the specified initial conditions.

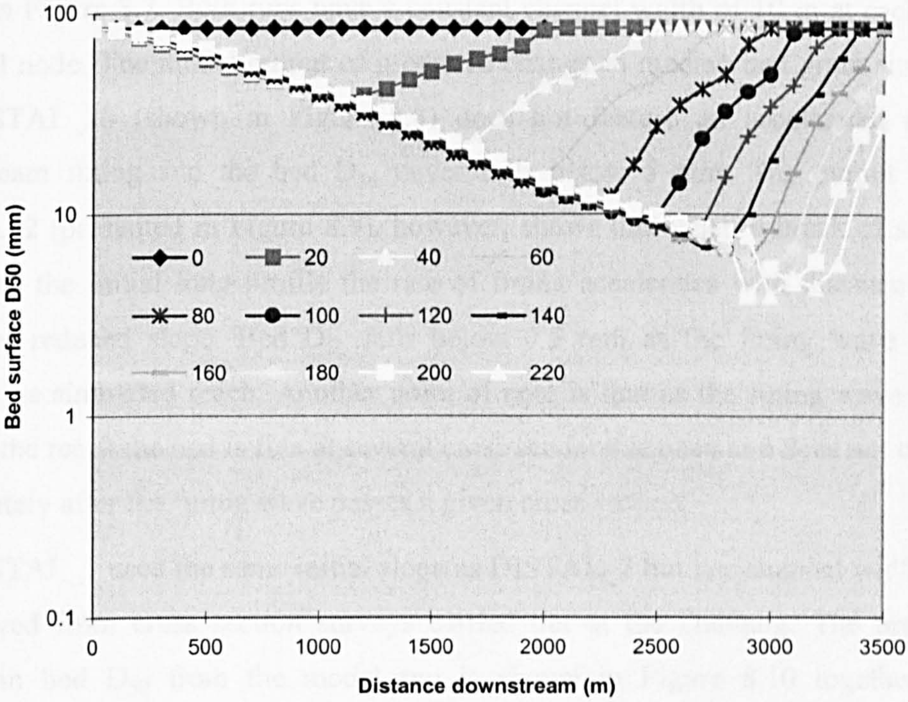


Figure 8.6: Change in bed surface D_{50} over time with distance downstream from the gravel-sand control run (SENSI_1). The numbers in the legend are x10^3 minutes of model time.

8.3.1 Variations in initial long profile

Two simulations were carried out to compare the impact of altering initial bed slope specifications on model output. Runs DISTAL_20 and DISTAL_2 have a smooth concave long profile, and a long profile with a break of slope (at 2800 m) respectively. All other parameters are identical in the runs. Plots of the initial long profiles used are shown in Figure 8.7. Both runs have a constant channel width of 10 m at each cross sectional node. The model output of predicted change in median bed grain size from run DISTAL_20 (shown in Figure 8.8) does not feature an accelerated rate of downstream fining and the bed D_{50} never falls below 5 mm. The output of run DISTAL_2 (presented in Figure 8.9), however, shows that when a break of slope is present in the initial long profile the rate of fining accelerates with distance in the region of reduced slope. Bed D_{50} falls below 0.5 mm as the fining wave moves through the simulated reach. Another point of note is that as the fining wave passes through the reach the bed is fine at several cross sections at once and does not coarsen immediately after the fining wave passes a given cross section.

Run DISTAL_1 used the same initial slope as DISTAL_2 but has channel widths that are derived from cross section surveys carried out at the Dubhaig. The predicted change in bed D_{50} from the model run is shown in Figure 8.10 together with annotations related to the channel width at each cross sectional node. The rate of fining is stronger in DISTAL_1 than DISTAL_2 due to the narrow channel widths downstream of the break of slope acting as a choke impeding the transport of the coarse fractions.

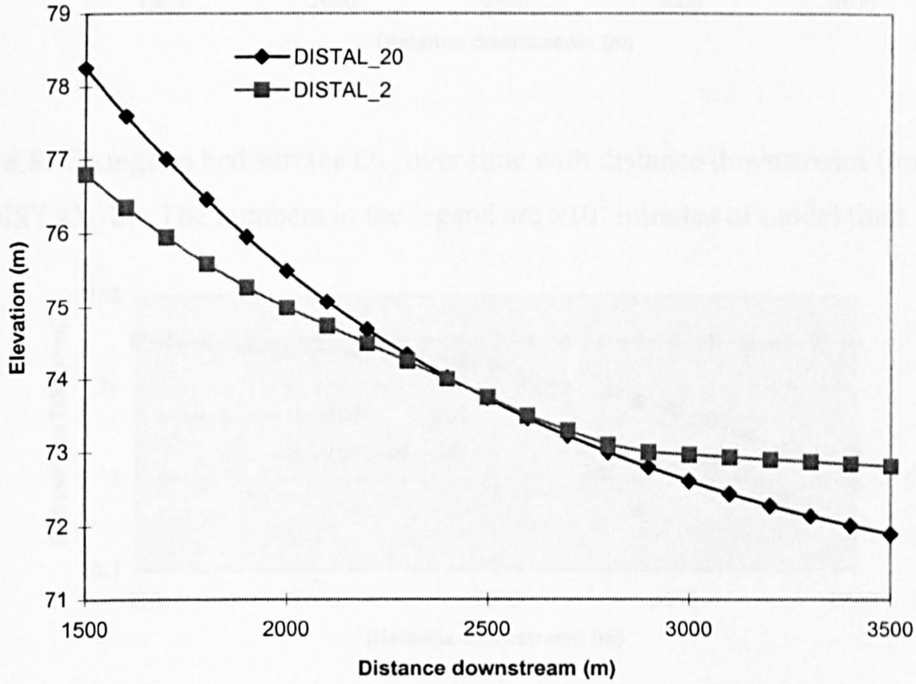


Figure 8.7: Initial bed surface slope specified in runs DISTAL_20 and DISTAL_2.

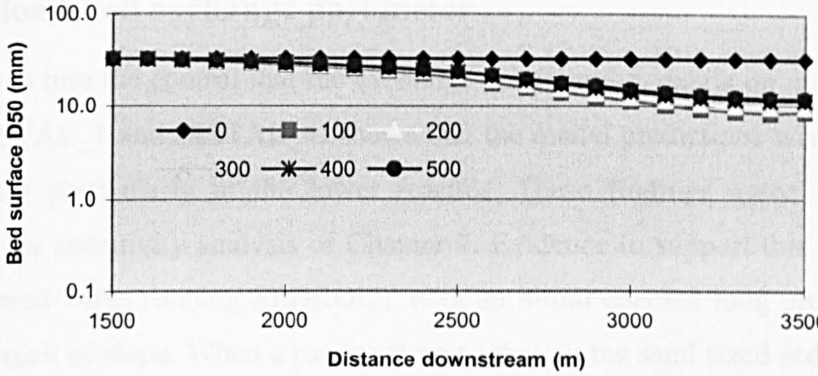


Figure 8.8: Change in bed surface D_{50} over time with distance downstream from run DISTAL_20. The numbers in the legend are x10^3 minutes of model time.

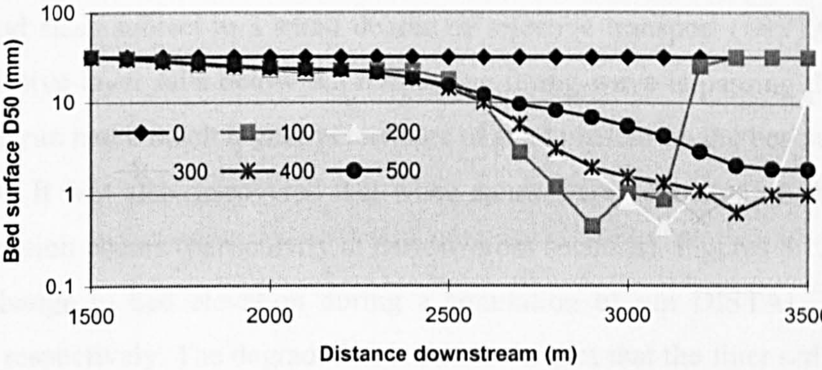


Figure 8.9: Change in bed surface D_{50} over time with distance downstream from run DISTAL_2. The numbers in the legend are x10^3 minutes of model time.

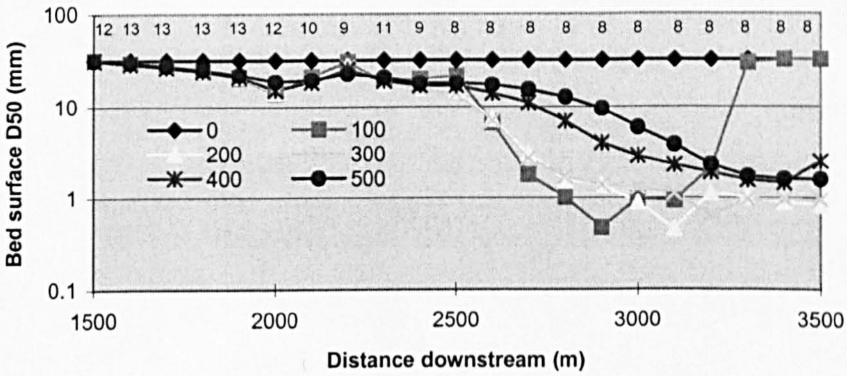


Figure 8.10: Change in bed surface D_{50} over time with distance downstream from run DISTAL_1. The numbers in the legend are x10^3 minutes of model time. Channel widths at each node are shown (in meters).

8.3.2 Bedload-bed exchange parameter

Investigations into the control that the exchange parameter, c, exerts on model output (in runs DISTAL_1 and DISTAL_4) shows that the model predictions were sensitive to this factor particularly in the lower reaches. These findings agree with those reported in the sensitivity analysis of Chapter 7. Evidence to support this hypothesis was discovered when running SEDROUT with an initial channel long profile which featured a break of slope. When a run is set up so that all the sand sized sediments are deposited in the subsurface (c = 0) and are at equal mobility (DISTAL_4), the D_{50} of the active layer does not fall below 2 mm at any time during the run, even when the fine wave is prograding through the reach (Figure 8.11). When the exchange parameter is set so that all the sands are deposited on the surface (c = 1), however, with the sand sizes subject to a small degree of selective transport (DISTAL_1), the D_{50} of the active layer falls below 0.5 mm as the fining wave is passing through the system. This run had a much higher percentage of sand present on the bed surface (see Figure 8.10). It was also discovered that when sands were deposited on the surface, more degradation occurs (particularly at narrow cross sections). Figures 8.12 and 8.13 show the change in bed elevation during a simulation of run DISTAL_1 and run DISTAL_4 respectively. The degradation is due to the fact that the finer sediments are easier to entrain than the coarser gravel sizes that dominate the surface if the exchange parameter is set such that all sand is deposited in the subsurface.

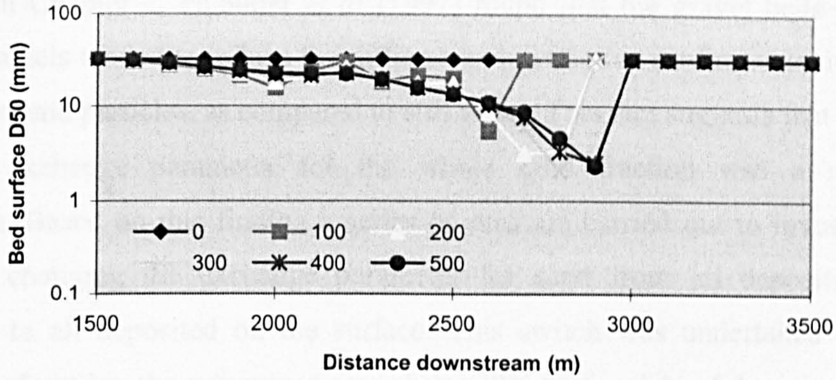


Figure 8.11: Change in bed surface D_{50} over time with distance downstream from run DISTAL_4. The numbers in the legend are $x10^3$ minutes of model time.

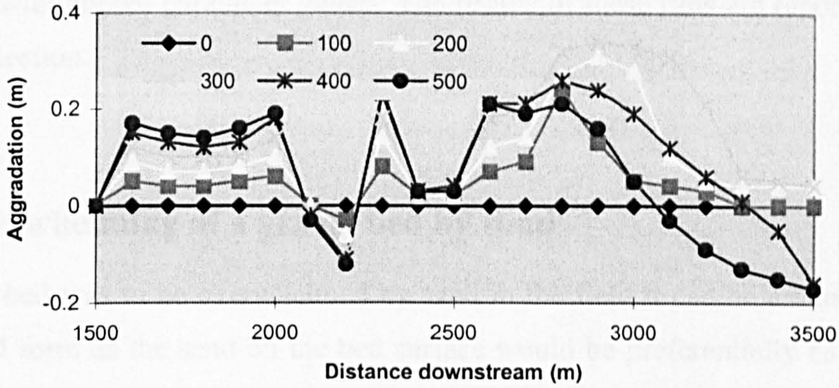


Figure 8.12: Change in bed elevation over time with distance downstream from run DISTAL_1. The numbers in the legend are $x10^3$ minutes of model time.

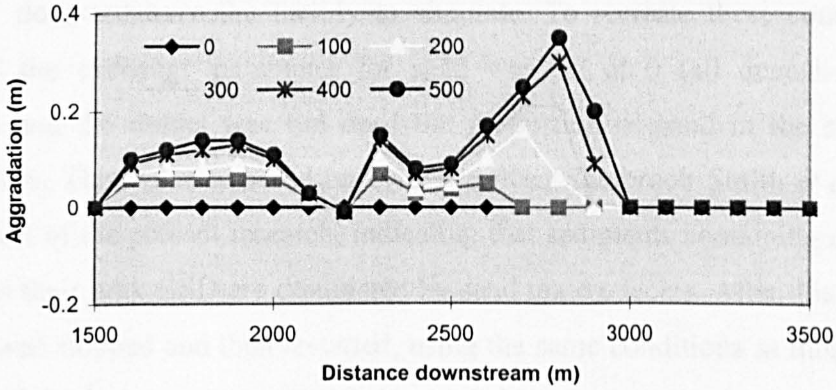


Figure 8.13: Change in bed elevation over time with distance downstream from run DISTAL_4. The numbers in the legend are $x10^3$ minutes of model time.

As noted in Chapter 2, Peloutier *et al* (1997) found that the gravel beds present in natural channels tended to reduce the differences in effective settling velocity of finer and coarser sand particles, as compared to still water. This fact suggests that the use of the same exchange parameter for the whole sand fraction was a reasonable assumption. Based on this finding a series of runs are carried out to investigate the impact of changing the exchange parameter for sand from all deposited in the subsurface to all deposited on the surface. This switch was undertaken when the proportion of sand in the subsurface rose above 30% to simulate of the overwhelming of a gravel bed by sand. SEDROUT allows the possibility of carrying out a simulation up until a particular point is reached by creating final slope and grain size files for the simulated reach at the end of each model run. A new run can then be started, using these files with altered parameter values. The results of these runs are reported in the following section.

8.3.3 Overwhelming of a gravel bed by sand

If a gravel bed was to be overwhelmed by sand in the field it can be assumed that a GST would form as the sand on the bed surface would be preferentially entrained at lower shear stresses than the gravel sediments. This would lead to a rapid increase in the flux of sand in transport with respect to gravel, as these finer sediments would no longer be hidden beneath the coarser and possibly armoured bed (Cui and Parker, 1998). The overwhelming of a gravel framework bed by sand is a situation which SEDROUT does not have the facility to simulate. To recreate these conditions in SEDROUT the exchange parameter for sand was set at 0 (all deposited in the subsurface) and the model was run until the proportion of sand in the subsurface exceeded 30%. This value is based on evidence from Sambrook Smith *et al* (1997), and Chapter 4 of the present research, indicating that sediments containing more than 30% sand in their bulk GSD are dominated by sand matrix facies. After this point the simulation was stopped and then restarted, using the same conditions as those present at the end of the last run, but with all the sand deposited on the surface (exchange parameter set at 1).

While carrying out these simulations it is discovered that the model output can be highly influenced by the GSD of the feed material that is supplied at the upstream end of the simulated reach. In the runs carried out for the sensitivity analysis (Chapter 7) the feed was calculated automatically so that no aggradation or degradation was allowed at $x = 0$ (the upstream limit). The sediment influx and GSD were calculated by SEDROUT to exactly match the transport capacity for each grain size at the top of the reach. During the experiment carried out here, however, while the condition of no aggradation or degradation is held, a GSD is specified so that the proportion of sand in the subsurface reaches the required value (30%). In all the runs the discharge is kept constant at 5 m^3/s, the initial bed had a D_{50} of 31 mm and contained 6% sand.

A first run was carried out with a fixed feed that had a GSD equal to that calculated from sediment deposited in Allt Dubhaig bedload trap after 31 combined flood events between 1991 and 1993 (run DISTAL_6). After running the model it was found that the feed contained too much sand (37.21% of the total sediment) and the required 30% of sand in the subsurface was reached by the second data dump. In an attempt to increase the amount of time taken for the proportion of sand in the bed to reach 30% a reduction in the amount of sand in the feed was required. By trail and error it was discovered that when the feed was specified as having 28% sand the 30% proportion in the subsurface was reached part-way through a run (DISTAL_13). Run DISTAL_13 was then renamed DISTAL_14 and run until the time that the proportion of sand in the subsurface first exceeded 30%. Run DISTAL_15 used the files created at the end of DISTAL_14 as its initial conditions and the bedload-bed exchange parameter was changed from 0 to 1 for the sand sizes. The proportion of sand in the subsurface rose above 30% at the cross section where the break of slope occurred in the initial long profile. In runs carried out on a smooth concave long profile the proportion of sand in the bed does not exceed 30% (DISTAL_16).

The results of these model runs show that as the fining wave is prograding through the reach the sections remain fine for a number of output dumps. These characteristics are unlike the sensitivity analysis runs, presented in the previous chapter, where coarsening of a node occurred immediately after the fining wave had passed. Rather, the sections where the proportion of sand in the subsurface is high, remain fine after the fining wave has passed through. Figure 8.14 shows the simulated downstream fining profile from the output of runs DISTAL_14 and DISTAL_15 combined.

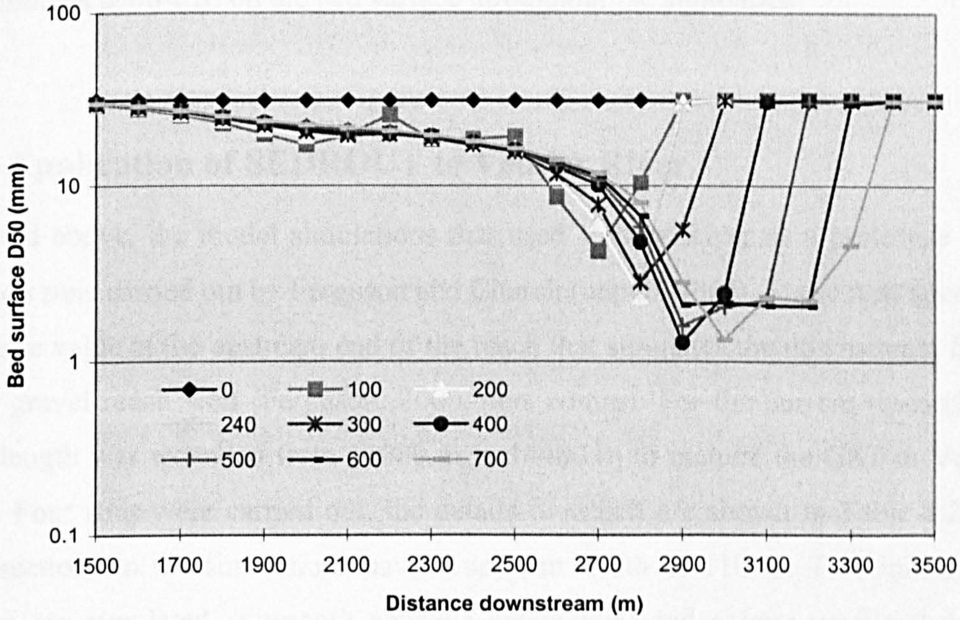


Figure 8.14: Change in bed surface D_{50} over time with distance downstream from runs DISTAL_14 and DISTAL_15 combined. The numbers in the legend are x10^3 minutes of model time.

The rate of fining also accelerates towards the cross section with the highest proportion of sand in the subsurface. These findings again indicate that for SEDROUT to generate an abrupt and spatially extensive GST a break of slope is required. The predicted GST generated by altering the exchange parameter is, however, no better than that generated by carrying out a run with a break of slope and all sediments deposited on the bed surface throughout the simulation.

8.3.4 Application of SEDROUT to Vedder River

As noted above, the model simulations that used Vedder River as a prototype were based on runs carried out by Ferguson and Church (unpublished). These runs specify a discharge value at the upstream end of the reach that simulates the downstream fining of the gravel reach well (Ferguson, 2000, pers comm). For the current research the reach length was extended from 11000 m to 14400 m to include the GST in Vedder Canal. Four runs were carried out, the details of which are shown in Table 8.2. All cross sections in the simulations have a uniform width of 110 m. Two initial long profiles are simulated, a smooth concave exponential and a long profile that was concave to 8800 m and was straight downstream of this distance. Plots of the initial long profiles are presented in Figure 8.15. Runs are also undertaken using two different initial bed grain size specifications: a constant coarse GSD (D_{50} = 35 mm) throughout the reach; and, the same coarse bed specified to 8400 m with finer grained sediments (D_{50} = 1 mm) downstream. Runs are carried out using the profile with a break of slope and break in grain size to compare the importance of these.

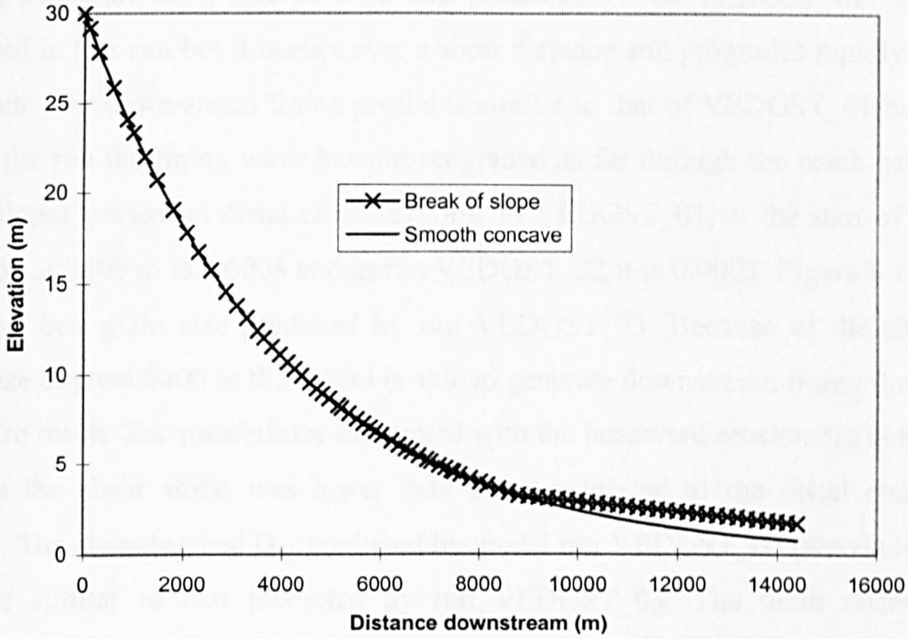


Figure 8.15: Initial bed surface slope specified in runs using Vedder River as a prototype.

Run VEDGST_01 (smooth concave initial slope and constant grain size) shows no accelerated or spatially extensive downstream fining. A plot of the change in bed surface D_{50} over time is shown in Figure 8.16. Because of the low slope in the distal part of the simulated reach the coarse initial bed GSD is present at the end of the run. The influence that the break of slope has on the predicted median bed grain size can be seen in Figure 8.17, derived from data produced by run VEDGST_02. A GST is produced in this run but it occurs over a short distance and progrades rapidly through the reach. The downstream fining profile is similar to that of VEDGST_01 but by the end of the run the fining wave has not prograded as far through the reach because of lower slopes present at distal cross sections. In VEDGST_01, at the start of the run, the slope at 8800 m is 0.0008 and in run VEDGST_02 it is 0.0003. Figure 8.18 shows changing bed grain size predicted by run VEDGST_03. Because of the change in grain size beyond 8400 m the model is able to generate downstream fining throughout the entire reach. The instabilities associated with the headward erosion did not reoccur because the shear stress was lower than those exhibited at the distal end of run BLT_2. The changing bed D_{50} predicted by model run VEDGST_04 (shown in Figure 8.19) is similar to that predicted by run VEDGST_03. The main difference is associated with the rate of bed coarsening in the lower reaches of the simulated stream as the model tended towards equilibrium. Because of the lower bed slope, and therefore shear stress, in run VEDGST_04 this coarsening occurred more slowly than in run VEDGST_03.

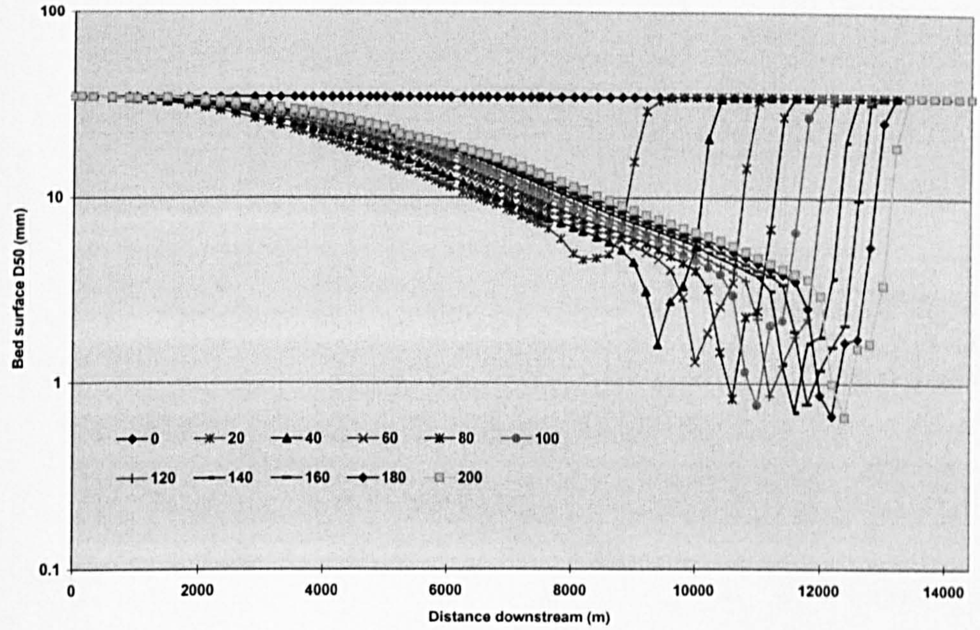


Figure 8.16: Change in bed surface D_{50} over time with distance downstream from run VEDGST_01. The numbers in the legend are x10^3 minutes of model time.

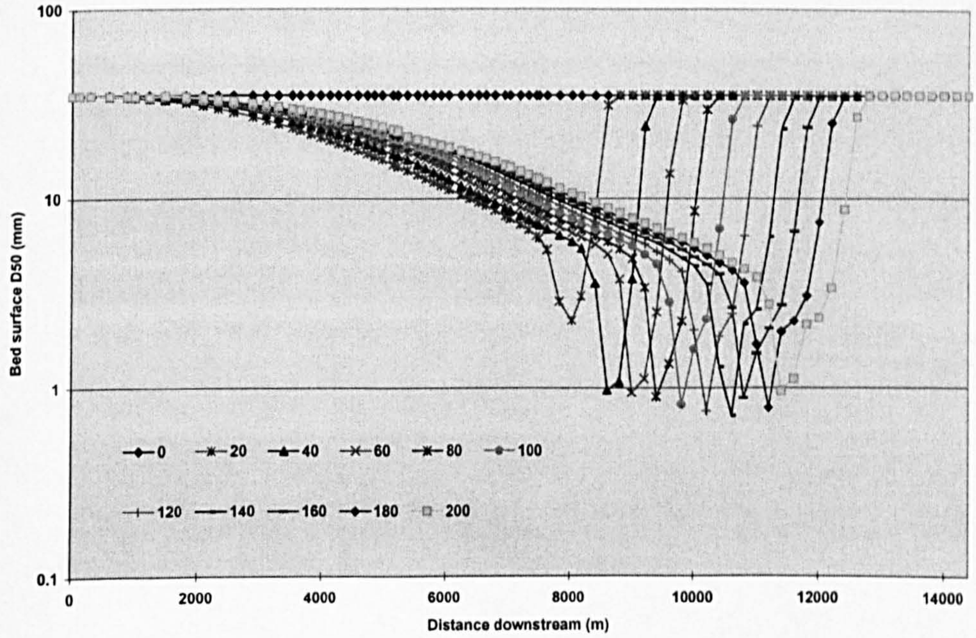


Figure 8.17: Change in bed surface D_{50} over time with distance downstream from run VEDGST_02. The numbers in the legend are x10^3 minutes of model time.

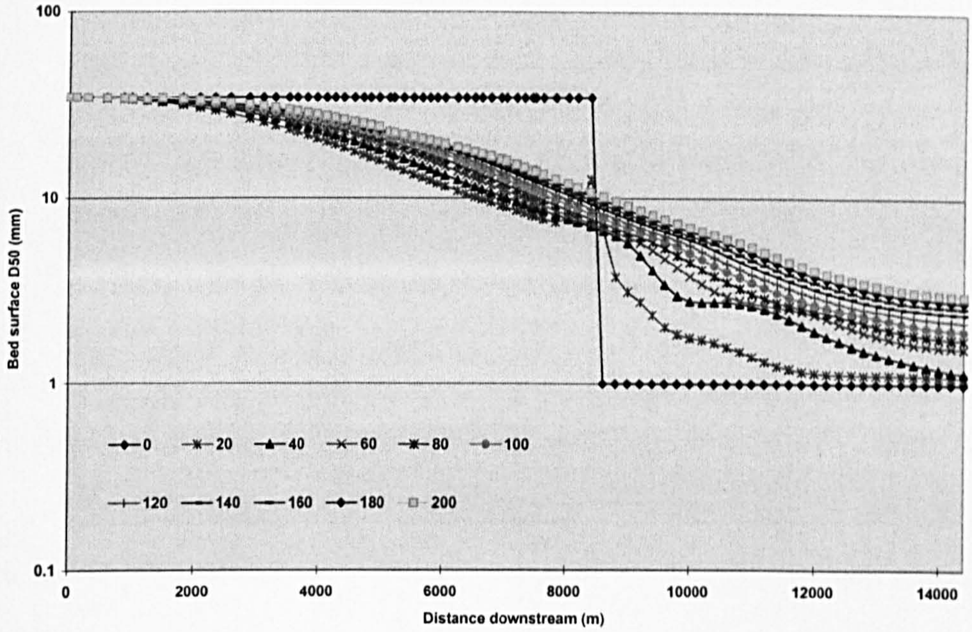


Figure 8.18: Change in bed surface D_{50} over time with distance downstream from run VEDGST_03. The numbers in the legend are x10^3 minutes of model time.

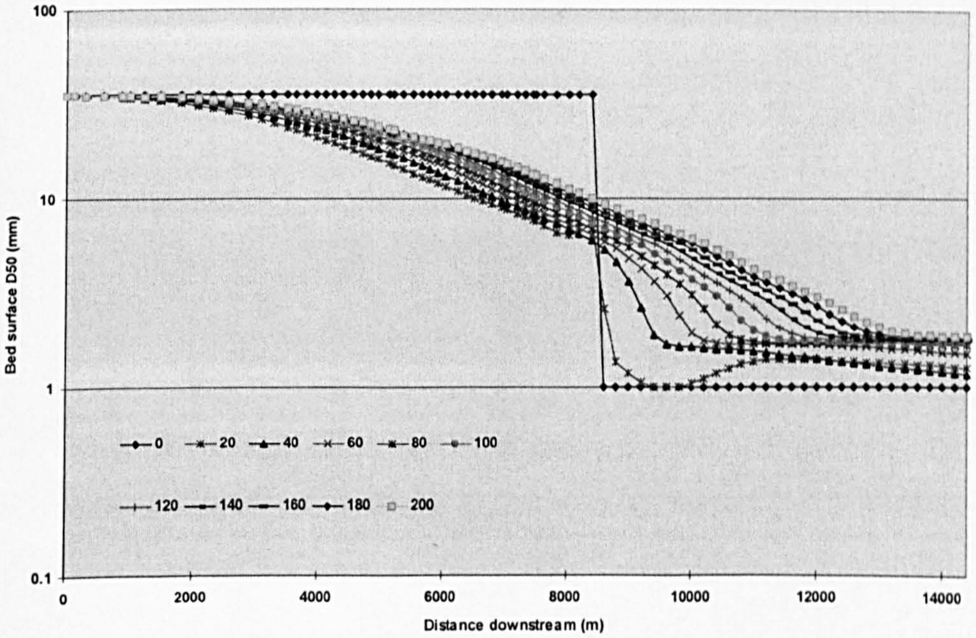


Figure 8.19: Change in bed surface D_{50} over time with distance downstream from run VEDGST_04. The numbers in the legend are x10^3 minutes of model time.

8.4 Unsuccessful Model Runs

On several occasions during the modelling investigations SEDROUT crashed, either at the start of or during a run. Many of these failures are associated with using data collected in the field as initial conditions for a model run. There are two possible causes for these failures: 1) SEDROUT was unable to cope with some of the initial, boundary or parameter conditions which are witnessed in the field and therefore is missing a process that may be important to GST development, or; 2) for certain conditions the coding of SEDROUT is inadequate for carrying out successful simulations. In many situations a combination of the causes may have been responsible for model crashes.

In almost all cases the hydraulics subroutine of SEDROUT was the cause of the failures. The model was unable to calculate the water surface slope (based on the specified bed slope and bed surface GSD) for many of the unsuccessful runs.

8.4.1 Instability caused by the exchange parameter

In runs DOGLEG_1 and DOGLEG_2 it was discovered that SEDROUT is sensitive to the exchange parameter value. In these runs the model crashed when the exchange parameter was set at 1 for sand (sediment deposited on the surface) but ran successfully when set at 0 (subsurface deposition) for these sizes. This difference between the two runs manifests itself as different proportions of sand available for transport on the bed surface. Both runs use the modified Parker (1990) bedload function, indicating that in some situations, where large proportions of sand are present in the active layer (as when the exchange parameter is set to 1 in DOGLEG_2), the enhanced bedload function is still unstable.

8.4.2 Simulations using real field data

To test the accuracy of SEDROUT at predicting change over several years, grain size and survey data collected in Allt Dubhaig in 1992 (by Sambrook Smith) is used as the initial conditions of a model run (BULK_1A). The aim is to compare the data collected for the present research with the output of the model run that uses the 1992 field data to specify its initial conditions. In all cases, however, when using real field GSD and survey information as the initial run conditions of a model run, SEDROUT crashes. In Allt Dubhaig the break of slope that is associated with accelerated downstream fining and the GST (detailed in Chapter 4) is of the same order of magnitude as that used in the DISTAL series of model runs. As noted in Chapter 4, in the prototype the slope changes from 0.002 to 0.0002 and the slope change in the DISTAL series was 0.003 to 0.0003. This change in gradient cannot, therefore, be the sole cause of the instability in the hydraulics routine of SEDROUT. The rate of change of grain size of the bed material in the Dubhaig is also rapid and in the model this would again lead to an abrupt change in the calculated water depth and shear stress over a short distance. The initial grain size data specified at the start of the runs using real field data exhibits a decrease in bed D_{50} from 4.9 mm to 0.3 mm and an increase in the proportion of sand in the bed from 25 to 93% over the space of 9 m. This rate of change in bed surface grain size is extreme even in the GST reach of a river and may be an artefact of sampling strategy used in 1992. Discussion of sampling techniques in gravel-sand bed rivers in Chapter 9 supports this inference. Inserting new sections in the region of rapid bed GSD change, and also smoothing the rate of grain size change in run BULK_1B, does not allow successful runs to be completed. Run BULK_1D was then carried out using real bed GSDs but with a standard cross section width (10 m) and a smooth concave long profile from a run which had been successfully carried out previously (run BLT_1). This run was still unsuccessful and crashed before the first data dump. Successful model runs have, however, been carried out on the gravel and sand data from both upstream and downstream of the rapid change in grain size. From these findings it can be assumed that the rapidly changing grain sizes is causing the failures, rather than the specified slope. A run using a straight long profile (BULK_1C) also crashed. Unfortunately, these discoveries did not allow model runs to be carried out using realistic field data as the initial run conditions. Field data collected for this thesis, and by others, cannot, therefore, be used as a test for model accuracy.

A similar difficulty is encountered when attempting to run SEDROUT using real field data collected from Vedder River. In this stream the bed GSDs fluctuated downstream in the distal gravel reach. The size fractions (in mm) in which the D_{84} occurs varied downstream as follows: 45-64, 32-45, 23-32, 32-45, 8-11, 23-32, 2.8-4, 0.7-1. Downstream fining, therefore, does not occur smoothly as is the assumed tendency in most streams with no lateral inputs. As the D_{84} is used by the model to calculate the water surface slope, some negative gradients were simulated between adjacent nodes. In the field these variations are likely to be smoothed out by changing grain sizes between the sections and also across the channel width. The grain size data being used to specify the initial conditions of run VED_1 was from point samples and any spatial variation in grain size at a cross section would therefore not be characterised.

8.5 Summary

The model simulations carried out for this research have been a partial success. The results are qualitatively realistic but quantitatively inaccurate. By introducing a break of slope into simulations it has been shown that an accelerated rate of downstream fining can be simulated. This rapid fining is not, however, of the same strength, or in the same location, as the GST found in Allt Dubhaig, the prototype upon which the model runs were based. This failure to recreate the conditions witnessed in the field could be due to a number of factors. Firstly, there may be an important process missing from model at present. Secondly, some processes may be simulated incorrectly by model. A third possibility is that the field data being used to run and test the model may not have been correctly collected to fulfil this purpose. As the model takes an average GSD at each node, the data used to test it must also be averaged across each channel cross section. Finally, the initial conditions used at the start of a model run may be unrealistic. These start-up conditions can impact on the simulated results, particularly for short runs.

Another major finding of the present chapter is that SEDROUT is not capable of simulating a realistic GST due to the rate of change of bed surface GSD occurring in the field. This fact means that the model could never recreate a realistic GST and therefore the evolution of the GST cannot be simulated. Investigations undertaken in this chapter showed that the specified bed surface GSD is the most likely cause of the model failures.

The differences between Allt Dubhaig and Vedder River are associated with the amount of time taken for the fine wave to pass through the simulated reach. This takes considerably longer for the Vedder and is associated with the low slope at the distal end of the reach. The most spatially rapid and extensive GST was generated by specifying a break of slope in the initial run conditions. Changing the exchange parameter when the proportion of sand in the subsurface rose above 30% caused the bed to coarsen before the fining wave had passed through the simulated reach.

Chapter 9. Discussion: forms and processes generating a GST

This chapter draws together the main findings of the thesis by linking the field and modelling aspects. By analysing both strands of investigation together inferences regarding the characteristics and processes occurring in a GST can be drawn. The field studies highlight areas where the one-dimensional model (SEDROUT) may be oversimplifying the natural conditions, limiting its accuracy and reliability, together with further improving the conceptual model of GST initiation and development.

As noted previously, numerical models are, in all cases, a simplification of the natural system which they are simulating. The degree of simplification can greatly impact upon the accuracy (correct results) and reliability (for the right reasons) of the model predictions. For this reason a clear knowledge of the structure of the chosen numerical model and the reason for its employment is vital.

During the course of the investigation it was therefore important to remember why SEDROUT was being used. The current research is aiming to elucidate the forms and processes important generating a GST. The model assists in this aim because it simulates width-averaged size-selective bedload transport which has been postulated as one of the main controlling processes. If the model does not generate a GST then it can be inferred that some important factors are missing. The detailed field characterisation of two GSTs, in Allt Dubhaig and Vedder Canal, indicates what these factors may be by providing a holistic view of a river in the region of a GST. The model is not being used to simulate the GST itself and therefore should not be seen as a predictive tool.

The chapter begins with a discussion regarding the need for, and extent to which, an internal validation of SEDROUT is necessary. This is followed by an assessment of the limitations of SEDROUT highlighted by the field research into the characteristics of rivers exhibiting a GST. Implications of this field research are also discussed, along with directions for further study, where potentially useful developments of SEDROUT to assist in the modelling of gravel-sand mixtures are outlined.

9.1 Internal validation of SEDROUT

The aim of the modelling aspect of the current research was to investigate the importance, in generating a GST, of width-averaged size selective bedload transport, rather than to produce an accurate numerical model for the behaviour of gravel-sand sediment mixtures. For this reason a rigorous internal validation of SEDROUT is not required. The fact that many aspects of the natural system are not included in the model, and the potential inaccuracy of specified initial run conditions, combine to further suggest that an internal validation is unnecessary here and may be potentially misleading. It is not possible to assess the accuracy of SEDROUT in predicting changing bed surface GSDs in the transition zone over time because the model did not run successfully using data collected in the field as initial conditions (see Section 8.4.2 of Chapter 8).

Parts of the field data collected for the current research, however, do support some of the model predictions, indicating that, although SEDROUT did not predict a GST like those present in the field, using realistic initial conditions, the internal workings of the model are at least similar to some of the processes occurring in the natural system.

The progradation of the gravel front exhibited by all model runs, presented in Chapters 7 and 8, is supported by field data of channel bed sedimentology change over time in Allt Dubhaig and Vedder Canal (see Chapter 5). Although the presence of accelerated and spatially rapid downstream fining during model runs was associated with a break of slope specified at the start of the model run, this fact does not imply that the conditions of the model run are completely unrealistic. This is because it is unlikely that any gravel bed river will flow down a smoothly concave profile for its entire evolution. The initial conditions for a run are, however, artificial with the same coarse bed GSD specified throughout the reach. For this reason the results should only be interpreted part way through each run to allow smoothing of the artificial initial grain size conditions. It should be noted that it is not necessary to wait until the model has reached equilibrium before analysing its predictions since many rivers exhibiting GSTs are not in equilibrium themselves. Evidence presented in the current research regarding the progradation of both the gravel front and gravel patches in the united gravel-sand reach suggest that the GST is likely to be a transitory feature at geological timescales. This finding calls into question the assumptions made by Parker and Cui (1998) and Cui and Parker (1998), during the development of their numerical model, when the GST was though of as a non-migrating feature over long timescales.

Probing investigations at Allt Dubhaig, presented in Chapter 5, indicate a progradation rate for the gravel of 0.8 m/year. Between 1992 and 1997 this would equate to 4 m movement of the gravel front, although its migration each year is dependant on the frequency of floods capable of entraining gravel over this period. The surficial gravel patches in the transition zone increased in length by between 11 and 15 m between 1992 and 1997. Although this rate of progradation is greater than that experienced by the gravel front itself this is accounted for by the fact that the gravel patches are found in the most active part of the channel. The patches are therefore mobile for a greater time period, increasing the transport rate of grains in these coarser patches. The rate of movement of the upstream fine gravel tracers is also likely to be greater than the progradation of the gravel front because these were all seeded on the bed surface and were therefore available for transport by all floods. Their virtual velocity would only decrease as they became integrated into the active layer. Fewer of the larger sized upstream tracers were buried (33 % of 20 mm tracers compared with 65 % and 63 % of the 10 and 14 mm tracers respectively) indicating that burial took place on a size selective basis. These findings agree with those of Wilcock (1997). A higher proportion of the downstream tracers were buried (between 85 and 98%). These were surveyed four months after the upstream tracers and the higher number of buried tracers may be due to greater vertical mixing between the surveys (Hassan and Church, 1994).

In an attempt to assess the rate of gravel progradation predicted by SEDROUT, the output of the runs investigating overwhelming of a gravel bed by sand (DISTAL_14 and DISTAL_15 in Chapter 8, Section 8.3.3) was interrogated. The rate of coarsening experienced in these runs after the fining wave had passed 2800 m, where a break of slope was present at the start of the run, was analysed as a surrogate for the coarsening experienced in the field as the gravel front progrades over a sandy bed. The time taken for the bed D_{50} at 2800 m to coarsen to a value similar to that present at 2700m (7.0 mm) at the first output dump (80000 minutes) of run DISTAL_15 was recorded. The D_{50} at 80000 minutes was close to the value exhibited by the gravel front in the field (between 9.6 and 6.2 mm at cross sections 5 and 10 respectively in 1997, see Chapter 4). After 1440000 minutes the D_{50} at 2800 m was 6.9 mm. This progradation of the coarse front therefore took 1360000 minutes of model time. The model runs were carried out with a specified discharge of 5 m^3. Hoey and Ferguson (1994) suggest that a run with this discharge and a duration of 500000 minutes is equivalent to 1 year of constant high flow or 100 years in real time. A run of 1360000 minutes therefore equates to approximately 2.7 years of constant high flow or 270 years of real time. From this evidence it took the simulated gravel front 270 years to prograde 100 m with an average progradation rate of 0.4 m/year. Although these calculations are crude, and the simulated strength of fining with distance is too low, they do indicate that the rate of progradation of the gravel front predicted by SEDROUT is of the same order of magnitude to that witnessed in the field.

9.2 Limitations of SEDROUT for the simulation of gravel-sand mixtures

The findings of this thesis show that the simplifications made during the construction of SEDROUT make the model unsuitable for simulating rivers with gravel-sand mixtures present on the bed. The importance of characteristics exhibited by GSTs in the field, but not incorporated in the model, requires assessment. Some aspects of natural GSTs which are not currently included in SEDROUT are: the overwhelming of a gravel-bed by sand and associated changes in bed surface sedimentology and surficial sand; the influence of sand on near bed hydraulics; downstream variations in grain size associated with bar / inter-bar storage; lateral sorting of sediment into patches of different grain size; and the dropout of sand from suspension.

9.2.1 Bed surface sedimentology and surficial sand

A crucial control on GST location is the relationship between the increasing bulk sand content in a gravel bed and the bed surface sedimentology. The capacity of a gravel bed to hold interstitial sand in its framework therefore requires quantification. This is of importance as once the gravel bed is filled with sand the finer sediments become available for transport at lower stresses, hence increasing the partial transport rates at shear stresses below those required to initiate equal mobility (Wilcock and McArdell, 1993; Lisle, 1995). The fine sands are therefore available for sorting into discrete lateral patches as they are in transport more often than gravel (Diplas and Parker, 1992; Paola and Seal, 1995), and therefore impact upon the near-bed hydraulics (Wilcock, 1993; Sambrook Smith *et al*, 1997). These two issues are discussed further in the following sections.

The detailed sedimentological investigation of Vedder Canal based upon both bulk grain size and surface sedimentology can be used to analyse the changes in bed surface sedimentology with increasing bulk sand content. This information can also be used to assess whether a quantitative assessment of the percentage range of sand in the bed surface can be obtained by a qualitative observation of the bed sedimentology. The proportion of sand required to cause a switch from a gravel framework to a sand matrix will also become clear following this analysis. The scheme used (adapted from Kodama, 1994b, based on visual observation) to classify the bed surface sedimentology of the Vedder River was detailed in Chapter 4 (see Section 4.2.3 and Table 4.2 for further details). Table 9.1 presents a comparison of the visual observation scheme with the bulk proportion of sand in the sample obtained by sieving.

As the facies classes 3M, 4 and 5 are made up of a sand matrix at the bed surface it can be assumed that sediments of this type will have sand available for transport at lower shear stresses than sand present in a gravel framework bed. The fines present in the matrix-supported beds are therefore available for entrainment at shear stresses below the equal mobility threshold. The 14% bulk sand content present in the Type 2 facies at N1.5, therefore, is not available for transport until the gravel in which these fines are stored is removed.

For a width-integrated sample containing 23% bulk sand (N4), only 6% of the channel width is made up of a matrix-supported sediment (3M). With a small further increase in bulk sand, to 30% (N6), the proportion of the bed surface across the channel width formed by a sand matrix facies jumps to 42%. From this evidence it can be seen that with only a 7% increase in the width-averaged bulk proportion of sand there is a 36% increase in the areal extent of sand matrix-supported facies. This results in a much greater availability of sand for entrainment from the bed surface.

Sampling within individual patches of different ambient grain size shows that only 36% bulk sand is necessary to form an entirely sand-matrix facies (Type 3M at N11, 10.2-16.6). When the proportion of bulk sand is increased to 41% (N10, 7.2-17.9) the bed surface appears to be almost entirely sand-dominated (Type 4).

Table 9.1: Proportions of the bed surface occupied by different sedimentologies in Vedder Canal, at each individual facies and width-averaged sample analysed, compared to the bulk sand percentage. Distances across the channel for the facies samples are in m. Note: sedimentology of N1.5 was not recorded. Information in the table is derived by merging the recorded sedimentology from N1 and N2.

Cross Section	Kodama Value (% of cross section covered)					bulk % sand
-	2	3F	3M	4	5	-
N1.5 (Width)	30	70				14
N4 (Width)		94	6			23
N5 (Width)		59		36	5	42
N6 (Width)		58	42			30
N7 (Width)		44	30	26		49
N8 (Width)		67	27	6		32
N9 (14.3-65)			100			50
N9 (70-74)					100	81
N9 (Width)		3	54	21	22	52
N10 (7.2-17.9)				100		41
N10 (23.3-43)					100	85
N10 (48-86)				100		49
N10 (Width)			6	64	30	58
N11 (10.2-16.6)			100			36
N11 (23.2-49)				100		75
N11 (56-90)					100	81
N11 (width)	10		16	42	32	74
N12 (10.1-64)				100		87
N12 (71-87)				100		67
N12 (Width)	6			63	31	82

Table 9.2: Limits of the proportions of sand in the different facies classes in Vedder Canal (derived from Table 9.1). Notes: Some proportions are estimated from samples which consist of more than one facies class but are dominated by the class of interest. For a sample to be included it must contain only the Kodama class of interest and those immediately coarser or finer to avoid bias.

Kodama Value	Max % sand	Min % sand
2	14	N/A
3F	30	14
3M	36	30
4	75	41
5	87	81

Table 9.2 above shows that the sedimentology type changes as the bulk fraction of sand in a sample increases. There is also very little overlap between the proportion of bulk sand in a sample and its allocated sedimentological type. This finding indicates that, although the scheme for assessing bed surface sedimentology is qualitative, the results produced are a reliable indicator to the actual range of sand volumes present in the bed. As the notation scheme was tried by both the author of the current research and two colleagues, who agreed on the classifications, operator variance is low. It can be seen from Tables 9.1 and 9.2 that the switch from a framework-supported gravel bed to a matrix-supported sand bed occurs as the proportion of bulk sand increases from 23% (N4, 94% 3F) to 36% (N11, 10.2-16.6, 100% 3M). As Table 9.1 shows when the width-averaged bulk sand is 30% (N6) almost half of the channel width (42%) is sand-matrix bedded. This supports the finding, of Carling and Reader (1982), Sambrook Smith *et al* (1997) and Wilcock (1998), that the threshold proportion of sand that a gravel-framework bed can hold interstitially lies between 23 and 36%.

A similar, although less detailed, investigation of the relationship between the bulk and areal extent of sand matrix sediments can be carried out using data collected from

Allt Dubhaig. Sedimentological information gathered from this stream is presented in Table 9.3 below.

Table 9.3: Proportions of the bed surface occupied by different sedimentologies in Allt Dubhaig, at each individual facies and width-averaged sample analysed, compared to the bulk sand percentage. Distances across the channel for the facies samples are in m. LB and RB are left bank and right bank respectively.

Cross Section	Proportion of width occupied by sand-matrix facies (%)	Bulk % sand
1 (width)	0	7
5 (width)	0	8
10 (width)	34	43
10 (LB-7)	0	14
10 (7-RB)	100	99
14 (width)	80	72
14 (LB-5)	100	69
14 (5-6)	50	32
14 (6-RB)	100	90
17 (width)	45	61
17 (LB-5)	100	100
17 (5-8)	50	29
19 (width)	100	100
23 (width)	89	92
27 (width)	100	98

The results of the sedimentological study carried out in the GST reach of Allt Dubhaig reflect the findings from Vedder Canal. There is a threshold in the proportion of the bed surface occupied by sand matrix sediments as the bulk sand content rises. With bulk sand contents up to 14% (10, LB-7) the channel bed is entirely composed of gravel framework sediments. As the bulk sand content rises to 29 % (17, 5-8) sand matrix sediments occupy 50% of the channel bed. This figure rises to 100% of the channel bed as the bulk sand content reaches 69% (14, LB-5). Although the relationship between bulk and areal sand content is not as clear as that exhibited by Vedder Canal, due to the more rigorous facies notation scheme employed in the Canadian river, the overall results agree well. There is a non-linear relationship between bulk and areal sand content and until the threshold bulk sand content of between 20 and 30% is reached very little sand is present on the bed surface.

This threshold in the capacity of a gravel-framework bed to hold interstitial sand is not simulated by SEDROUT as the exchange parameter for sand sizes remains fixed during the runs. In the field, as the sand proportion in the bed subsurface increases, more sand is deposited on the bed surface. In the model, however, this overwhelming does not occur and, if specified at the start of a run, all sands continue to be deposited in the bed subsurface regardless of the bulk bed sand content. Model runs were specifically undertaken to address this shortcoming (see runs DISTAL_14 and DISTAL_15 in Chapter 8). These showed that a switch from subsurface to surface sand deposition during a run, as the subsurface sand content approached 30%, created a more stable, slowly prograding GST.

The near-bankfull high flow specified during all model runs also limits the development of a GST since almost all grain sizes can be entrained at this discharge. With lower flows occurring for the majority of the time in the field any surficial sand can be transported without the need for the removal of gravel. Partial transport therefore occurs with size selective entrainment of the finer fractions (Ikeda and Iseya, 1988; Wilcock and McArdell, 1993;1997). For this reason, to simulate the behaviour of gravel-sand mixtures realistically, variable hiding factors are required for each individual size fraction depending on its relative proportion within the bed grain size mix. When sands dominate the bed, these are removed, even at low shear stresses. This situation is likely to become crucial in the field in a reach where sand patches are beginning to form on the surface as the threshold of a gravel bed to hold interstitial sand is approached (Wilcock, 1998). Where no fines are present on the bed surface, at low stresses little sediment is entrained. Once patches begin to form, however, a threshold is crossed and this fine sediment is preferentially entrained compared to the gravel. These fines will greatly increase the flux of sand, providing more sand for deposition. Sand-rich sediments therefore begin to dominate downstream. This acts to further increase the mobility of sand with respect to gravel, reducing the importance of hiding for these finer size fractions (Ikeda and Iseya, 1987;1988). In effect, the size of the exponent x in equation 2.14 increases for sand, moving the sediment transport system towards simple Shields and away from near-equal mobility, by increasing the availability of the fine fraction. These inferences agree with the findings of Kuhnle (1993a,b). He observed during flume experiments that, for a bed composed of a bimodal gravel-sand mixture, parts of the bed surface were composed of 100% sand. The sand from these areas was entrained in a manner similar to that exhibited by a bed containing no gravel. An implication of strong selective transport of sand is therefore a sharper GST. If the proportion of sand in the bed were lowered then no sand would be transported at low transport rates because the fines would be trapped in the interstices of the gravel bed (Kuhnle, 1993a,b). Wilcock (1993) suggested that exponent x in Equation 2.14 would vary between 0 for unimodal sediments to 1 for bimodal sediments with peaks in the sand and medium gravel sizes.

To simulate the variable mobility of sand numerically, the hiding exponent, x, could be linked to the overall bed GSD. In a gravel bed, containing less than 30% sand, transport is only slightly size selective. As the threshold is crossed the critical shear stress for entrainment of both gravel and sand is reduced (Wilcock, 1998; Ikeda and Iseya, 1988). The critical stress for sand drops more than for gravel, however, and therefore sediment transport takes occurs on a stronger size selective basis. Referring back to the bed GSD could inform a model when the threshold is approached, and therefore when to reduce the critical shear stress and increase the strength of size selective transport.

9.2.2 Influence of sand on near-bed hydraulics

The above section highlights the fact that there is an abrupt switch in surface sedimentology from gravel framework to sand matrix with a small increase in bulk bed surface sand content. This fact brings into question the reliability of the hydraulics routine of SEDROUT. The model employs a function using bed D_{84} (k_S in Equation 2.12) to calculate roughness. This, in turn, affects flow depth and therefore shear stress (see Equation 2.2). In a gravel framework bed, using a function of D_{84} to calculate roughness may be representative. As the proportion of sand in the bed increases, however, the surface sedimentology changes abruptly from a gravel framework to a sand-matrix. In a sand matrix bed containing approximately 30% bulk sand the D_{84} will be of gravel size. Sambrook Smith *et al* (1997) suggested that the bed will behave hydraulically as a sand bed and this hypothesis is supported by the findings of the current research. The hydraulics routine of the model, therefore, does not represent accurately the conditions occurring in the field. As with the hiding factors discussed above, the definition of k_S in the roughness equation utilised in the hydraulics routine must be a variable, dependant on the proportion of sand in the bed surface. With increasing sand content, k_S will decrease although the reduction will be non-linear as the sand threshold is reached. A rapid drop in k_S would act to lower the water depth by reducing roughness. This would, in turn, reduce the shear stress and therefore impact on the mobility of sediments (Sambrook Smith and Ferguson, 1996). As the excess stress for sand is significantly lower than that for gravel, however, some sand would remain in transport, increasing the sharpness of the GST.

Although the fine gravel tracer experiments carried out in Allt Dubhaig (and reported in Chapter 5) suggested that the overpassing of gravel on sand did not occur, anecdotal evidence from Vedder Canal suggests that this may be a process in some streams. Downstream of the distal limit of the gravel bed (N10 to N12) there were fine gravel deposits in the troughs of large sand ripples. The overpassing of gravel on a sand-dominated bed is another process is not simulated by SEDROUT. Size selective bedload transport is assumed during runs although if gravel overpassing occurs the coarse gravels would move more rapidly than the finer sands by virtue of their increased exposure (Wilcock, 1993; Sambrook Smith *et al*, 1997).

9.2.3 Bed grain size sorting in the region of a GST

Since the current research is investigating the characteristics of GSTs it is necessary to study the changes in bed grain size occurring in a downstream direction as a river switches from a gravel-dominated to a sand-dominated bed. This streamwise grain size sorting occurs over a relatively short distance compared to the spatial rate of fining exhibited by the gravel bed channel upstream. As Chapters 4 and 5 noted, however, significant grain size sorting also occurs in the cross-stream dimension in channels with both gravel and sand sediments present on their beds.

Downstream sorting

Both Allt Dubhaig and Vedder Canal exhibited a decrease in the modal grain size for gravel in a streamwise direction. Abrasion is unlikely to be the cause of this fining as it occurs over such a short distance. At Allt Dubhaig in 1999 the modal gravel size reduced from 16-23, 16-23, 11-16, 8-11, 4-5.6 mm, to no gravel mode over a distance of approximately 190 m. The magnitude of the gravel mode also decreased with distance downstream (see Table 4.7 in Chapter 4). This finding indicates that strong size selective transport is occurring over this distance as a result of the increased sand supply associated with the switch from a gravel framework to a sand matrix bed.

The smooth downstream fining trend in Vedder Canal is complicated by the presence of bars in the gravel-bedded part of the channel (at N4, N6 and N8). These sections exhibited a higher bed D_{50}, and lower sand proportion, than the intervening sections N5 and N7. It can be inferred that the gravel bars are acting as stores for some of the coarser bed sediment. If this is the case in other rivers then it follows that the position of the last gravel bar is crucial for the location of the GST since the bar would retain the coarsest bed sediments. Downstream of this only the finer gravels and sands would be present in the bed. This process would enhance the development of a GST as the finer downstream gravel present beyond the last gravel bar has a smaller volume of pore space available to hold sand. Less sand would be needed to fill the pores before fines are deposited on the surface initiating a switch from a clast-supported gravel bed to a matrix-supported sand bed.

The decrease in bed D_{50}, and increase in the proportion of sand, associated with the switch from a gravel framework to a sand matrix bed was of similar magnitude to the coarsening-fining cycles associated with the bar / inter-bar samples (see Figure 4.13 in Chapter 4). This finding indicates that some threshold of sand content in the gravel framework bed must have been crossed downstream of the last gravel bar, to cause the switch to a sand matrix as a gravel framework bed dominated even at the inter-bar sections. The visual abruptness of the GST is therefore a product of a decrease in the proportion of the bed surface composed of gravel facies and bars rather than an abrupt change in grain size or bulk sand proportion in the bed.

An important difference between the width-averaged GSDs from the Vedder and those from other rivers which exhibit GSTs, such as the Allt Dubhaig, is that there was no development of a clear bimodal sediment mixture in the transition reach, although vaguely bimodal sediments were present at some sections. This bimodality was best developed in Vedder Canal at sections which do not exhibit gravel bars (N5 and N7, see Figure 9.1). A number of samples showed GSDs which have large amounts of sediment found in what has traditionally been though of as the grain size gap (N9 and N11, see Figure 9.2). This has implications for the sampling of fluvial sediments, since it has often been assumed that there is a relative paucity of sediments between approximately 1 and 8 mm in diameter (Shaw and Kellerhals, 1982; Chapter 2, Section 2.3 of this thesis). The apparent lack of sediments of this size may be due to the fact that in the past sediments were not sampled across the entire channel width including the wetted perimeter. The gap size fractions occur in the tail of most point GSDs (the fine tail in a gravel framework and the coarse tail in a sand matrix). These sizes may therefore look insignificant at a particular location but, as they are present in both gravel framework and sand matrix beds, their importance grows for width-averaged GSDs. The grain size gap sediments may have been located in relatively narrow strips of the channel width and sampling at only barhead locations would therefore not include these sediments. Using the bimodality index of Sambrook Smith *et al* (1997, Equation 4.1 of this thesis), therefore, may be misleading as it implies a paucity of sediment exists between the sand and gravel modes. Samples from barheads used to characterise the GSD of a stream would also skew the grain size estimates towards the coarse fraction of the bed material as the sediments at cross sections containing bars tend to be coarser than sediments present in inter-bar areas. If these samples were used as the initial conditions for a model run the output would be unreliable as the GSDs are not fully representative of the natural conditions.

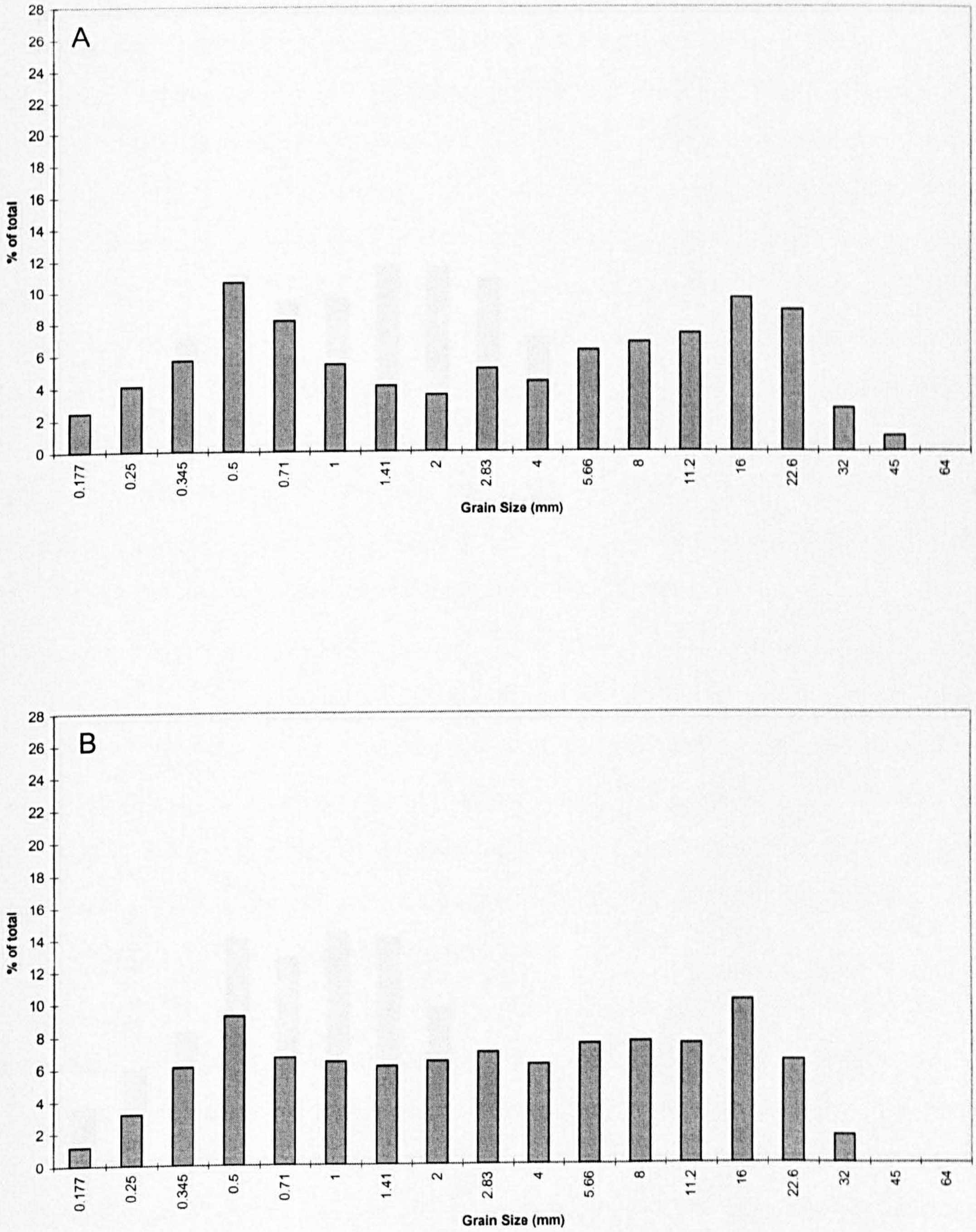
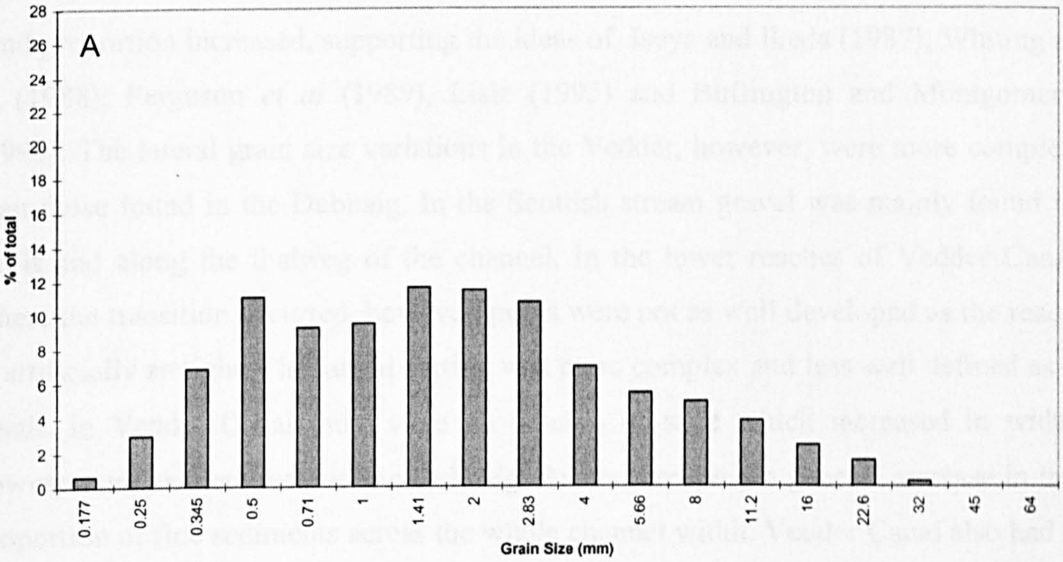


Figure 9.1: Histogram plots of bed surface GSD from Vedder Canal at N5 (a) and N7 (b).



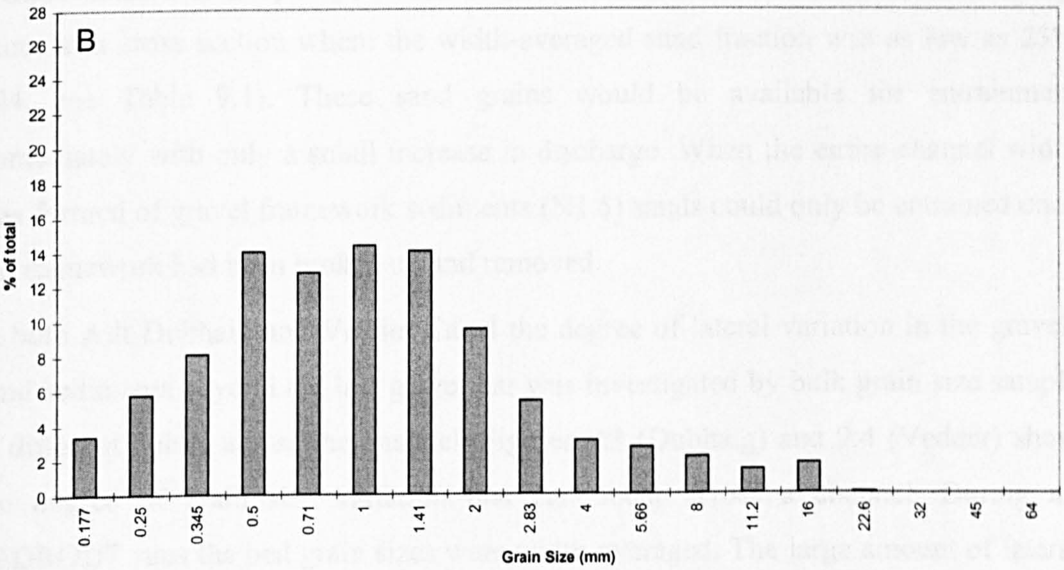


Figure 9.2: Histogram plots of bed surface GSD from Vedder Canal at N9 (a) and N11 (b).

Lateral sorting

As noted above, there are similarities between the bed surface facies of Allt Dubhaig and Vedder Canal. Both rivers exhibited pronounced lateral sorting downstream as the sand proportion increased, supporting the ideas of Iseya and Ikeda (1987), Whiting *et al* (1988), Ferguson *et al* (1989), Lisle (1995) and Buffington and Montgomery (1999). The lateral grain size variations in the Vedder, however, were more complex than those found in the Dubhaig. In the Scottish stream gravel was mainly found in pools and along the thalweg of the channel. In the lower reaches of Vedder Canal where the transition occurred, however, pools were not as well developed as the reach is artificially straight. The lateral sorting was more complex and less well defined as a result. In Vedder Canal there were no patches of sand which increased in width downstream, as there were in the Dubhaig. Rather there was a general increase in the proportion of fine sediments across the whole channel width. Vedder Canal also had a higher width-depth ratio giving more scope for mid-channel bars which further complicated the lateral sorting of different grain-size patches.

The presence of the grain size patches may be crucial to the formation and location of a GST. Grain size sampling in Vedder Canal showed that sand matrix patches could occur at a cross section where the width-averaged sand fraction was as low as 23% (N4, see Table 9.1). These sand grains would be available for entrainment immediately with only a small increase in discharge. When the entire channel width was formed of gravel framework sediments (N1.5) sands could only be entrained once the framework had been broken up and removed.

In both Allt Dubhaig and Vedder Canal the degree of lateral variation in the gravel-sand sediments beyond the last gravel bar was investigated by bulk grain size sample at different points across the channel. Figures 9.3 (Dubhaig) and 9.4 (Vedder) show the degree of grain size variation that can occur across a channel. During all SEDROUT runs the bed grain sizes were width-averaged. The large amount of lateral sorting exhibited by gravel-sand bed rivers was therefore not included.

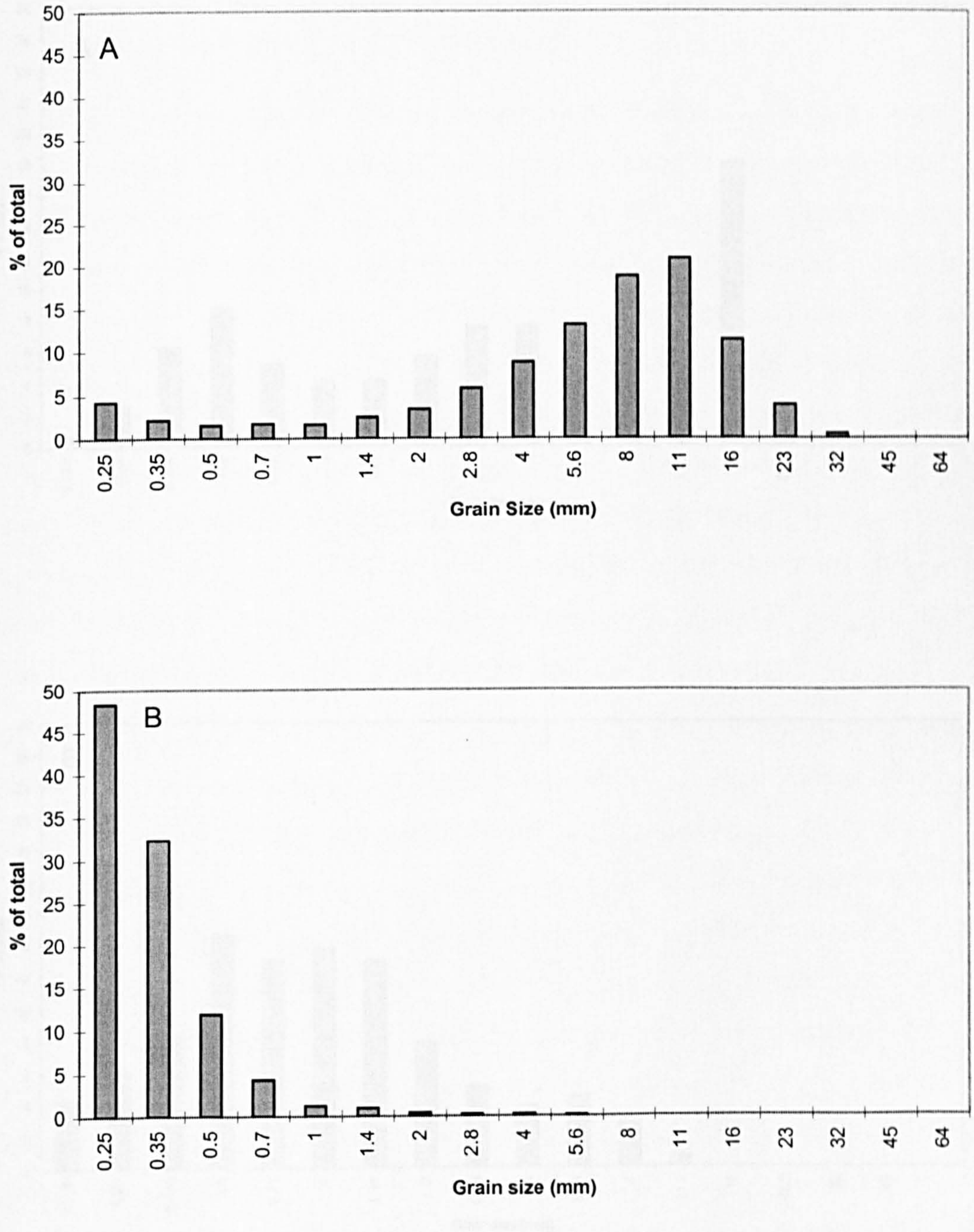


Figure 9.3: Histogram plots showing lateral variation at the same distance (cross section 10) downstream from Allt Dubhaig. Sample A is from left bank - 7 m, and B is from 7 m - right bank.

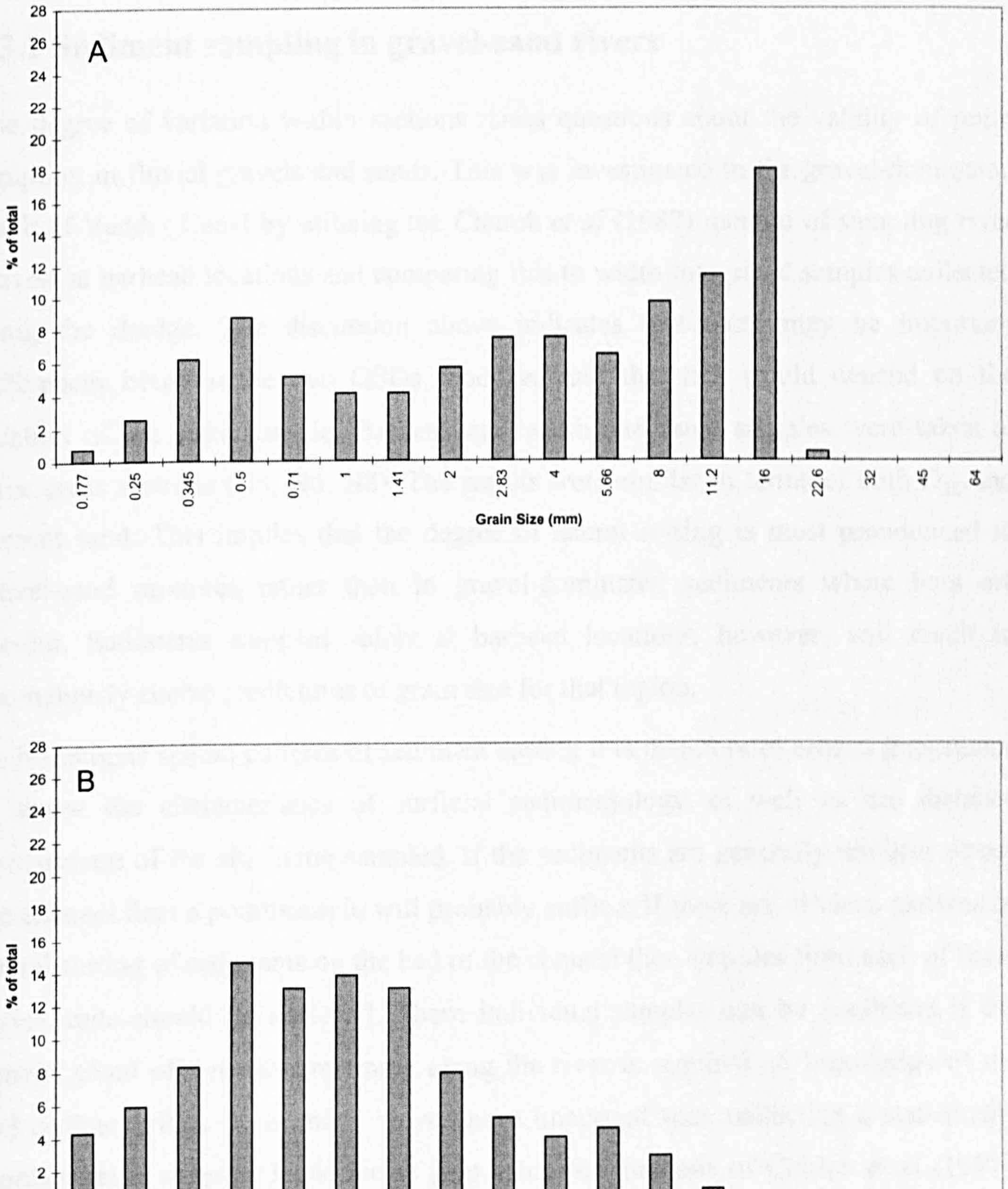


Figure 9.4: Histogram plots showing lateral variation at the same distance downstream (N11) from Vedder Canal. Sample A is from 10.2-16.6 and B is from 23.2-49 m across the channel.

9.3 Broader implications of the current research for the study of gravel-sand bed rivers

9.3.1 Sediment sampling in gravel-sand rivers

The degree of variation within sections raises questions about the validity of point sampling in fluvial gravels and sands. This was investigated in the gravel-dominated reach of Vedder Canal by utilising the Church *et al* (1987) method of sampling river gravels at barhead locations and comparing this to width-integrated samples collected using the dredge. The discussion above indicates that there may be important differences between the two GSDs produced and that this would depend on the location of the point sample. Barhead and width-integrated samples were taken at three cross sections (N4, N6, N8). The results were similar in terms of both D_{50} and percent sand. This implies that the degree of lateral sorting is most pronounced in gravel-sand mixtures rather than in gravel-dominated sediments where bars are present. Sediments sampled solely at barhead locations, however, will result in anomalously coarse predictions of grain size for that region.

To investigate spatial patterns of sediment sorting it is therefore of crucial importance to know the characteristics of surficial sedimentology as well as the distance downstream of the site being sampled. If the sediments are generally uniform across the channel then a point sample will probably suffice. If there are obvious patterns of lateral sorting of sediments on the bed of the channel then samples from each of these facies units should be retrieved. These individual samples can be combined if the general trend of downstream fining along the river is required. A knowledge of the bed sedimentology is, in many ways, more important than collecting a statistically representative sample of one facies type using the methods of Church *et* al (1987), Ferguson and Paola (1997) or Petrie and Diplas (2000). It is important to note that both a point sample and a width integrated sample may miss a small part of the bed that is locally finer or coarser than elsewhere at a given distance downstream. The presence of a sand patch will, however, ensure that entrainment of fines can occur at lower shear stresses than a width averaged or point sample would imply since the fine sediments would not be hidden in the interstices of coarser particles, but available for transport at the bed surface (see Ferguson *et al*, 1989; Seal and Paola, 1995; Wilcock, 1998). If these patches make up a large enough proportion of the bed then the transport of fines relative to coarse sediments will greatly increase, initiating a GST. It is therefore the presence of sand patches on the bed that is the key control on GST development.

9.3.2 Influence of dropout from suspension on GST formation

The water surface profile surveyed across the GST in Allt Dubhaig exhibited a clear break of slope immediately downstream of the last gravel bar. This reduction in bed slope can be expected to be associated with a relative decrease in gravel transport with respect to sand. In Allt Dubhaig, sand matrix sediments began to dominate the bed surface approximately 60 m downstream of the break of slope. Kodama (1994a,b) found a similar sedimentological change downstream of a break of slope in the Watarase River, Japan. It was inferred that in this stream some of the gravel was transported beyond the break of slope but its mobility was greatly inhibited by the reduction in shear stress. As there is a break of slope it might also be expected that sand would be deposited from suspension as the shear stress fell.

The bed slope of Vedder Canal, however, does not exhibit a break of slope in the region of the last gravel bar, meaning dropout from suspension is unlikely to be the cause of the persistence of the GST in this stream. The GST may have formed initially as the result of a break of slope at the head of the Canal when it was first built but there is no evidence of this change in gradient today. The fact that the GST in Vedder Canal occurs without the presence of a break of slope suggests that in some cases transitions can remain in place through changes in bedload transport dynamics alone. The switch from a gravel framework to a sand-matrix bed with only a small increase in bulk sand content, and the lateral sorting of sediments into patches of different grain size assist in the preservation of the GST in this stream.

9.3.3 Influence of flood frequency and magnitude on GST location

The hydrological regime of a stream is crucial in determining its bed structure and bedload transport characteristics (Reid *et al*, 1985; Reid and Frostick, 1986; Laronne *et al*, 1994; Reid and Laronne, 1995). In most perennial streams, floods are infrequent and the discharge is dominated by extended periods of low flow. During these low flow periods, fine sediments are winnowed from the gravel framework bed. These fines are supplied to the channel downstream. The low flows also act to stabilise the gravel framework as the coarse grains become locked together and any remaining fines settle into the interstices, further strengthening the framework (Reid *et al*, 1985). The entrainment of gravel during floods is therefore delayed as the gravel grains are secured in place by the matrix and their coarse neighbours (Hassan and Church, 2000).

In regions where there are no extended periods of low flow, for example in dryland areas, no winnowing of the fine grains can occur and the gravel bed does not have the opportunity to become armoured with the grains interlocked. In this situation coarser grains would be prevalent in the bedload and its GSD would be similar to that of the bed. The development of a GST would therefore be precluded as few fine sediments would be preferentially entrained.

If this hypothesis is correct it follows that GSTs would be most prevalent in reaches of perennial streams with long periods of low flow, which allow partial and selective transport of sands, and lead to the development of a strong armour layer. The occurrence of frequent floods would lead to less armouring and structuring of the bed surface and therefore more gravel would be transported with respect to sand. The presence of a well defined, armoured, structured gravel bar upstream of the GST is therefore crucial in defining where the transition occurs. This hypothesis agrees with the results from Vedder Canal regarding the width-averaged downstream fining trend. In the Canadian stream smooth fining was punctuated by gravel bars acting as stores for coarse sediments and the GST itself occurred immediately downstream from the last gravel bar.

9.4 Directions for further research

The current research aimed to answer a number of specific questions. These are presented in the following, concluding chapter. The study has, however, raised a series of further issues related to both the modelling and field aspects of the investigations. A number of these issues follow on from the questions asked at the end of Chapter 2, where more detailed studies are required, and some are noted for the first time here.

9.4.1 Development of one-dimensional sediment routing models for gravel-sand mixtures

The findings of the current research indicate that a number of modifications could be made to SEDROUT to improve its simulations of gravel-sand mixtures. The degree to which each of these modifications would improve model predictions requires assessment. These potential improvements and inclusions are listed below:

- a variable bedload-bed exchange parameter dependant on gravel framework sand content to allow the overwhelming of a gravel bed by sand
- a variable hiding factor for individual size fractions based on their proportions in the overall bed GSD
- a function that uses a variable grain size parameter to define bed roughness depending on the overall bed GSD
- a method of defining variable depth across a channel allowing variable shear stress and bedload transport rates across the simulated width
- a method of including lateral sorting into different grain size patches to allow a variable critical shear stress for entrainment at different positions across the simulated width

The latter two suggestions of potential model modifications cannot practically be achieved within a one-dimensional sediment routing model.

9.4.2 Field research in gravel-sand bed rivers

The results of field investigations presented in this thesis have shown that the characterisation of gravel-sand bed rivers is complex. Further studies are required to elucidate the following:

- the formation of a GST without the presence of a sharp break of slope

- how widespread the threshold in the amount of interstitial sand that can be held in a gravel framework is, and if the figure always in the region of 30%
- the change in near-bed hydraulics as the sand threshold is crossed
- the manifestation of lateral grain size sorting in other single thread streams
- the complexity of lateral grain size sorting in other channel types, such as braided rivers
- the occurrence of lateral sorting in gravel-dominated channels
- the processes causing sorting of gravel and sand into patches in the united gravel-sand reach
- the influence of flood frequency and magnitude, and the importance of periods of low flow, on GST location

Chapter 10. Conclusions

Towards the end of Chapter 2 (Section 2.5) a number of specific research questions were posed that the current research attempts to answer. The questions, grouped into three objectives, tackle issues associated with elucidating the forms and processes occurring along a gravel bed river as the proportion of sand increases. Answers to the questions are presented in this chapter, using the findings of this thesis as evidence.

10.1 Characterisation of contemporary GST sedimentology

- How does bed grain size change in a streamwise direction through a GST?

Chapter 4 outlined the downstream reduction in bulk bed grain size upstream of, and through, the GST. The magnitude and grain size of the gravel mode decreases and the percentage of sand increases along the river. The rate of change of bed grain size downstream is higher in the GST than in the gravel reach upstream but is not as sharp as visual inspection indicates. The decrease in grain size is not smooth, however, and in Vedder Canal is punctuated by coarser than average and finer than average bed GSDs associated with bar / inter-bar reaches of the channel.

- To what extent does lateral sorting of grain sizes occur in the united gravel-sand reach?

In the GST reach of Allt Dubhaig and Vedder Canal there is clear evidence of lateral sorting of sediments into grain size patches. It is shown, in Chapters 4 and 9, that this lateral sorting into patches of sand matrix and gravel framework sediments can occur at width-averaged sand contents of less than 30%. This lateral sorting will impact on the mobility of individual grain sizes as sand is available at a shear stress lower than that necessary to mobilise a gravel framework.

- Can a quantitative assessment of the proportion of sand in the bed be obtained by a qualitative observation of bed surface sedimentology?

Chapter 9 included a discussion of the accuracy of the modified Kodama (1994b) technique to assess bed surface sand content. An analysis indicates that, while not giving an exact value as would be obtained by sieving, for a given gravel-sand mixture, the Kodama values can be transferred into an accurate range of bulk bed surface sand.

- How can gravel-sand sediments be sampled to obtain representative bed GSDs for a particular distance downstream?

The evidence presented in Chapter 4, and the discussion in Chapter 9, suggest that careful sampling of sediments for the characterisation of gravel-sand bed rivers is necessary, due to the extent of lateral sorting. Ideally, a knowledge of bed surface sedimentology is required prior to the collection of samples to ensure that they are representative of the cross section as a whole. In addition, this thesis indicates that barhead samples may lead to anomalously coarse GSDs being assigned to particular reaches of gravel-dominated rivers.

- What proportion of bulk sand is required to cause a switch in bed surface sedimentology from gravel framework to sand matrix?

Chapters 4 and 9 showed that there is a threshold in the amount of sand a gravel framework bed can hold, above which the channel bed becomes sand matrix dominated. This threshold lies between 20 and 30% bulk bed sand content.

- What is the relationship between bulk and areal sand content in gravel-sand bed streams?

During the discussion in Chapter 9 it was found that there is a non-linear relationship between the bulk and areal extent of sand in a gravel-sand bed channel. With small increases in the bulk sand proportion there are large increases in the area of bed surface sand. In Vedder Canal, for example, a 7% increase in the width-averaged bulk sand content causes a 36% increase in the bed surface occupied by sand matrix sediments.

10.2 Channel change and mechanisms of GST formation

- To what extent does size selective transport occur in distal fine gravels immediately upstream of the GST?

A fine-gravel tracer experiment, presented in Chapter 5, carried out in the distal gravel reach of Allt Dubhaig showed no evidence of size selective transport operating. It is postulated that this is due to the combined effects of only a small number of flows occurring that are capable of entraining the tracers, and low virtual velocities on very gentle slopes. If these tracers had remained in place for a longer period then they might have experienced size selective transport.

- Is there evidence of processes causing sorting of fine gravel into patches in the united gravel-sand reach?

Although the tracers seeded in the united gravel-sand reach did undergo size selective transport, there is no evidence of processes sorting these coarser particles into grain size patches. Conversely, the tracers tend to remain on the bed type on which they are seeded, and those seeded on gravel are more mobile than those seeded on sand. The fact that the bed surface sedimentology is strongly related to channel morphology may go some way to explaining this apparent lack of patch sorting. The gravel patches in Allt Dubhaig occur in the deepest part of the channel and are hence more active than the sand patches which occur on the insides of bends or shallower parts of straight reaches. As such, the tracers seeded on sand patches experience a lower shear stress than those seeded on gravel, impeding their mobility. The transport distances may also have been insufficient to transport the tracers from the patches on which they were seeded.

- How does the surface morphology and sedimentology of the gravel front and united gravel-sand reach evolve over time periods of 10^0 to 10^2 years?

Surface sedimentological maps of Allt Dubhaig surveyed in 1992 and 1997, presented in Chapter 5, show progradation of gravel patches occurring in the united gravel-sand reach. Probing in the transition zone of the Dubhaig also indicates that the gravel front and gravel patches may prograde downstream over a longer timescale. The construction of a small weir downstream of the GST, however, appears to have reversed this temporarily, shifting the three zones of differing bed sediment (gravel, united gravel-sand, sand) upstream.

- Are lateral inputs of fine sediment crucial for the formation of a GST?

Repeated cross section surveys at Allt Dubhaig and Vedder Canal, discussed in Chapter 5, show little lateral erosion between 1992 and 1997. In Allt Dubhaig, floodplain coring further supports this lack of channel movement. Bank sediment GSD analysis at Allt Dubhaig, indicating grains considerably finer than the fine mode in the bed, stable channel banks in Vedder Canal and the lack of tributary inputs all add weight to the argument that the GSTs in these streams are not the result of a lateral input of fine sediment.

- Do GSTs form without a sharp reduction in bed slope over a short distance?

In Allt Dubhaig there is an order of magnitude reduction in gradient immediately upstream of the GST (see Chapters 4 and 5). There is, however, no evidence of a sharp break of slope in the region of the GST in Vedder Canal. A break of slope was most likely present at the head of the Canal when it was first constructed. This suggests that while a break of slope may be necessary for a GST to form initially, the spatially rapid reduction in grain size can remain in place after the break of slope has been removed through aggradation.

10.3 Modelling the GST through width-averaged bedload sorting

- How does a one-dimensional model of size selective gravel bedload sorting need to be modified to be capable of simulating gravel-sand mixtures?

The introduction of sand into a gravel sediment routing model (SEDROUT) creates a number of unexpected problems, which were discussed in Chapter 7. By far the most complex of these is instabilities in the Parker (1990) surface-based bedload transport relation which is used in SEDROUT. This problem manifests itself in model runs with a wide range of grain sizes present on the simulated bed. In this case, transport rate dropped with increasing shear stress, associated with the method the model used to calculate the bed surface GSD from the subsurface (the, so called, straining function). An alternative function was introduced to allow the current research to be completed.

- Can a GST be formed through numerical modelling of width-averaged size selective bedload sorting alone?

Model runs, presented in Chapter 8, with a smooth concave channel long profile did not develop a spatially accelerated rate of downstream fining. This was the case even when the fining wave was passing through the simulated reach.

- Can a GST be formed through numerical modelling of width-averaged size selective bedload sorting in the presence of a spatially rapid break of slope?

The introduction of a spatially rapid break of slope caused the development of a spatially accelerated rate of downstream fining. This rapid fining was not, however, of the same strength as that witnessed in the field.

- Can a GST be formed through numerical modelling of width-averaged size selective bedload sorting and the overwhelming of a gravel bed by sand?

A series of model runs, presented in Chapter 8, were undertaken to simulate the overwhelming of a gravel bed by sand in the presence of a break of slope. Sand was deposited into the bed subsurface until it composed 30% of the sediments and all sand was then deposited on the channel surface. These runs generated a GST and the cross sectional nodes remained fine for a number of model output dumps, rather than the rapid coarsening experienced by all the other runs carried out for the current research. The predicted GST was no more abrupt, however, than that generated by the run featuring only a sharp break of slope in the long profile.

- How can a one-dimensional model of size selective gravel bedload sorting be improved to simulate accurately the behaviour of gravel-sand mixtures?

A number of improvements could be made to SEDROUT to improve its prediction of the behaviour a gravel-sand mixtures. These include: a bedload-bed exchange parameter that is variable depending on bed surface and subsurface sand content; variable hiding factors for each size fraction dependent on their relative proportions in the bed; and a method for accounting for lateral variations in grain size and channel depth which will influence the mobility of different grain sizes.

References

Ackers, P. 1992. 1992 Gerald Lacey memorial lecture - Canal and river regime in theory and practice 1929-92. *Proceedings of the Institution of Civil Engineers*, **96**, 167-178.

Adams, J. 1979. Wear of unsound pebbles in river headwaters. *Science*, **203**, 171-172.

Adams, J. 1980. Contemporary uplift and erosion of the Southern Alps, New Zealand. *Geological Society of America Bulletin*, Part 2, **91**, 1-114.

Allan, A.F. and Frostick, L. 1999. Framework dilation, winnowing, and matrix particle size: the behaviour of some sand-gravel mixtures in a laboratory flume. *Journal of Sedimentary Research*, **69**, 21-26.

Andrews, E.D. 1979. Scour and fill in a stream channel, East Fork River, western Wyoming. *USGS Professional Paper*, **1117**.

Andrews, E.D. 1980. Effective and bankfull discharge of streams in the Yampa Basin, Colorado and Wyoming. *Journal of Hydrology*, **46**, 311-330.

Andrews, E.D. 1983. Entrainment of gravel from naturally sorted riverbed material. *Geological Society of America Bulletin*, **94**, 1225-1231.

Andrews, E.D. 2000. Bed material transport in the Virgin River, Utah. *Water Resources Research*, **36**, 585-596.

Andrews, E.D. and Erman, D.C. 1986. Persistence in the size distribution of surficial bed material during an extreme snowmelt flood. *Geological Society of America Bulletin*, **22**, 191-197.

Andrews, E.D. and Nankervis, J.M. 1995. Effective discharge and the design of channel maintenance flows for gravel-bed rivers. In Costa, J.E., Miller, A.J., Potter, K.W. and Wilcock, P.R. (eds). *Natural and Anthropogenic Influences in Fluvial Geomorphology*, 151-164, American Geophysical Union Monograph, **89**, Washington DC.

Andrews, E.D. and Parker, G. 1987. Formation of a coarse surface layer as the response to gravel mobility. In Thorne, C.R., Bathurst, J.C. and Hey, R.D. (eds). *Sediment Transport in Gravel Bed Rivers*, 269-325, John Wiley and Sons. Chichester, England.

Ashworth, P.J. and Ferguson R.I. 1986. Interrelationships of channel processes, changes and sediments in a proglacial river. *Geografiska Annaler*, **25**, 627-634.

Ashworth, P.J. and Ferguson, R.I. 1989. Size-selective entrainment of bed load in gravel bed streams. *Water Resources Research*, **25**, 627-634.

Ashworth, P.J., Ferguson, R.I., Ashmore, P.E., Paola, C., Powell, D.M. and Prestegaard, K.L. 1992. Measurements in a braided river chute and lobe. 2. sorting of bedload during entrainment, transport and deposition. *Water Resources Research*, **28**, 1887-1896.

Bagnold, R.A. 1966. An approach to the sediment transport problem from general physics. *USGS Professional Paper*, **422-I**.

Bagnold, R.A. 1973. The nature of saltation and bedload transport in water. *Proceedings of the Royal Society*, London, **A322**, 473-504.

Batalla, R.J. 1997. Evaluating bed-material transport equations using field measurements in a sandy gravel-bed stream, Arbucies, NE Spain. *Earth Surface Processes and Landforms*, **22**, 121-130.

Batalla, R.J. and Sala, M. 1995. Effective discharge for bedload transport in a sub-humid Mediterranean sandy gravel-bed river, Arbucies, north-east Spain. In Hickin, E.J. (ed). *River Geomorphology*, 93-103, John Wiley and Sons. Chichester, England.

Bergeron, N.E. and Caronneau, P. 1999. The effect of sediment concentration on bed roughness. *Hydrological Processes*, **13**, 2583-2589.

Beschta, R.L. and Jackson, W.L. 1979. The intrusion of fine sediments into a stable gravel bed. *Journal of Fisheries Research Board*, Canada, **36**, 204-210.

Bradley, W.C. 1970. Effects of weathering on abrasion of granitic gravel, Colorado River, Texas. *Geological Society of America Bulletin*, **81**, 61-80.

Bradley W.C., Fahnestock, R.K. and Rowekamp, E.T. 1972. Coarse sediment transport by flood flows on Knik River, Alaska. *Geological Society of America Bulletin*, **83**, 1261-1284.

Bray, D.I. 1982. Flow resistance in gravel bed rivers. In Hey, R.D., Bathurst, J.C. and Thorne, C.R. (eds). *Gravel Bed Rivers: Fluvial Processes*, 109-137, John Wiley and Sons. Chichester, England.

Brewer, P.A. and Lewin, J. 1993. In-transport modification of alluvial sediment: field evidence and laboratory experiments. *Spec. Publs. Int. Ass. Sediment.*, **17**, 23-35.

Brewer, P.A., Leeks, G.J.L. and Lewin, J. 1992. Direct measurement of in-channel abrasion processes. In Erosion and sediment transport monitoring programmes in river basins. *International Association of Hydrological Sciences Publication*, Proceedings of the Oslo Symposium, 21-29, **210**.

Bridge, J.S. and Dominic, D.D. 1984. Bedload grain velocities and sediment transport rates. *Water Resources Research*, **20**, 476-490.

Brierley G.J. and Hickin, E.J. 1985. The downstream gradation of particle sizes in the Squamish River, British Columbia. *Earth Surface Processes and Landforms*, **10**, 579-606.

Buffington, J.M. and Montgomery, D.R. A procedure for classifying textural facies in gravel-bed rivers. *Water Resources Research*, **35**, 1903-1914.

Burkham, D.E. and Dawdy, D.R. 1976. Resistance equation for alluvial-channel flow. *Journal of Hydraulics Division,* ASCE, **102**, 1479-1489.

Campbell, I.A. 1970. Erosion rates in the Steveville Badlands, Alberta. *Canadian Geographer*, **14**, 202-216.

Campbell, I.A. 1977. Stream discharge, suspended sediment and erosion rates in the Red Deer River basin, Alberta, Canada. *International Association of Hydrological Sciences Publication*, **122**, 244-259.

Carling, P.A. 1983. Threshold of coarse sediment transport in broad and narrow natural streams. *Earth Surface Processes and Landforms*, **8**, 1-18.

Carling, P.A. 1984. Deposition of fine and coarse sand in an open-work gravel bed. *Canadian Journal of Fisheries and Aquatic Sciences*, **41**, 263-270.

Carling, P.A. and Reader, N.A. 1982. Structure, composition and bulk properties of upland stream gravels. *Earth Surface Processes and Landforms*, **7**, 349-365.

Charlton, F.G., Brown, P.M. and Benson, R.W. 1978. The hydraulic geometry of some gravel rivers in Britain. *Hydraulics Research Station Report*, **IT180**, Wallingford, England.

Church, M. 1978. Palaeohydrological reconstructions from a Holocene valley fill. In Miall, A.D. (ed). *Fluvial Sedimentology*, 743-772, Canadian Society of Petroleum Geologists Memoir, **5**, Calgary, Canada.

Church, M. and Kellerhals, R. 1978. On the statistics of grain size variation along a gravel river. *Canadian Journal of Earth Sciences*, **15**, 1151-1160.

Church, M., McLean, D.G. and Wolcott, J.F. 1987. River bed gravels: sampling and analysis. In Thorne, C.R., Bathurst, J.C. and Hey, R.D. (eds). *Sediment Transport in Gravel Bed Rivers*, 43-88, John Wiley and Sons. Chichester, England.

Church, M., Hassan, M.A. and Wolcott, J.F. 1998. Stabilizing self-organised structures in gravel-bed stream channels: field and experimental observations. *Water Resources Research*, **34**, 3169-3179.

Church, M., Wolcott, J.F. and Fletcher, W.K. 1991. A test of equal mobility in fluvial sediment transport: behaviour of the sand fraction. *Water Resources Research*, **27**, 2941-2951.

Cui, Y., Parker, G. and Paola, C. 1996. Numerical simulation of aggradation and downstream fining. *Journal of Hydraulic Research*, **34**, 185-204.

Cui, Y.T. and Parker, G. 1998. The arrested gravel front: stable gravel-sand transitions - part 2: general numerical solution. *Journal of Hydraulic Research*, **36**, 159-182.

Dade, W.B. and Friend, P.F. 1998. Grain-size, sediment-transport regime, and channel slope in alluvial rivers. *Journal of Geology*, **106**, 661-675.

Davis, W.M. 1899. The geographical cycle. *Geographical Journal*, **14**, 481-504.

Davis, W.M. 1902. Base-level, grade and peneplain, *Journal of Geology*, **10**, 77-111.

Dawson, M.D. 1988. Sediment size variation in a braided reach of the Sunwapta River, Alberta, Canada. *Earth Surface Processes and Landforms*, **13**, 599-618.

Deigaard, R. 1982. Longitudinal sorting of grain sizes in alluvial rivers. Proc. Euromech. **156,** *Mechanics of sediment transport*, 231-236.

Deigaard, R. and Fredsoe, J. 1978. Longitudinal grain sorting by current in alluvial streams. *Nordic Hydrology*, **9**, 7-16.

Dietrich, W.E., Kirchner, J.W., Ikeda, H. and Iseya. 1989. Sediment supply and the development of the coarse surface layer in gravel-bed rivers. *Nature*, **340**, 215-217.

Diplas, P. 1987. Bedload transport in gravel-bed streams. *Journal of Hydraulic Engineering*, **113**, 277-292.

Diplas, P. and Parker, G. 1985. Pollution of gravel spawning grounds due to fine sediment. *St. Anthony Falls Hydraulic Laboratory Project Report*, **240**, University of Minnesota, Minneapolis.

Diplas, P & Parker, G, 1992, Deposition and removal of fines in gravel-bed streams, In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 313-329, John Wiley and Sons. Chichester, England.

Dunkerley, D.L. 1990. The development of armour in the Tambo River, Victoria, Australia. *Earth Surface Processes and Landforms*, **15**, 405-415.

Einstein, H.A. 1968. Deposition of suspended particles in a gravel-bed. *Journal of Hydraulics Division,* ASCE, **91**, 225-247.

Egiazaroff, I.V. 1965. Calculation of nonuniform sediment concentrations. *Journal of Hydraulic Engineering*, **91**, 225-247.

Egaizaroff, I.V. 1967. Sediment transport mechanics - initiation of motion. *Journal of Hydraulics Division,* ASCE, **93**, 281-287.

Engelund, F. and Fredsoe, J. 1976. A sediment transport model for straight alluvial channels. *Nordic Hydrology*, **7**, 293-300.

Fenton, J.D. and Abbott, J.E. 1977. Initial movement of grains on a stream bed: the effects of relative protrusion. *Proceedings of the Royal Society*, London, **A352**, 523-537.

Ferguson, R.I. and Ashworth, P.J. 1991. Slope-induced changes in the channel character along a gravel-bed stream: the Allt Dubhaig, Scotland. *Earth Surface Processes and Landforms*, **16**, 65-82.

Ferguson, R.I. and Paola, C. 1997. Bias and precision of percentiles of bulk grain size distributions. . *Earth Surface Processes and Landforms*, **22**, 1061-1077.

Ferguson, R.I. and Wathen, S.J. 1998. Tracer pebble movement along a concave river profile: virtual velocity in relation to grain size and shear stress. *Water Resources Research*, **34**, 2031-2038.

Ferguson, R.I., Prestegaard, K.L. and Ashworth, P.J. 1989. Influence of sand on hydraulics and gravel transport in a braided gravel-bed river. *Water Resources Research*, **25**, 635-643.

Ferguson, R.I., Hoey, T.B., Wathen, S.J. and Werritty, A. 1996. Field evidence for rapid downstream fining of river gravels through selective transport. *Geology*, **24**, 179-182.

Ferguson, R.I., Hoey, T.B., Wathen, S.J., Werritty, A., Hardwick, R.I. and Sambrook Smith, G.H. 1998. Downstream fining of river gravels: integrated field, laboratory and modelling study. In Klingeman, P.C., Beschta, R.L., Komar, P.D. and Bradley, J.B. (eds). *Gravel-Bed Rivers and the Environment*, 85-114, Water Resources Publications, Highlands Ranch, Colorado.

Frostick, L.E. Lucas, P.M. and Reid, I. 1984. The infiltration of fine matrices into coarse-grained alluvial sediments and its implications for stratigraphic interpretation. *Journal of the Geological Society*, London, **141**, 955-965.

Gasparini, N.M., Tucker, G.E. and Bras, R.L. 1999. Downstream fining through selective particle sorting in an equilibrium drainage network. *Geology*, **27**, 1079-1082.

Gilbert, G.K. 1914. The transportation of debris by flowing water. *USGS Professional Paper*, **86**.

Gomez, B. 1991. Bedload transport. *Earth Science Reviews*, **31**, 89-132.

Gomez, B. and Church, M. 1991. An assessment of bedload sediment transport formulae for gravel bed rivers. *Water Resources Research*, **25**, 1161-1186.

Griffiths G.A. 1979. Recent sedimentation history of the Waimakariri River, New Zealand. *New Zealand Journal of Hydrology*, **28**, 63-75.

Hack, J.T. 1957. Studies of longitudinal stream profiles in Virginia and Maryland. *USGS Professional Paper*, **294B**.

Hassan, M.A. and Church, M. 1994. Vertical mixing of coarse particles in gravel bed rivers: a kinematic model. *Water Resources Research*, **30**, 1173-118-5.

Hassan, M.A. and Church, M. 2000. Experiments on surface structure and partial sediment transport on a gravel bed. *Water Resources Research*, **36**, 1885-1895.

Hey, R.D. 1979. Flow resistance in gravel-bed rivers. *Journal of Hydraulics Division*, ASCE, **105**, 365-379.

Hey, R.D. and Thorne, C.R. 1986. Stable channels with mobile beds. *Journal of Hydraulics Division*, ASCE, **112**, 671-689.

Higgins, R.J., Pickup, G. and Cloke, P.S. 1987. Estimating the transport and deposition of mining waste at OK Tedi. In Thorne, C.R., Bathurst, J.C. and Hey, R.D. (eds). *Sediment Transport in Gravel Bed Rivers*, 243-260, John Wiley and Sons. Chichester, England.

Hoey, T.B. 1992. Temporal variations in bedload transport rates and sediment storage in gravel-bed rivers. *Progress in Physical Geography*, **15**, 319-338.

Hoey, T.B. and Bluck, B.J. 1999. Identifying the controls over downstream fining of river gravels. *Journal of Sedimentary Research*, **69**, 40-50.

Hoey, T.B. and Ferguson, R.I. 1994. Numerical simulation of downstream fining by selective transport in gravel bed rivers: model development and illustration, *Water Resources Research*, **30**, 2251-2260.

Hoey, T.B. and Ferguson, R.I. 1997. Controls of strength and rate of downstream fining above a river base level. *Water Resources Research*, **33**, 2601-2608.

Horton, R.E. 1945. Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology. *Geological Society of America Bulletin*, **56**, 275-370.

Howard, A.D. 1980. Thresholds in river regimes. In Coates, D.R. and Vitek, J.D. (eds). *Thresholds in Geomorphology*, 227-258, George Allen and Unwin, Boston.

Howard, A.D. 1987. Modelling fluvial systems: rock-, gravel- and sand-bed channels. In Richards, K.S. (ed). *River Channels: Environment and Process*, 69-94, Blackwell, England.

Ichim, I. and Radoane, M. 1990. Channel sediment variability along a river: a case study of the Siret River (Romania). *Earth Surface Processes and Landforms*, **15**, 211-225.

Ikeda, H. 1970. On the longitudinal profiles of the Asake, Mitaki and Utsbe Rivers, Mie Prefecture. *Geographical Review of Japan*, **43**, 2353-2364.

Ikeda, H. and Iseya, F. 1987. Thresholds in the mobility of sediment mixtures. In Gardiner, V. (ed). *International Geomorphology* , 1, 561-570, John Wiley and Sons, New York.

Ikeda, H. and Iseya, F. 1988. Experimental study of heterogeneous sediment transport. *University of Tsukuba, Environmental Research Center Papers*, **12**.

Iseya, F. and Ikeda, H. 1987. Pulsations in bedload transport rates induced by a longitudinal sediment sorting: a flume study using sand and gravel mixtures. *Geografiska Annaler*, **69A**, 15-27.

Jackson, W.L. and Beschta, R.L. 1982. A model of two-phase bedload transport in an Oregon coast range stream. *Earth Surface Processes and Landforms*, **7**, 517-527.

James, C.S. 1990. Prediction of entrainment conditions for nonuniform, noncohesive sediments. . *Journal of Hydraulic Research*, **28**, 25-41.

Jones, L.F. and Humphrey, N.F. 1997. Weathering-controlled abrasion in a coarse-grained, meandering reach of the Rio-Grande: implications for the rock record. *Geological Society of America Bulletin*, **109**, 1080-1088.

Kamphuis, J.W. 1974. Determination of sand roughness for fixed beds. *Journal of Hydraulic Research*, **12**, 193-203.

Keulegan, G.H. 1938. Laws of turbulent flow in open channels. *J. Res. Natl. Bur. Stand.*, **21**, 707-741.

Kirchner, J.W., Dietrich, W.E., Iseya, F. and Ikeda, H. 1990. Variability of critical shear stress, friction angle, and grain protrusion in water-worked sediments. *Sedimentology*, **37**, 647-672.

Knighton, A.D. 1980. Longitudinal changes in size and sorting of stream-bed material in four English rivers. *Geological Society of America Bulletin*, **91**, 55-62.

Knighton, A.D. 1989. River adjustment to changes in sediment load: the effects of tin mining on the Ringarooma River, Tasmania, 1875-1984. *Earth Surface Processes and Landforms*, **14**, 333-359.

Knighton, A.D. 1991. Channel bed adjustment along mine-affected rivers of northeast Tasmania. *Geomorphology*, **4**, 205-219.

Knighton, A.D. 1998. *Fluvial forms and processes: a new perspective*. Arnold, Great Britain.

Knighton, A.D. 1999. The gravel-sand transition in a disturbed catchment. *Geomorphology*, **27**, 325-341.

Kodama, Y. 1992. Effect of abrasion on downstream grain-size reduction in the Watarase River, Japan: fieldwork and laboratory experiment. *University of Tsukuba, Environmental Research Center Papers*, **15**.

Kodama, Y. 1994a. Downstream changes in the lithology and grain size of fluvial gravels, the Watarase River, Japan: evidence of the role of abrasion in downstream fining. *Journal of Sedimentary Research*, **64A**, 68-75.

Kodama, Y. 1994b. Abrupt channel slope change in the Lower Watarase River caused by thresholds in gravel and sand mobility. *Geographical Review of Japan*, **67A-5**, 311-324.

Kodama, Y. 1994c. Experimental study of abrasion and its role in producing downstream fining in gravel-bed rivers. *Journal of Sedimentary Research*, **64A**, 76-85.

Komar, P.D. 1987. Selective grain entrainment by a current from a bed of mixed sizes: a reanalysis. *Journal of Sedimentary Petrology*, **57**, 203-211.

Komar, P.D. 1988. Sediment transport by floods. In Baker, V.R., Kochel, R.C. and Patton, P.C. (eds). *Flood Geomorphology*, 97-111, Wiley-Interscience, New York.

Komar, P.D. 1989. Flow-competence evaluations of the hydraulic parameters of floods: an assessment of the technique. In Beven, K. and Carling, P. (eds). *Floods: Hydrological, Sedimentological and Geomorphological Implications*, 107-134, John Wiley and Sons. Chichester, England.

Komar, P.D. and Li, Z. 1986. Pivoting analyses of the selective entrainment of sediments by shape and size with application to gravel threshold. *Sedimentology*, **33**, 425-436.

Komar, P.D. and Shih, S-M. 1992. Equal mobility versus changing bedload grain sizes in gravel bed streams. In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 73-106, John Wiley and Sons. Chichester, England.

Kuenen, Ph.H. 1956. Experimental abrasion by pebbles, 2: rolling by current. *Journal of Geology*, **64**, 336-368.

Kuhnle, R.A. 1993a. Fluvial transport of sand and gravel mixtures with bimodal size distributions. *Sedimentary Geology*, **85**, 17-24.

Kuhnle, R.A., 1993b. Incipient motion of sand-gravel mixtures. *Journal of Hydraulic Engineering*, **119**, 1400-1415.

Kuhnle, R.A. and Southard, J.B. 1988. Bedload transport fluctuations in a gravel bed laboratory channel. *Water Resources Research*, **24**, 247-260.

Lane, E.W. 1955. The importance of fluvial morphology in hydraulic engineering. *Proc. ASCE*, **81**, 1-17.

Lane, S.N. 1998. Hydraulic modelling in hydrology and geomorphology: a review of high resolution approaches. *Hydrological Processes*, **12**, 1131-1150.

Lapointe, M. 1992. Burst-like sediment suspension events in a sand bed river. *Earth Surface Processes and Landforms*, **17**, 253-270.

Laronne, J.B. and Reid, I. 1993. Very high rates of bedload sediment transport in desert ephemeral rivers. *Nature*, **366**, 148-150.

Laronne, J.B., Reid, I., Yitshak, Y. and Frostick, L.E. 1994. The non-layering of gravel streambeds under ephemeral flood regimes. *Journal of Hydrology*, **159**, 353-363.

Leopold, L.B. 1992. Sediment size that determines channel morphology. In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 297-311, John Wiley and Sons. Chichester, England.

Leopold, L.B. and Langbein, W.B. 1962. The concept of entropy in landscape evolution. *USGS Professional Paper*, **500A**.

Leopold, L.B. and Maddock, T. 1953. The hydraulic geometry of stream channels and some physiographic implications. *USGS Professional Paper*, **252**.

Li, Z. and Komar, P.D. 1986. Laboratory measurements of pivoting angles for applications to selective entrainment of gravel in a current. *Sedimentology*, **33**, 413-423.

Lisle, T.E. 1979. A sorting mechanism for a riffle-pool sequence. *Geological Society of America Bulletin*, **90**, 1142-1157.

Lisle, T.E. 1995. Particle size variations between bed load and bed material in natural gravel bed channels. *Water Resources Research*, **31**, 1107-1118.

Lisle, T.E. and Madej, M.A. 1992. Spatial variation in armouring in a channel with high sediment supply. In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 277-291, John Wiley and Sons. Chichester, England.

Lisle, T.E. and Hilton, S. 1999. Fine bed material in pools of natural gravel bed channels. *Water Resources Research*, **35**, 1291-1304.

Mackin, J.H. 1948. Concept of a graded river. *Geological Society of America Bulletin*, **59**, 463-512.

Martin, Y. 1992. Sediment budget from morphology: Vedder River, British Columbia. *Unpublished M.Sc. Thesis*, Department of Geography, University of British Columbia.

Martin, Y. and Church, M. 1995. Bed-material transport estimated from channel surveys: Vedder River, British Colombia. *Earth Surface Processes and Landforms*, **20**, 347-361.

McLean, D.G. 1980. Flood control and sediment transport study of the Vedder River. *Unpublished M.A.Sc Thesis*, Department of Civil Engineering, University of British Columbia.

Meade, R.H. 1985. Wavelike movement of bedload sediment, East Fork River, Wyoming. *Environ. Geol. Water Sci.*, **7**, 215-225.

Meyer-Peter, E. and Muller, R. 1948. Formulas for bed-load transport. *Proc. 2nd Cong. IAHR*, Stokholm, **2**, 39-64.

Milhous, R.T. 1973. Sediment transport in a gravel-bottomed stream. *Unpublished Ph.D. Thesis*, Department of Civil Engineering, Oregon State University.

Milhous, R.T. 1982. Discussion of Klingeman, P.C. and Emmett, W.W. Gravel bedload transport processes. In Hey, R.D., Bathurst, J.C. and Thorne, C.R. (eds). *Gravel Bed Rivers: Fluvial Processes*, 141-179, John Wiley and Sons. Chichester, England.

Montgomery, D.R., Panfil, M.S. and Hayes, S.K. 1999. Channel-bed mobility response to extreme sediment loading at Mount Pinatubo. *Geology*, **27**, 271-274.

Morisawa, M. 1985. *River forms and process*. Longman, New York.

Mulder, T. and Syvitski, J.P.M. 1996. Climatic and morphological relationships of rivers: implications of sea-level fluctuations on river loads. *Journal of Geology*, **104**, 509-523.

Naden, P.S. 1988. Models of sediment transport in natural streams. In Anderson, M.G. (ed). *Modelling Geomorphological Systems*, 217-258, John Wiley and Sons. London, England.

Nicholas, A.P., Ashworth, P.J., Kirkby, M.J., Macklin, M.G. and Murray, T. 1995. Sediment slugs: large-scale fluctuations in fluvial sediment transport rates and storage volumes. *Progress in Physical Geography*, **19**, 500-519.

Paintal, A.S. 1971. Concept of critical shear stress in loose open boundary channels. *Journal of Hydraulic Research*, **9**, 91-113.

Paola, C. and Seal, R. 1995. Grain size patchiness as a cause of selective deposition and downstream fining. *Water Resources Research*, **31**, 1395-1407.

Paola, C., Parker, G., Seal, R., Sinha, S.K., Southard, J.B. and Wilcock, P.R. 1992. Downstream fining by selective deposition in a laboratory flume. *Science*, **258**, 1757-1760.

Parker, G. 1990. Surface-based bedload transport relationship for gravel rivers. *Journal of Hydraulic Research*, **28**, 417-436.

Parker, G. 1991a. Selective sorting and abrasion of river gravels, I: theory. *Journal of Hydraulic Engineering*, **117**, 131-149.

Parker, G. 1991b. Selective sorting and abrasion of river gravels, II: applications. *Journal of Hydraulic Engineering*, **117**, 150-173.

Parker, G. 1998. Advances in sediment transport processes and fluvial geomorphology. In Klingeman, P.C., Beschta, R.L., Komar, P.D. and Bradley, J.B. (eds). *Gravel-Bed Rivers and the Environment*, 7-14, Water Resources Publications, Highlands Ranch, Colorado.

Parker, G. and Cui, Y.T. 1998. The arrested gravel front: stable gravel sand transitions in rivers - part 1: simplified analytical solution. *Journal of Hydraulic Research*, **36**, 75-100.

Parker, G. and Klingeman, P.C. 1982. On why gravel bed streams are paved. *Water Resources Research*, **18**, 1409-1423.

Parker, G. and Sutherland, A.J. 1990. Fluvial armor. *Journal of Hydraulic Research*, **28**, 529-543.

Parker, G., Klingeman, P.C. and McLean, D.G. 1982. Bedload and size distribution in paved gravel-bed streams. *Journal of Hydraulics Division, ASCE*, **108**, 544-571.

Peloutier, V., Hoey, T.B. and Herbertson, J.G. 1997. Experimental study of fine sediment infiltration into porous media. In Watt, J. (ed), *Proceedings, 3rd International Conference on River Flood Hydraulics*, 463-472.

Petrie, J. and Diplas, P. 2000. Statistical approach to sediment sampling accuracy. *Water Resources Research*, **36**, 597-605.

Pickup, G. 1984. Geomorphology of tropical rivers, 1: landforms, hydrology and sedimentation in the Fly and Lower Purari, Papau New Guinea. *Catena Supplement*, **5**, 1-17.

Pizzuto, J.E. 1984. Bank erodibility of shallow sandbed streams. *Earth Surface Processes and Landforms*, **9**, 113-124.

Pizzuto, J.E. 1992. The morphology of graded gravel rivers: a network perspective. *Geomorphology*, **5**, 457-474.

Pizzuto, J.E. 1995. Downstream fining in a network of gravel-bedded rivers. *Water Resources Research*, **31**, 753-759.

Rana, S.A., Simons, D.B. and Mahmood, K. 1973. Analysis of sediment sorting in alluvial channels. *Journal of Hydraulics Division,* ASCE, **99**, 1967-1980.

Reid, I. and Frostick, L.E. 1985. Role of settling, entrainment and dispersive equivalence and of interstice trapping in placer formation. *Journal of the Geological Society*, **142**, 739-746.

Reid, I. Frostick, L.E. and Brayshaw, A.C. 1992. Microform roughness elements and the selective entrainment and entrapment of particles in gravel-bed rivers. In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 243-249, John Wiley and Sons. Chichester, England.

Reid, I. and Laronne, J.B. 1995. Bedload sediment transport in an ephemeral stream and a comparison with seasonal and perennial counterparts. *Water Resources Research*, **31**, 773-781.

Reid, I., Bathurst, J.C., Carling, P.A., Walling, D.E. and Webb, B.W. 1997. Sediment erosion, transport and deposition. In Thorne, C.R., Hey, R.D. and Newson, M.D. (eds). *Applied Fluvial Geomorphology for River Engineering and Management*, 95-135, John Wiley and Sons. Chichester, England.

Rice, S. 1998. Which tributaries disrupt downstream fining along gravel-bed rivers? *Geomorphology*, **22**, 39-56.

Rice, S. 1999. The nature and controls on downstream fining within sedimentary links. *Journal of Sedimentary Research*, **69**, 32-39.

Rice, S. and Church, M. 1998. Grain size along two gravel-bed rivers: statistical variation, spatial pattern and sedimentary links. *Earth Surface Processes and Landforms*, **23**, 345-363.

Rinaldo, A. 1999. Hydraulic networks in nature. *Journal of Hydraulic Research*, **37**, 847-859.

Rhoads, B.L. 1989. Longitudinal variation in the size and sorting of bed material along six arid-region mountain streams. *Catena Supplement*, **14**, 38-45.

Robinson, R.A.J. and Slingerland, R.L. 1998. Origin of fluvial grain-size trends in a foreland basin: the Pocono-Formation of the central Appalachian Basin. *Journal of Sedimentary Research*, **68**, 39-56.

Russell, R.J. 1968. Where most grains of very coarse sand and fine gravel are deposited. *Sedimentology*, **11**, 31-38.

Sambrook Smith, G.H. 1994. The gravel-sand transition along river channels. *Unpublished Ph.D. Thesis*, Department of Geography, University of Sheffield.

Sambrook Smith, G.H. 1996. Bimodal fluvial bed sediments: origin, spatial extent and processes. *Progress in Physical Geography*, **20**, 402-417.

Sambrook Smith, G.H. and Ferguson, R.I. 1995. The gravel-sand transition along river channels. *Journal of Sedimentary Research*, **A65**, 423-430.

Sambrook Smith, G.H. and Ferguson, R.I. 1996. The gravel-sand transition: flume study of channel response to reduced slope. *Geomorphology*, **16**, 147-159.

Sambrook Smith, G.H., Nicholas, A.P. and Ferguson, R.I. 1997. Measuring and defining bimodal sediments: problems and implications. *Water Resources Research*, **33**, 1179-1185.

Schroeder, R. 1991. Test of Hack's slope to bed material relationship in the Southern Eifel Uplands, Germany. *Earth Surface Processes and Landforms*, **16**, 731-736.

Schumm, S.A. and Stevens, M.A. 1973. Abrasion in place: a mechanism for rounding and size reduction of coarse sediments in rivers. *Geology*, **1**, 37-40.

Seal, R., Paola, C., Parker, G., Southard, J.B. and Wilcock, P.R. 1997. Experiments on downstream fining of gravel: 1. narrow channel runs. *Journal of Hydraulic Engineering*, **123**, 874-884.

Seal, R., Toro-Escobar, C., Cui, Y., Paola, C., Parker, G., Southard, J.B. and Wilcock, P.R. 1998. Downstream fining by selective deposition: theory, laboratory and field observations. In Klingeman, P.C., Beschta, R.L., Komar, P.D. and Bradley, J.B. (eds). *Gravel-Bed Rivers and the Environment*, 85-114, Water Resources Publications, Highlands Ranch, Colorado.

Sear, D.A. 1993. Fine sediment infiltration into gravel spawning beds within a regulated river experiencing floods: ecological implications for salmonids. *Regulated Rivers: Research and Management*, **8**, 373-390.

Sear, D.A. 1996. Sediment transport processes in pool-riffle sequences. *Earth Surface Processes and Landforms*, **21**, 241-262.

Shaw, J. and Kellerhals, R. 1982. The composition of recent alluvial gravels in Alberta river beds. *Alberta Research Council Bulletin*, **41**, Calgary.

Shea, J.A. 1974. Deficiencies of clastic particles of certain sizes. *Journal of Sedimentary Petrology*, **44**, 985-1003.

Shields, A. 1936. Anwendung der Ahnlichkeitsmechanik und der Turbulenz-forschung auf die Geschiebebewegung. *Miteil. dur Preussischen Versuch. fur Wasserbau u. Schiffbau*, Berlin.

Shulits, S. 1941. Rational equation of river bed profile. *EOS, Transactions of the American Geophysical Union*, **22**, 622-630.

Simons, D.B. and Simons, R.K. 1987. Differences between gravel- and sand-bed rivers. In Thorne, C.R., Bathurst, J.C. and Hey, R.D. (eds). *Sediment Transport in Gravel Bed Rivers*, 3-15, John Wiley and Sons. Chichester, England.

Sinclair, F.N. 1961. A history of the Sumas drainage, dyking and development district. *Chilliwack Historical Society*, Chilliwack, British Columbia.

Sinha, S.K. and Parker, G. 1996. Causes of concavity in longitudinal profiles of rivers. *Water Resources Research*, **32**, 1417-1428.

Sissons, J.B. 1974. Late-glacial ice cap in the central Grampians, Scotland. *Transactions of the Institute of British Geographers*, **62**, 95-114.

Snow, R.S. and Slingerland, R.L. 1987. Mathematical modelling of graded river profiles. *Journal of Geology*, **95**, 15-33.

Sternberg, H. 1875. Untersuchungen uber Langen und Querprofil geschiebefuhrender Flusse. *Z. fur Bauwesen*, **25**, 483-506.

Sutherland, A.J. 1987. Static armour layers by selective erosion. In Thorne, C.R., Bathurst, J.C. and Hey, R.D. (eds). *Sediment Transport in Gravel Bed Rivers*, 243-260, John Wiley and Sons. Chichester, England.

Tait, S.J., Willetts, B.B. and Maizels, J.K. 1992. Laboratory observations of bed armouring and changes in bedload composition. In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 205-225, John Wiley and Sons. Chichester, England.

Talling, P.J. 2000. Self-organisation of river networks to threshold states. *Water Resources Research*, **36**, 1119-1128.

Thompson, D.M., Wohl, E.E. and Jarrett, R.D. 1996. Revised velocity-reversal and sediment-sorting model for a high-gradient, pool-riffle stream. *Physical Geography*, **17**, 142-156.

Toro-Escobar, C.M., Parker, G. and Paola, C. 1996. Transfer function for the deposition of poorly sorted gravel in response to streambed aggradation. *Journal of Hydraulic Research*, **34**, 35-53.

Tucker, G.E. and Slingerland, R.L. 1997. Drainage basin responses to climate change. *Water Resources Research*, **33**, 2031-2047.

Van Niekerk, A., Vogel, K.R., Slingerland, R.L. and Bridge, J.S. 1992. Routing of heterogeneous sediments over movable bed: model development. *Journal of Hydraulic Engineering*, ASCE, **118**, 246-261.

Vogel, K.R., Van Niekerk, A., Slingerland, R.L. and Bridge, J.S. 1992. Routing of heterogeneous size-density sediments over movable stream bed: model verification and testing. *Journal of Hydraulic Engineering*, ASCE, **118**, 263-279.

Wathen, S.J., Ferguson, R.I., Hoey, T.B. and Werritty, A. 1995. Unequal mobility of gravel and sand in weakly bimodal river sediments. *Water Resources Research*, **31**, 2087-2096.

Werritty, A. 1992. Downstream fining in a gravel-bed river in southern Poland: lithologic controls and the role of abrasion. In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 333-350, John Wiley and Sons. Chichester, England.

Whiting, P.J., Dietrich, W.E., Leopold, L.B., Drake, T.G. and Shreve, R.L. 1988. Bedload sheets in heterogeneous sediment. *Geology*, **16**, 105-108.

Wiberg, P.L. and Smith, J.D. 1987. Calculations of the critical shear stress for motion of uniform and heterogeneous sediments. *Water Resources Research*, **23**, 1471-1480.

Wilcock, P.R. 1992. Experimental investigation of the effect of mixture properties on transport dynamics. In Billi, P., Hey, R.D., Thorne, C.R. and Taconni, P. (eds). *Dynamics of Gravel Bed Rivers*, 109-139, John Wiley and Sons. Chichester, England.

Wilcock, P.R. 1993. Critical shear stress of natural sediments. *Journal of Hydraulic Engineering*, **119**, 491-505.

Wilcock, P.R. 1997. Entrainment, displacement and transport of tracer gravels. *Earth Surface Processes and Landforms*, **22**, 1125-1138.

Wilcock, P.R. 1998. Two-fraction model of initial sediment motion in gravel-bed rivers. *Science*, **280**, 410-412.

Wilcock, P.R. and McArdell, B.W. 1993. Surface-based fractional transport rates: mobilization thresholds and partial transport of a sand-gravel sediment. *Water Resources Research*, **29**, 1297-1312.

Wilcock, P.R. and McArdell, B.W. 1997. Partial transport of a sand/gravel sediment. *Water Resources Research*, **33**, 235-245.

Wilcock, P.R., Barta, A.F., Shea, C.C., Kondolf, G.M., Matthews, W.V.G. and Pitlick, J. 1996. Observations of flow and sediment entrainment on a large gravel-bed river. *Water Resources Research*, **32**, 2897-2909.

Wolcott, J. 1988. Nonfluvial control of bimodal grain-size distributions in river-bed gravels. *Journal of Sedimentary Petrology*, **58**, 979-984.

Yatsu, E. 1955. On the longitudinal profile of the graded river. *Transactions of the American Geophysical Union*, **36**, 655-663.