ABSTRACT

Contemporary and Historic River Channel Change at Swinhope Burn, Weardale:
A Study of River Response to Flood Events in an Upland, Gravel-Bed Stream
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The response of upland gravel-bed streams to floods has long been associated with high levels of erosion and deposition, often resulting in major changes in channel form. This largely stems from observations of upland channels immediately following large floods. The aim of this thesis is to identify the importance of flood events in contemporary and historic river channel change in an upland gravel-bed stream.

Swinhope Burn, in upper Weardale, Northern Pennines is used as a basis for this study. An assessment of channel planform change over the historical period was made using historical maps and air photographs. The study reach has retained a stable meandering pattern over a period of 180 years, with a temporary but dramatic change to a straight, low sinuosity, partly braided channel identified in the 1844 Tithe Map. The probable cause is an increase in coarse sediment supply generated by floods in the 1820's and upstream mining activities.

The passage of a major flood through Swinhope Burn in February, 1997 produced very little channel change with erosion being the same order of magnitude as deposition, indicating that the study reach is stable even during overbank flows. A sediment tracing experiment designed to assess the importance of sediment exchanges between the bed and lateral inputs from eroding banks and bluffs demonstrates the importance of within channel movements and the minor local influence of lateral sediment sources.

This study shows that contemporary channel response to flood events is through vertical rather than lateral adjustment in channel form which is substantiated by channel planform stability over the historical period. Long-term channel stability is attributed to the presence of a local base-level imposed by the Greenly Hills moraine, which has resulted in a low channel gradient which inhibits coarse bedload transport and frequent, major channel change.

CONTEMPORARY AND HISTORIC RIVER CHANNEL CHANGE AT SWINHOPE BURN, WEARDALE: A STUDY OF RIVER RESPONSE TO FLOOD EVENTS IN AN UPLAND, GRAVEL-BED STREAM

by

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This is to certify that none of the material offered has previously been submitted by me for a degree in this or any other University.

Signature ....M. Danks..........

Date ..22/12/98..

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CHAPTER 1

INTRODUCTION

1.1 Overview

Upland gravel-bed streams have long been characterised by their powerful and dynamic nature which has been associated with very high rates of erosion, transport and deposition (Acreman, 1983; Ferguson and Werritty, 1983; Carling, 1989). Since upland catchments are often small with high gradients, particularly in the headwaters, the effects of a flood event can be catastrophic in terms of river channel change and bedload transport (Carling, 1986; Harvey, 1986).

Although a number of studies have documented the response of upland streams to rare, high magnitude floods (Newson, 1980; Coxon et al, 1989; McEwen, 1989a), there has been a lack of studies monitoring channel response to more frequent events such as the mean annual flood. Important exceptions include Milne (1982b, 1983a) and Ferguson and Werritty (1983) in terms of the adjustment of cross-sectional form, and Carling (1989), Ashworth and Ferguson (1989) and Carling and Hurley (1987) in terms of rates of bedload transport. Very few studies combine observations of both historical and contemporary channel change which is critical if present levels of geomorphic activity in upland basins are to be set in a broader context (Werritty and Ferguson, 1980). The continuing debate over the significance of rare, high magnitude floods in comparison with the lower magnitude mean annual flood in causing river channel change will only be resolved if contemporary floods are put in a historical perspective and insight into the nature and rates of post-flood recovery is gained (Anderson and Calver, 1977, 1980). This point is highlighted by Newson (1989) who identifies a need 'to record the impacts of contemporary floods and to continuously review those of historical floods'. There is also continuing controversy over the interaction between climatic fluctuations, land-use change, and the hydrological regime of upland gravel-bed streams in governing river channel change over the historical period. In the Northern Pennines, probably the most important change in land-use has been historic metal mining. Although a number of studies have examined the role of historic metal mining on river channel change

(Lewin et al, 1977, 1983), fewer have examined the combined influence of historic floods and metal mining on channel metamorphosis (Macklin, 1986).

This lack of research into the role of flood events in determining both contemporary and historic river channel change in the British uplands provides justification for this thesis. In particular, the assumption that upland gravel-bed streams are highly dynamic in terms of channel change and bedload movement has been hitherto largely unchallenged. This is not surprising since observations of upland streams following flood events is likely to lead to an overestimation of the role of high magnitude flows in causing channel changes. In fact, the initial motivation for examining the role of flood events in contemporary and historical river channel change in Swinhope Burn, upper Weardale, originated from observing the Swinhope Burn basin immediately following a high magnitude flood event which affected the basin in January, 1995. The flood had a peak flow of 287 m^3s^{-1} recorded on the River Wear, of which Swinhope Burn is a tributary. The January 1995 flood to date is the largest flood on record and led to extensive bank undercutting, deposition of large cobbles on the outside of meander bends and considerable structural damage to fences and walls bounding the channel within the Swinhope Burn basin. In general, the study reach gives the impression of a highly dynamic upland stream in terms of both the throughput of coarse sediment and in terms of numerous lateral inputs of coarse sediment from cut-banks where the stream impinges on the valley-side slopes. In order to determine whether Swinhope Burn is a dynamic gravel-bed stream characterised by high rates of erosion, transport and deposition, post-flood changes in channel form and grain-size of bed material need to be monitored and the fluvial sediment dynamics characterised.

Swinhope Burn was chosen as the site at which to examine the role of flood events in contemporary and historic river channel change for a number of reasons. Firstly, Swinhope Burn gives the impression of a dynamic upland stream, with many lateral sediment inputs from eroding and cut banks and an active throughput of sediment. The meandering channel pattern, particularly in the lower reaches, provides an opportunity to examine the relationship between channel sinuosity and erosion and deposition during flood events. The distinct step in the middle reaches of the long profile of Swinhope Burn, formed by the partial closure of the valley system by the Greenly Hills moraine, allows the effects of local base-level control on sediment dynamics and channel form to be determined. Since Swinhope Bottoms is essentially a zone of sedimentation, where coarse sediment entering the reach becomes trapped within the channel, it is potentially a site at which the long-term sediment record can be examined. This setting also provides an opportunity to investigate the significance of glacial heritage in determining the sensitivity of historic and contemporary channel response to flood events. All these points provide justification for choosing Swinhope Burn as an interesting study site, despite its relatively low channel gradient.

1.2 Aims and objectives

The aim of this thesis is to identify the importance of flood events in contemporary and historic river channel change in Swinhope Burn, Weardale. In order to test the hypothesis that floods have a significant role to play in both historic and contemporary river channel change, analysis of historical maps and documents over a period of 180 years and field observations over a period of two years form the basis of the research.

The first objective is to identify the role of flood events in determining the nature and extent of historic river planform change along a 1.4 km reach of Swinhope Burn from 1815 to the present, using historical maps and air photographs. The probable cause of river planform change is determined using field evidence and local historic flood documentation. The second objective is to examine the response of an upland stream to floods in terms of changes in channel form and the mean grain-size of the bed material. The final secondary objective is to examine within reach and through reach sediment dynamics during flood events to determine whether Swinhope Burn has a dynamic bed with active throughput of sediment or is predominantly stable.

1.3 Thesis structure

The thesis is organised into eight chapters. Before the main body of the project, relevant literature is reviewed in Chapter 2, which provides the context for the present study. Chapter 3 describes the study site, a 1.4 km reach of Swinhope Burn, a right bank tributary of the River Wear, upper Weardale, County Durham and details the field methods and techniques used in the study. Chapter 4 identifies the nature and extent of historic river channel change in Swinhope Burn and seeks to determine possible causes of observed channel planform change using historic flood documentation and field evidence. The structure of the stream is described in Chapter 5 with reference to the influence of bank cohesion, bed topography and channel gradient on changes in the cross-sectional geometry and planform of the study reach. Chapter 6 identifies downstream changes in the cross-sectional form of the channel and mean-grain size of bed material following the passage of a large flood event through the Swinhope Burn basin on 19th and 20th February, 1997. This information is used to test the hypothesis that Swinhope Burn is an active gravel-bed channel in which major channel change occurs in response to flooding. The results of a small sediment tracing experiment, using painted and magnetic tracer pebbles, designed in order to identify within reach and through reach sediment dynamics are discussed in Chapter 7. Finally Chapter 8 provides a summary of the thesis, outlines the major conclusions and suggests future research.

CHAPTER 2

LITERATURE REVIEW

2.1 Introduction

The objective of this chapter is to review literature which examines the role of flood events in historic and contemporary river channel change in U.K. upland gravel-bed streams particularly with reference to studies in the Northern Pennines (Figure 2.1).

Research over the past 30 years has shown that one of the major causes of river channel change in upland, gravel-bed streams is the occurrence of flood events. However, there is continuing debate over whether 'rare' high magnitude floods are responsible for the observed, sometimes irreversible, river channel change (Anderson and Calver, 1977; Newson, 1980; Harvey, 1986) or whether the 'annual' flood is geomorphologically of greater significance (Hitchcock, 1977; Carling, 1988). The rate of post-flood recovery of upland streams is critical to this debate, since if channel recovery from a high magnitude flood is rapid, due to channel reworking by lesser, more frequent flows, the geomorphic significance of a 'rare' flood is less (Werritty, 1982; Carling, 1986; Coxon et al, 1989).

Since very large floods tend to occur relatively infrequently and because post-flood recovery often takes many years, the significance of these rare events can only be determined from a knowledge of historic channel planform change within a basin. Accordingly, channel planform change in response to floods of varying magnitude and frequency over the historical period is a continuing research topic in the British uplands. The utility of historical maps and air photographs in identifying the nature and extent of changes in channel planform over timescales ranging from upwards of 300 years (Lewin, 1987), 200 years (Werritty and Ferguson, 1980; Macklin, 1986) and 30 years (Werritty, 1982) has been demonstrated. The development of flood dating techniques has made it possible to identify the cause of observed changes in channel planform and aid the reconstruction of flood histories in a basin (Rumsby and Macklin, 1994).

Figure 2.1 Map of the U.K. showing research sites mentioned in the literature

An understanding of the response of upland, gravel-bed streams to flood events also requires a detailed knowledge of contemporary channel processes during high flow events. Consequently, sediment dynamics within coarse-bedded channels during flood events (Carling, 1989; Ashworth and Ferguson, 1989; Hoey, 1992) and the morphological adjustment of the channel in response to flood events (Ferguson and Werritty, 1983; Carling, 1986; Harvey, 1991) are well-documented for the British uplands. Improvements in measuring bedload transport, and in particular, pebble tracing techniques have resulted in a much clearer understanding of sediment dynamics in upland gravel-bed streams (Carling, 1987; Ferguson et al, 1996).

However, in order to explain both contemporary and historic river channel response to flood events, the interaction between changes in land-use, climatic fluctuations and the hydrological regime of upland gravel-bed streams also needs to be investigated (McEwen, 1989a; Macklin et al, 1992; Rumsby and Macklin, 1994). Probably, the most important change in land-use in the British uplands, and particularly in the Northern Pennines over the past 300 years has been the onset of historic metal mining which introduced large amounts of coarse sediment and fine metaliferous waste to many headwater streams in the Northern Pennines, which resulted in widespread channel aggradation and impaired growth of riparian vegetation (Aspinall et al, 1986; Rumsby and Macklin, 1994).

2.2 Scope of Chapter

This chapter firstly examines the relative magnitude and frequency of flood events in upland gravel-bed streams and their geomorphic significance. The role of flood events in historical and contemporary river channel change is assessed with particular reference to the morphological response of upland stream channels to floods and bedload transport during floods. Finally, the role of climate and land-use change in determining flood regime and river channel change is discussed.

2.3 The geomorphic significance of floods - the question of magnitude and frequency

The concept of the relative effectiveness of infrequent, high magnitude floods and smaller more frequent events coupled with the significance of the post-flood recovery period is a recurring research theme (Wolman and Miller, 1960; Anderson and Calver, 1977, 1980). This idea is appropriate to the British uplands where flood events, often occurring as a result of snowmelt coupled with heavy rainfall within small, steep, river basins, can result in catastrophic geomorphic change. Pioneering research by Anderson and Calver (1977) identifies three requirements for the qualitative assessment of the effects of a flood: a record of the pre-flood landscape, monitoring during the passage of a flood and subsequent moderation of features produced. Similarly, Newson (1980) suggests that studies of flood effectiveness require regular surveys throughout the recovery period following major flooding.

Although it is difficult to generalise on the importance of the magnitude and frequency of flood events to river channel change, it appears that, with the exceptions of Anderson and Calver (1977) and Wolman and Gerson (1978), since the early 1980's the prevailing view has been that it is events of high magnitude and low frequency that are most important in the evolution of upland river channels and their floodplains. This may reflect the high incidence of major floods recorded during this period, for example, flooding on the Ardessie Burn, Wester Ross, Scotland (Acreman, 1983), the Noon Hill flash floods in upper Teesdale (Carling, 1986), the major flood on the Yellow River in Ireland (Coxon et al, 1989) and in the Howgill Fells (Harvey, 1986). However, the scarcity of meteorological and hydrological information for most upland areas has generally led to an underestimation of the frequency of flooding, particularly where flooding is localised and results from convectional summer storms (Newson, 1980). For example, in the Northern Pennines, where there is a lack of detailed rainfall and flood data for isolated headwater catchments, the geomorphic effects of severe thunderstorms and resulting flood events are poorly documented, with the notable exception of Carling (1986). This is despite a long history of flooding in the Pennine uplands (Archer, 1992).

This section reviews literature over the past four decades in which the development of ideas relating to the importance of the magnitude and frequency of flood events in controlling river channel change is examined, using examples from the British uplands. Key factors influencing channel response to floods of varying magnitude are identified as sediment supply, the sequence of flood events and antecedent rainfall conditions. The nature and extent of post-flood recovery of upland streams following flood events is detailed and its significance in relation to the magnitude and frequency concept is considered.

2.3.1 The significance of high magnitude/low frequency flood events

The effects of catastrophic flooding resulting from high magnitude and low frequency events have been well documented for the British uplands with return intervals of 100 years (Harvey, 1986, 1991), 150 years (Anderson and Calver, 1977, 1980), and 1000 years (Newson, 1980; McEwen, 1989a). These authors share the view that it is these rare flood events which are responsible for the major changes to river channels and floodplains in these areas.

Wolman and Gerson (1978) suggest that extremely rare floods with recurrence intervals of 500 years or more are likely to result in irreversible changes in hillslopes and river channels. However, Newson (1980) reviews the concept of the geomorphic effectiveness of floods proposed by Wolman and Gerson (1978) detailing two rare flood events which occurred in the small upland Plynlimon experimental catchment in mid-Wales during August 1973 and August 1977. The distinction between geomorphic 'effectiveness' of floods and geomorphic 'work' performed by floods, popularised by Wolman and Gerson (1978) are both demonstrated in the Plynlimon example. The geomorphic 'effectiveness' of these floods was determined using a series of photographs taken during the post-flood period and 'work' achieved by these floods was determined using bedload traps. Newson (1980) demonstrated that although the first Plynlimon flood was more 'effective' on hillslopes, the second was more 'effective' within the channel, despite peak discharges of similar return periods. The Ardessie flood, in Wester Ross, Scotland Highlands (Acreman, 1983) was similar to the floods recorded by Newson (1980) in mid-Wales in terms of sediment transport, however, it was both a 'channel' and a 'slope' flood. It was estimated that the flood transported approximately 14 years of non-dissolved sediment output, and led to widespread deposition in the form of several large boulder and coarse gravel bars and an alluvial fan.

Carling (1986) documents the catastrophic effects of the Noon Hill flash flood of July, 1983 which centered on the sparsely populated watersheds between Teesdale and Weardale in the Northern Pennines. This flood occurred 4 km west of Swinhope Burn basin, on which the present study is centered. The flood was of high magnitude and had an estimated return interval of between eight and twenty years. The flood was extremely abrupt with streamflow levels rising from base flow to peak flow in less than 15 minutes with the entire discharge being generated over a very small area. The geomorphic effects on the channels and slopes bounding the channel was severe. Similarly, Coxon et al (1989) document the effects of a flood in a small, steep, upland catchment in County Leitrim, Ireland which had a recurrence interval on the order of hundreds of years. However, whereas the Yellow River has a catchment area of 14.8 km^2 the three catchments affected by the Noon Hill flash floods in the Northern Pennines (Carling, 1986) had considerably smaller catchments ranging between 1.86 km^2 in West Grain to 7.14 km^2 on Ireshope Burn. Consequently it is difficult to compare the geomorphic effects of the Noon Hill flash flood and the flash flood on the Yellow River in terms of the recurrence interval of each flood. However, major channel changes described for the Yellow River (Coxon et al, 1989) are very similar to those recorded by Carling (1986) in the Northern Pennines, Anderson and Calver (1980) on Exmoor and Harvey (1991) in the Howgill Fells.

Harvey (1986) describes a flash flood which resulted from an intense convectional storm in the headwaters of the River Lune in the Howgill Fells in June, 1982. This provided an ideal opportunity to assess the importance of flood events with a recurrence interval of 100 years in landform adjustment. Since the streams described by Harvey (1986, 1991) have catchment areas ranging from 0.89 km^2 to 8.26 km^2 these basins are of comparable size to those described by Carling (1986) on the watershed between Weardale and Teesdale. However, the flood recurrence interval of 100 years is much greater than the flood described by Carling (1986) which had an estimated recurrence interval of 8-20 years. The massive input of coarse sediment generated by the 100 year flood coupled with high flood peaks led to the development of unstable, wide, braided channels, replacing the former narrow, meandering channel. Air photographs indicated that the flood of June, 1982 brought about more channel change than during the whole of the previous period since 1948. However, Harvey (1991) concludes that in the Howgill Fells, the fluvial system showed adjustment of channel morphology to hillslope sediment supply in two contexts, firstly with regard to fairly frequent events with a return period of one to five years and secondly, major floods.

2.3.2 The significance of low magnitude/high frequency flood events

In contrast, low magnitude, high frequency floods (e.g. ten times a year, Hitchcock, 1977 and two to four times a year, Harvey, 1982) have also been shown to be important in controlling river channel change. Harvey's (1977) research in the Howgill Fells was one of the first studies to examine the effect of the magnitude and frequency of major discharges on small, upland gravel-bed streams. This was in contrast with earlier research such as Wolman and Miller (1960) who dealt exclusively with lowland streams or large rivers. Harvey (1977) claimed that it was events of moderate magnitude and frequency resulting from an annual flood which were capable of carrying out the most geomorphic 'work' and controlling the morphology of both gullied slopes and stream channels in an upland environment. Observations at Grains Gill, a tributary of Carlingill in the Howgill Fells (Harvey, 1977), showed that coarse sediment builds up in debris cones at the base of gullies. This sediment was removed by floods which occurred approximately once a year, resulting in major channel change in the main valley. Likewise, Harvey et al (1982), studied three different fluvial environments in upland north-west Britain ranging from the steep head waters of Grains Gill, the braided Langden Brook and the meandering River Cound. They concluded that flood events which occurred between two and four times a year and once every two years were dominant in controlling river channel change, but that more frequently occurring events caused minor adjustments to channel morphology.

Similarly, the development of channel pattern in upland streams in response to flood events with a return interval of 0.9 years is demonstrated for Langden Brook, in the Forest of Bowland (Hitchcock, 1977) and Great Eggleshope Beck, in the Northern Pennines (Carling, 1988). The development of channel pattern in response to floods with a return interval of one year is demonstrated for the River Feshie in the

Cairngorms (Werritty and Ferguson, 1980). Hitchcock (1977) and Werritty and Ferguson (1980) both identify stream re-working of coarse superficial deposits as a key factor in determining channel response to the annual flood. Carling (1988), confirmed the 'dominant discharge concept' for steady-state alluvial channels, arguing that although large discharges individually transport more sediment than flows which occur once or twice a year, they occur too infrequently to be of overall significance in sediment transport and river channel change. However, discharges in excess of bankfull are required for extensive channel change to occur (Carling, 1987). Carling (1987) argues that bankfull discharge which has a recurrence interval of slightly less than one year, is the effective discharge in small upland gravel-bed streams unless the stream is unable to adjust its form freely. For example, in Carl Beck in the Northern Pennines, form adjustment is constrained by a compacted bed and banks and even overbank flows are not adequate to alter channel form (Carling 1988). Similarly, detailed morphological data for eleven coarse bedload upland catchments in northern England and southern Scotland, suggests that the size of these coarse bedload channels and channel changes are related to 'relatively high but frequently occurring flows' (Milne 1983a).

Carling (1988) supports the original concept of Wolman and Miller (1960) that events of modest magnitude and high frequency are more important than larger, rarer events in transporting sediment but stresses that the Wolman-Miller model should be applied only to alluvial streams which are capable of adjusting their boundaries.

2.3.3 Post-flood recovery

The rate of post-flood recovery of upland streams is critical to the debate on the significance of magnitude and frequency of flooding (Carling, 1986; Coxon et al, 1989). If post-flood recovery is relatively rapid and widespread (Werritty, 1982; Carling, 1986), the long-term geomorphic significance of the flood is less. Conversely, if post-flood recovery is less rapid or more localised (Anderson and Calver, 1977, 1980; Milne, 1982a) the geomorphic significance of the flood is greater regardless of the magnitude and frequency of the event.

Pioneering work by Wolman and Gerson (1978) suggests that in temperate regions, river channels are widened by floods with a recurrence interval of between 50 and 200 years but that the channels may undergo significant modification by subsequent smaller flows and recovery to their pre-flood state can occur over a timescale of years or even months. Carling (1986) acknowledges the persistence of large depositional features formed by the Noon Hill flash floods but equally stresses the rapidity of the re-adjustment of the channel to its pre-flood state. Carling (1988) describes the catchments in the Northern Pennines, which were affected by the Noon Hill flash flood as being 'insensitive systems that recover rapidly', within one tenth of the rare event recurrence interval.

Coxon et al (1989) record an equally catastrophic event on the Yellow River, but once again observes rapid post-flood recovery. Likewise, Newson (1980) describes post-flood recovery in the Plynlimon Experimental catchment, mid-Wales, as taking the form of undercutting of banks opposite point bars which were created by the flood and associated increased channel sinuosity. Similarly, Werritty and Ferguson (1980) identify braiding of river channels as the immediate response of channels to flood events with recovery of the system taking the form of an increased tendency for the channel to meander. More moderate flows aid the process of post-flood recovery by gradually eroding gravel shoals and boulder dumps (Newson, 1980) and re-working bar and bed material (Werritty and Ferguson. 1980). However, in the case of the gravelly River Feshie the process of post-flood recovery occurred over a period of 20 years. Werritty (1982) concludes from his observations of flash floods on the River Nethy, an upland gravel-bed stream in the Scottish Highlands, that smaller competent flows with a recurrence interval of around twice a year modify the erosional and depositional features produced by 'rare floods'.

Therefore, rapid post-flood recovery inevitably throws doubt on the actual geomorphic significance of major flood events for long-term river channel changes and highlights the importance of subsequent more frequent and moderate flow events and their ability to modify the effects of higher magnitude events.

Conversely, Anderson and Calver (1977) assess the persistence of landscape features formed by the August, 1952 flood on Exmoor, which had an estimated recurrence interval of 150 years. At the time of reinvestigation, 20 years after the flood event, it was still apparent that a large erosive event had taken place, and it was estimated that channel deepening may survive the mean recurrence interval of the flood. Milne (1982a) suggests that the effects of a large flood in 1931 on Harthope Burn, Cheviots, have persisted and former channel courses have not become vegetated owing to the presence of very coarse bedload material and a lack of overbank fines. Equally, Acreman (1983) suggests that following a rare flood event on the Ardessie Burn, recovery of this upland, coarse-grained stream may be largely limited due to the lack of bedload transport during low flows.

Harvey (1986) having documented the catastrophic flooding in the Howgills, reported five years later (Harvey, 1991), that post-flood recovery was primarily restricted by the available supply of sediment from erosional gullies adjacent to the stream which are only incorporated in the bedload of streams by floods with a return interval of two to five years. Variations in post-flood recovery of channels from one valley to another in the Howgill Fells were largely determined by sediment supply and catchment size. Harvey (1991) concludes that as long as there is an ample supply of sediment to the system, channels will remain unstable.

However, as Anderson and Calver (1977, 1980) point out, although the landforms produced by 'rare' flood events with return intervals in the order of hundreds of years do persist, the sequence of flood events and the time between such events is critical to river response and the development of the channel and adjacent valley slopes. This view is shared by Hitchcock (1977), Werritty (1982), Carling (1986), Coxon et al (1989) and Harvey (1986, 1991) from observations on other upland streams.

Examples of major flooding in British upland catchments described in the foregoing discussion, highlight the importance of intense rainfall associated with localised, convectional summer storms, since many of these floods occurred during the summer months (Anderson and Calver, 1977, 1980; Newson, 1980; Acreman, 1983; Carling, 1986; Harvey, 1986; Coxon et al, 1989). Since summer convective storms tend to be relatively short-lived and localised, their influence on flooding and channel changes in the British uplands may have been largely underestimated. However, in contrast, Knighton (1973) in his study of riverbank erosion in relation to streamflow conditions in the River Bollin-Dean argued that although summer storms had high rainfall intensities, they were of short duration and, unless they were of very high magnitude, caused little or no bank erosion. Although winter storms tended to be of lower magnitude they were more effective in eroding channel banks, because of increased bank wetting which reduced bank material resistance. Knighton (1973) concluded that channel form adjustment is related to the nature of a streams regime and particularly flow duration rather than the seasonality of flooding.

Also, the importance of sediment supply in determining whether or not a flood will be geomorphologically effective has been clearly demonstrated (Hitchcock, 1977; Anderson and Calver, 1980; Harvey, 1982; Acreman, 1983; McEwen, 1989a) with the coupling between slopes and channels being a particularly important factor (Newson, 1980; Harvey, 1986, 1991). Harvey (1991) argues that fairly frequent flood events of moderate magnitude are needed to incorporate sediment from erosional gullies into the bedload of streams, so that channel change can occur. This is likely to be determined by long-term climatic fluctuations which control the magnitude and frequency of storm events.

2.4 The role of flood events in historic river channel change

Direct evidence of flood events in Britain can be very difficult to determine, in part because the rapid growth of vegetation can obscure erosional and deposition features over time. Equally, when structural damage caused to walls, fences and buildings by a flood is repaired evidence of the extent of flooding is destroyed. However, historic maps, air photographs, and flood dating techniques can aid the reconstruction of flood histories within a basin.

The following review identifies a number of studies throughout the British uplands, which illustrate the importance of using historical maps and air photographs to determine the nature and extent of river channel change over timescales ranging from 30 (Werritty, 1982) to 200 years (Macklin, 1986; McEwen, 1989a). However, although maps and air photographs provide evidence of changes in channel planform over the historical period (Werritty and Ferguson, 1980; Macklin and Aspinall, 1986) flood dating techniques are required in order to identify particular floods which may have been responsible for the observed channel change.

The utility of historic maps and air photographs in determining historical channel planform change is discussed with particular reference to their advantages and limitations. Various dating techniques used to identify the cause of observed channel planform change are then identified. The section concludes with a brief review of the nature and causes of river planform change in the British uplands, with specific reference to the role of floods and mining activities on river channel change.

2.4.1. The use of historic maps and air photographs

Historic maps and air photographs have been used over the past few decades to identify changes in channel planform and floodplain sedimentation in upland gravel-bed streams (Ferguson and Werritty, 1980) and in piedmont zones (Mosley, 1975; Lewin, 1987) resulting from climatic fluctuations (McEwen, 1989b; Tipping, 1994), land-use change (Macklin, 1986), high magnitude flood events (Milne, 1982a) or a combination of all three (McEwen, 1989a).

Since national topographic mapping commenced around the 1860's and extensive air photography began in the 1940's studies of historic river channel change are generally limited to around 150 years. However, older maps, such as William Roy's Military Survey of Scotland (Werritty and Ferguson, 1980; McEwen, 1989) or Tithe and Estate maps (Hooke and Kain, 1982; Lewin, 1987) and enclosure plans may be available for a particular catchment. In cases where estates have been surveyed and mapped, changes in channel planform have been examined for upwards of 300 years (Lewin, 1987; Figure 2.2). However, where older source material is used, its reliability must be checked and if possible confirmed with other contemporary sources.

Difficulties experienced in comparing historical maps and air photographs over historical timescales for the Scottish braided River Feshie are outlined by Werritty and Ferguson (1980). Comparisons of 19th century maps, based on topographic survey with 20th century maps based on large-scale aerial photography can be problematical (Werritty and Ferguson, 1980; Milne, 1982a). A similar problem is experienced in interpreting and comparing air photographs over a 30 year period since photographs from different sources may be taken from different angles, some vertical and others

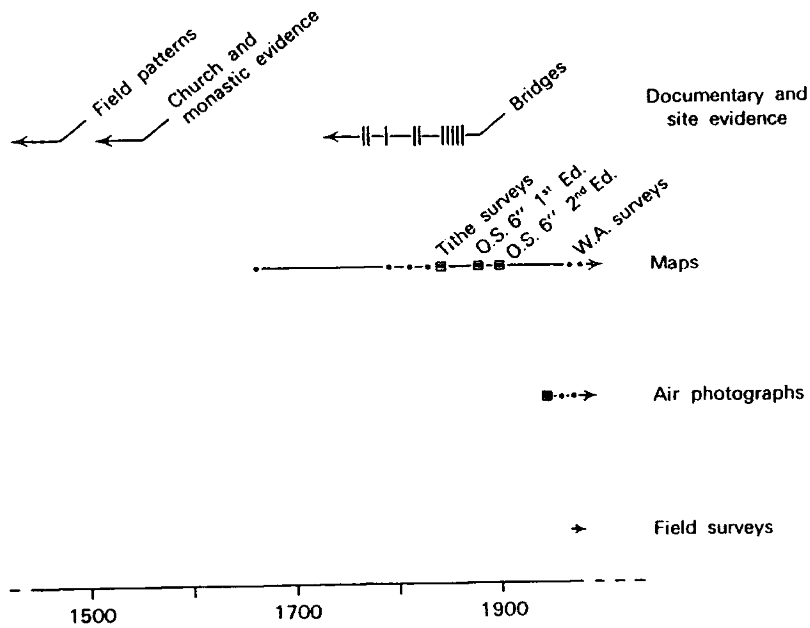


Figure 2.2 Historical sources for river channel changes in the Severn Basin (Lewin, 1987)

oblique (Werritty and Ferguson, 1980). Hooke and Kain (1982) identify specific problems in the use of maps for studies of river planform change which stem from factors such as the methods of survey used and the manner of representing the channel, for example, gravel bars are not always mapped. Hooke and Kain (1982) suggest that as much background as possible on survey methods and mapping conventions should be obtained.

However, it is possible, using maps and air photographs, to identify quite detailed changes in channel planform over a period of hundreds of years. If the coverage of maps and air photographs is detailed enough it is possible to identify the effects of, and the gradual recovery of a system from, a major flood event (Werritty and Ferguson, 1980; Milne, 1982a). Although maps and air photographs can identify changes in channel planform over the historical period dating techniques can identify the timing of the observed change (Milne, 1982a).

2.4.2 Dating techniques

Flood dating techniques depend largely upon the nature of information available for a particular catchment. For example, in the remoter areas of the British uplands, gauged discharge data and historic documentation relating to floods is likely to be relatively sparse (McEwen, 1994). However, in the Pennine and Welsh uplands, contamination of fluvial sediment with heavy metals from historic mining activities allows the use of the trace metal dating technique (Lewin et al, 1977, 1983; Macklin, 1997b). In many cases, a number of dating techniques are used in conjunction with one another. Extreme rainfall data (McEwen, 1989b) and continuous daily rainfall data (McEwen, 1989a; Rumsby and Macklin, 1994), when used in conjunction with flow data and historic flood documentation can aid the reconstruction of flood histories and provide an estimate of the return period of extreme flood events.

Since gauged flow data and rainfall records are not readily available for isolated upland catchments, historic flood documentation is often the only means by which a historic flood record can be established. Although flood documentation was available for the Tyne basin from 1700 onwards, Rumsby and Macklin (1994) also made a qualitative assessment of flood magnitude using relative stage height, often marked by floodstones, and by assessing the nature and extent of reported flood damage. Similarly, Macklin et al (1992b) used newspapers, local books and meteorological journals in conjunction with floodstones which marked the heights of the great floods of 1771 and 1815 in order to elucidate the flood history at Low Prudhoe on the River Tyne. However, documentary flood records are often biased in their tendency to record large overbank floods and tend to underrepresent the actual number of flood events (Macklin et al, 1992b) (Figure 2.3). McEwen (1987) warns that historic flood documentation is rather subjective and can be less reliable than some other sources as it tends to concentrate on the human impact of floods rather than the impact of floods on river channels and floodplains. Also it is often difficult to interpret the severity of a flood as different reports of the same flood event may be contradictory. Hooke and Kain (1982) argue that written sources are prone to exaggeration and that the lyrical style of some writing may distort the truth.

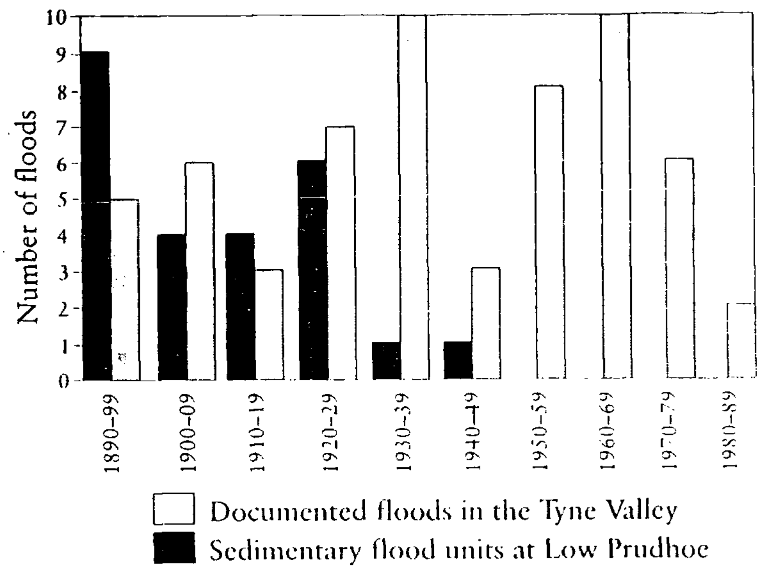


Figure 2.3 A comparison of floods documented in the lower Tyne valley and the sedimentary flood record at Low Prudhoe between 1890 and 1989 (Macklin et al, 1992b)

An alternative approach to an absence of gauged discharge data was used by Milne (1982a) on Harthope Burn, Cheviots, who estimated the size of flows generated within the basin by using daily discharge data for other adjacent gauged catchments. As one of the adjacent catchments received flow from Harthope Burn , it was used to identify the timing and magnitude of major floods which may have affected the study reach.

On streams where gauged discharge data are limited, the analysis of long-term and extreme rainfall can aid the reconstruction of the history of flood events (Rowling, 1989; Rumsby and Macklin, 1994), particularly when used in conjunction with historical maps and air photographs (McEwen 1989a; 1989b). McEwen (1989b) analysed rainfall records for the middle River Tweed in Scotland over the past 100 years and found that the most extreme rainfall totals corresponded with major flood events which then resulted in river channel change.

More precise dating of individual flood events can be derived using lichenometry to date flood deposits (McCarroll, 1994). This technique has been particularly useful when used in conjunction with maps, air photographs and flow data (Milne, 1982a; Macklin, 1986; McEwen, 1994) and in conjunction with analysis of the stratigraphic record of Holocene alluvial sequences to date flood deposits such a cobble-boulder bars, splays, berms and lobes (Macklin and Aspinall, 1986; Macklin et al, 1992a; Macklin et al, 1994; Macklin, 1997b). The analysis of alluvial stratigraphy is a relatively recent technique for reconstructing flood histories and was first used in Britain in the lower Tyne valley (Macklin et al, 1992b) and in the headwater streams of the Northern Pennines (Macklin et al, 1992a). This allows estimates to be made of the magnitude and frequency of rare and large floods through the analysis of fine-grained alluvial deposits. Since very high rates of vertical accretion are observed for the lower Tyne valley, it was possible for Macklin et al (1992b) to recognise and date individual historic flood events.

However, of particular importance in areas with a history of metal mining, such as the Northern Pennines, is trace metal analysis which is used for the relative dating of stratigraphic sequences (Lewin et al, 1977, 1983; Macklin and Dowsett, 1989; Macklin et al, 1992b; Passmore and Macklin, 1993; Macklin, 1997b). Since the history and development of metal mining is well-documented for the Northern Pennine Orefield (Raistrick and Jennings, 1965) trace metal analysis provides important dating markers within the stratigraphic sequence. For example, Macklin et al (1992) found that a rise in lead concentrations in the upper 90 cm of alluvium at Low Prudhoe, River Tyne basin, was in response to the revival of lead mining in the headwaters of the Tyne in the Northern Pennines.

Another useful relative dating technique is the assessment of soil characteristics such as maturity and the presence of buried and truncated profiles which helps to date the relative ages of erosional and depositional features formed by floods (Smith and Boardman, 1989). However a drawback associated with using this technique is that if flood events are closely spaced, it may be difficult to differentiate between them. In addition, gaps may exist in the record if a lower magnitude flood is followed by one of higher magnitude. Other field evidence such as dated bridges, and medieval ridge and furrow patterns can provide an indication of the extent of lateral channel migration (Lewin, 1987).

To summarise, since there is very little direct evidence of floods in the British uplands and since river channels and floodplains rapidly re-adjust to their pre-flood state, even after a major flood (Carling, 1986; Coxon et al, 1989) maps and air photographs are a vital source of evidence in determining the role of floods in river channel change. However, often the coverage of maps and air photographs is not comprehensive enough to provide information either on the frequency of flooding or post-flood recovery and it is often not possible to date a specific flood event which is responsible for the observed change in channel planform. This brief review of dating techniques has suggested that it is preferential to use a variety of techniques in order to identify the causes of observed river channel change and establish a historic flood chronology for a basin. Each technique has both benefits and drawbacks and they are best used in conjunction with one another (Milne, 1982a; McEwen, 1989a, 1989b; Macklin et al, 1992b; Rumsby and Macklin, 1994).

2.4.3 Historic channel planform change in the British uplands

Research suggests that historic channel planform change in the British uplands is quite widespread with almost 35% of all upland channels undergoing some degree of planform change between 1870 and 1950 (Hooke and Redmond, 1989). For example, the obliteration of a sinuous channel pattern and the formation of a fragmented braided channel pattern in response to a single high magnitude flood event has been identified for Dorback Burn, Cairngorms (Werritty, 1982), Harthope Burn, Cheviots (Milne, 1982a), Hoaroak Water, Exmoor (Anderson and Calver, 1980), and the River Severn at Welshpool (Lewin, 1987). A similar pattern of historic planform change has been reported for the River Dee, Aberdeenshire (McEwen, 1989a) which occurred in response to a more general increase in moderate to extreme flood events. Conversely, in the piedmont zone of the River Bollin, Cheshire, a series of large flood events during the period 1872 to 1973 resulted in channel widening and meander cut-off rather than the development of a braided channel pattern characteristic of coarse-bedded upland streams (Mosley, 1977).

Channel pattern changes associated with the downstream movement of sediment waves, generated by historic metal mining, have been documented for the Northern Pennines, using maps and air photographs (Macklin, 1986; Macklin, 1997a). Channel pattern changes resulting from inputs of coarse mining waste often takes the form of switching from a single thread meandering stream to a multi-thread braided pattern (Macklin, 1986). For example, active coarse sedimentation over the width of the valley floor of the River West Allen, Northern Pennines, with channel division and the formation of mid-channel bars were found to result from mining within an upland catchment (Macklin and Aspinall, 1986).

The effects of historic mining activities and high magnitude flood events can produce a similar change in channel planform from a meandering to a braided channel, due to inputs of coarse sediment. This makes it difficult to pinpoint the actual cause of river channel change (Lewin et al, 1983). Macklin (1986), however, argues that in the case of the River Nent, Blagill, Northern Pennines, the coincidence of metal mining and a series of large floods in the mid 1800's combined to produce a braided channel system which persisted for approximately 100 years before reverting to a single-thread meandering channel (Figure 2.4). Macklin (1997a) suggests that many rivers in the Northern Pennines experienced an increase in sediment supply to their valley floors and this was combined with a change in flood frequency in the latter half of the 19th century. A similar scenario is described for the River Ystwyth in upland mid-Wales (Lewin et al, 1977, 1983) where it was possible to conclude that channel planform had responded to a change in sediment supply. However, Lewin et al (1977) stress that the input of mining waste has a more localised effect on channel planform change than an input of coarse sediment generated by flood events.

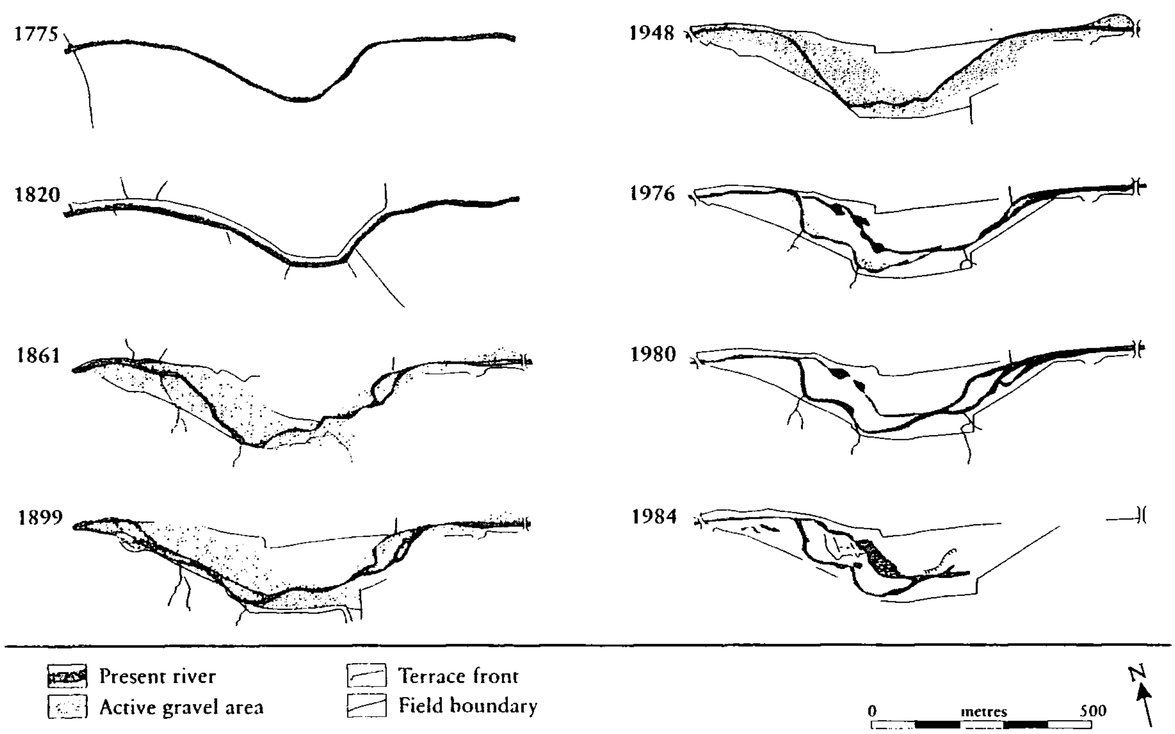


Figure 2.4 Channel change at Blagill, River Nent, 1775-1984 (Macklin, 1986)

Although historic channel planform change is widespread in the British uplands (Hooke and Redmond, 1989) there is evidence to suggest that whole streams, or particular reaches of streams, have remained relatively stable over many centuries. Werritty and Ferguson (1980) argue that over a time period of 200 years the River Feshie has retained distinctive channel patterns in some reaches as a result of spatial variations in valley topography. Similarly, McEwen (1989a) reports that over a period of 100 years, 69% of reaches sampled on the River Dee, Aberdeenshire had remained relatively stable. It appears that even though large flood events have occurred within a basin during the period of record, floods have had a very limited impact on the river channel (McEwen, 1989a; McEwen, 1994). Historical maps for Low Prudhoe on the river Tyne, revealed that there had been very little change in channel planform over the period of 130 years (Macklin et al 1992b). Likewise, Lewin (1987) found that meander bends had remained stable at Llandinam on the middle River Severn for a period of around 100 years.

Similarly, Warburton et al (1993), using map and air photograph evidence for the upland, braided Ashley River, New Zealand, found that channel planform had been remarkably stable over a period of 130 years.

However, McEwen (1994) in a historic planform reconstruction of the River Coe, Scottish Highlands suggests that there are marked temporal and spatial variations in channel stability and types of channel change. Lewin (1987) argues that spatial variety over short distances in both rates and patterns of historic planform change is to be expected, although he also suggests that in the case of the River Severn, the greater degree of lateral channel migration occurs in the piedmont zone because of the downstream relationship between available stream power and width of alluvial valley floor. Lewin (1987) further suggests that channel planform change in the British uplands is dominated by lateral channel migration, progressively eroding valley fills with rates of change being largely determined by land-use change and other human influences.

2.5 The role of flood events in contemporary river channel change

2.5.1 Response of upland stream channels to floods

Downstream variations in channel plan and cross-sectional form of upland, gravel-bed streams in response to flood events depends on a range of interrelated factors at both the basin and local scale. At the basin scale, channel gradient (Milne, 1983a; Ferguson et al, 1996), upstream drainage area (Carling, 1986) and valley floor width (Anderson and Calver, 1977; Milne, 1983a) influence the role of flood events on river channel change. Sediment supply (Harvey, 1991) bank cohesivity (Milne, 1983a), the pool-riffle and meander-bend sequence (Milne, 1982b; Carling, 1991) and the presence of bars (Ferguson and Werritty, 1983) produce more localised variations in channel response.

This section identifies those factors recognised as influencing changes in the cross-section and planform of upland gravel-bed streams following high magnitude flow events. The methods and techniques used to establish the nature and extent of channel change in response to flood events are identified. Finally, downstream patterns of aggradation and degradation of the channel resulting from flood events and erosional and depositional features associated with the passage of a flood through upland gravel-bed streams are described.

Factors which influence the response of upland stream channels to floods

The sequence of flood events has been shown to be a critical factor governing the extent of channel change (Hitchcock, 1977; Werritty, 1982; Ferguson and Werritty, 1983; Carling, 1986; Harvey, 1986; 1991; Coxon et al, 1989). A high magnitude event can make the channel unstable which when followed by a subsequent lower flow causes a greater extent of channel change than would normally have been expected. Lane et al (1996) have suggested that inter-arrival periods between peaks in flood discharge are as significant as the peaks themselves.

Channel gradient varies downstream and this has consequences for both stream power and channel pattern (Ferguson, 1981). For example, Ferguson and Ashworth (1991) found that as channel gradient becomes more gentle, meander bends become increasingly stable. Milne (1983a) highlights the influence of low channel gradient on the adjustment of upland stream channels to floods, by arguing that high amplitude 'goose-neck' meander bends may form where a low gradient channel flows through fine-grained cohesive sediments. In upland gravel-bed streams local base-level control is identified as the cause of rapid downstream fining of bed material and a change in channel pattern from near-braided to sinuous (Ferguson and Ashworth, 1991; Ferguson et al, 1996). Milne (1979) argues that if a locally low gradient exists, this can lead to reduced stream competence and a weak relationship between the width and depth of the channel. Ferguson (1987) has suggested that when considering the role of channel slope on the development of channel pattern, in many cases, the modern channel slope is relict and may reflect past rather than present alluvial sedimentation. In other words, channel slope may not reflect adjustment of the channel, through aggradation or degradation, to the present sediment supply from upstream.

Anderson and Calver (1980) have suggested that if a large flood event is to be instrumental in causing major river channel change, sufficient upstream drainage area is needed in order to generate discharges. However, this may be questionable bearing in mind the example of the Noon Hill flash flood (Carling, 1986) where a very large discharge was generated from a small area in the Langdon Head basin (3.5 km^2).

Anderson and Calver (1977) suggested that where there is stream confinement resulting from a narrow valley floor, as in the case of Cannon Hill during the 1952 flood on Exmoor, flooding will result in cross-sectional channel changes, with channel adjustment occurring through bank undercutting and channel incision. It is suggested that in confined reaches, characteristic of small upland streams, hillslope stability controls the response of cross-sectional form of the channel to the passage of a flood. This is in contrast to flooding on the neighboring Hoaroak Water, Exmoor, where there was sufficient valley width to attenuate the flow and develop new channels. In this case, changes in channel planform resulted.

Changes in sediment supply to a stream may be the result of climatic changes, major floods or land-use change, e.g. mining activity and deforestation (Ferguson, 1987). If the change in sediment supply is transient, a sediment slug may move downstream as a diffuse wave, marked by 'sedimentation zones' (Church, 1983) which may temporarily alter the channel pattern from a meandering to a braided one. The location of abundant sediment sources during major flood events have been identified as streamside scars and bank erosion (Harvey, 1986; 1991), tributary channels (Carling, 1986) and landsliding (Coxon et al, 1989).

However, it is clear that the calibre and quantity of the sediment supply is very important if channel change is to result from flood events (Werritty, 1982; Carling, 1986; Harvey, 1986). An abundant supply of coarse sediment combined with the morphology of the existing channel and the presence of numerous inactive channels are identified as critical to channel and floodplain response to flood events within in the River Feshie in the Scottish Highlands (Werritty and Ferguson, 1980) and in Langden Brook (Hitchcock, 1977). However, Ferguson (1987) has suggested that, in general terms, the size of bed material supplied to the channel is more important in determining channel pattern changes than the amount of sediment supplied, through the influence of bed material size on stream roughness and stream competence.

Harvey (1991) identified the sudden increase in availability of coarse sediment, generated by a rare flood event in June, 1982, as a key factor in controlling the switching of upland channels from a meandering to a braided pattern. Figure 2.5 shows channel response and recovery from the 1982 flood. Large sub-rounded and angular boulders were fed into the stream system due to trenching in gullies and tributary channels. Sediment input from streamside scars and bank erosion caused more localised channel change. As debris cones blocked pre-flood channels, streams were diverted across the floodplain. Channel blockage by coarse boulder and cobble sediment 'plugs' was a direct cause of major channel avulsion on open bends. The geomorphic effects of the flood were to steepen the gradient and widen the channel which prompted the switch of channel pattern from meandering to braided.

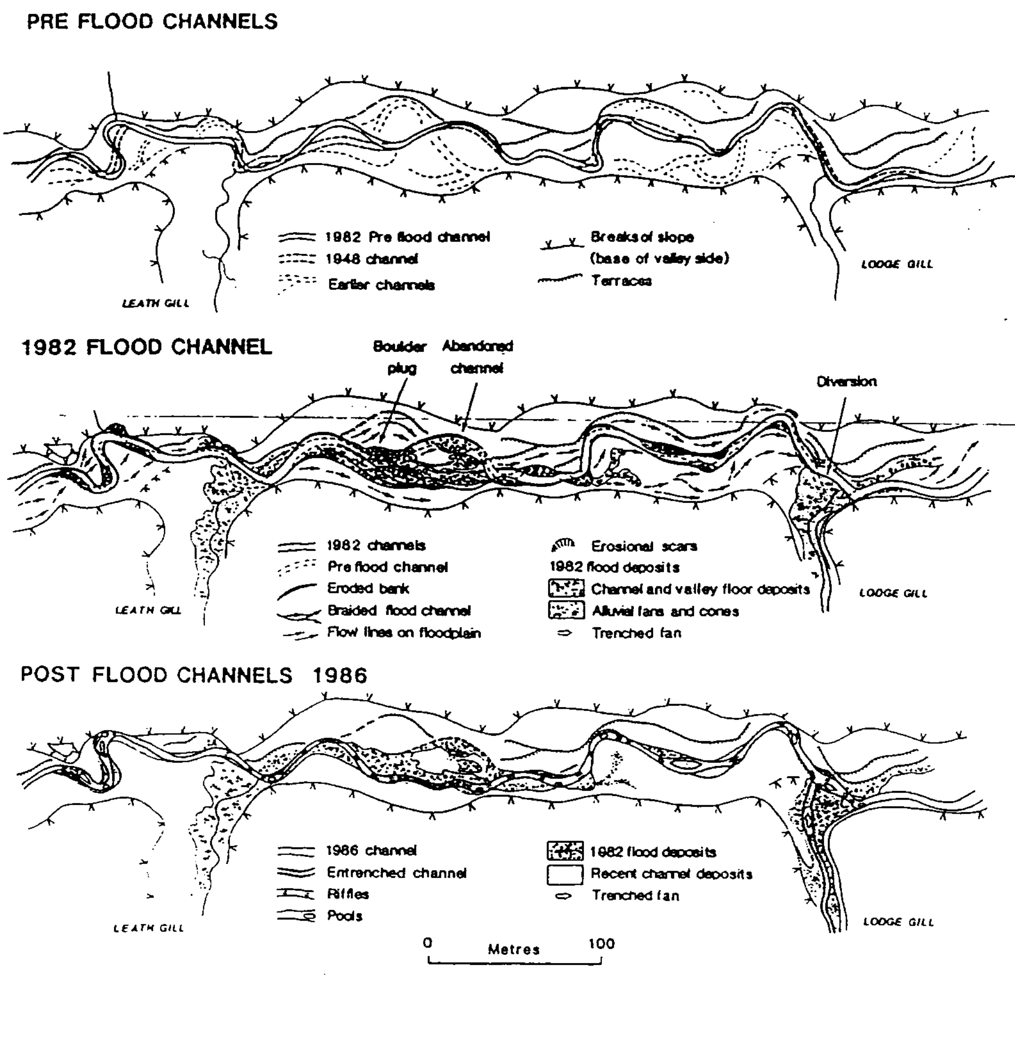


Figure 2.5 Channel response and recovery from the 1982 flood in the Howgill Fells. Example reach in Bowderdale (Harvey, 1991)

A similar shift in channel pattern is described for Langden Brook in the Forest of Bowland (Hitchcock, 1977). Hitchcock (1977), identified the abundant supply of coarse sediment derived from the reworking of channel bars and from incohesive channel banks as the key factor instigating river channel change, particularly through the development of a braided channel pattern. Milne (1983a) identifies sediment supply combined with the cohesiveness of the channel banks and pattern of valley-side confinement as determining the way in which the mean annual flood controls cross-sectional scale and channel planform within the uplands. Similarly, the critical role played by sediment supply in determining the geomorphological response of upland gravel-bed streams to high magnitude floods is demonstrated by Carling (1986). Carling (1986) suggests that during the Noon Hill flash flood, although entrenchment of bedrock occurred in the upper reaches of streams on the watershed between Weardale and Teesdale in the Northern Pennines, since the energy input was localised and abrupt, downstream inputs of large quantities of sediment reduced levels of erosion in the lower reaches.

Lane et al (1996) have suggested that the nature of channel change in a meltwater stream of a glacierised catchment is primarily controlled by variations in upstream sediment supply. Although a local relationship between increasing discharge and channel erosion was identified, this was in the absence of upstream sediment supply. Change in channel morphology was dominated by the relative timing of discharge and sediment supply waves from upstream and by the existing river channel morphology.

The composition of channel banks influences where bank erosion is most likely to occur during a flood and consequently conditions patterns of channel sedimentation (Milne, 1982b; Ferguson, 1981). Research by Hooke (1979, 1980), on lowland Devon rivers suggested that the sediment composition of banks was the dominant factor controlling the rate of bank erosion in response to floods. Banks with higher silt-clay composition were observed to be more resistant to erosion and experience less slumping. However, it was also suggested that bank erosion was controlled by a complex combination of factors including local slope and secondary circulations within the channel during high flow events and position on the meander bend.

Milne (1983a) using examples from eleven upland gravel-bed streams in Northern Britain and the Scottish uplands suggests that irregular cross-sectional geometries, characteristic of upland gravel-bed streams, result from localised changes in the cohesivity of bank sediments. Where cohesive bank material is located so that the current from upstream meets the cohesive sediment at a high angle, stronger secondary circulations are established leading to the development of very tight meander bends (Milne, 1983a). Because these bends have greater cross-sectional asymmetry, with excessive widths across well-developed point bars, secondary currents are reinforced, bend curvature increases and erosion is further concentrated at the bend apex during flood events. This process encourages the development of high amplitude, 'goose-neck' meander bends along channels with highly cohesive banks and particularly within channels where the gradient is relatively low.

Similarly, Milne (1983a) using the example of the upland gravel-bed Stanhope Burn, observed that in segments of channel with coarse grained non-cohesive banks the channel was virtually straight, wide and shallow. However, where the channel banks were cohesive, the channel course was largely confined, producing a highly sinuous, narrow and deep channel. Extensive pockets of coarse floodplain sediments have been observed to coincide with exceptionally long riffles in upland gravel-bed streams (Milne, 1982c). Where coarse material is supplied to the channel from cut banks, the base of the bank is protected from deep scouring during high flow events which decreases the width:depth ratio in these locations (Milne, 1982a). Non-cohesive banks combined with high shear stresses generated by major flood events has resulted in channel widening on the River Feshie in the Scottish Highlands at a rate of seven metres per year (Ferguson and Werritty, 1983).

However, upland stream channels flowing through alluvial deposits often have banks with a composite structure of cohesionless sand, gravel and cobbles, overlain by cohesive silt clay deposited by overbank flow. During high flow events, a higher rate of bank erosion occurs from the lower cohesionless bank than from the upper more cohesive bank. This process of bank undercutting which produce cantilevers of cohesive material which eventually become unstable and collapse into the channel has been described for the River Severn at Morfodian, Powys, Wales (Thorne and Tovey, 1981).

Bank vegetation can affect channel pattern changes by restricting lateral erosion (Ferguson, 1987). Where a dense layer of vegetation covers the floodplain, widespread erosion of the floodplain, resulting from overbank flow can be prevented (Anderson and Calver, 1977).

The presence of bars in upland gravel-bed streams have an important influence on channel morphology during floods. Bars are major stores of bedload and tend to encourage bank erosion of the opposite bank through flow diversion away from the bar surface. Observations of hydraulic conditions during floods on the River Feshie, in the Scottish uplands, have identified how the episodic elongation of diagonal bars by deposition can cause bank erosion of adjacent channels of up to 10 metres (Ferguson and Werritty, 1983). As the normal pattern of flow of convergence into and divergence out of riffles is reversed during major flood events, (Keller, 1971) flow diverges over bar tops and margins and reconverges into the deeper pools. The result is the downstream advance of diagonal bars and since the flow is divergent, the bar progrades. A consequence of this is downstream riffle migration and abandonment of former pool heads. This process can lead to a wave of downstream bank erosion as lateral aggradation of the bar face constricts the channel alongside which may erode. The process of diagonal bar progradation may result in a chute being incised through the middle of the bar or alternatively a chute may develop along the margin of the bar which is riffle free. In this way channel change occurs as the chute becomes more active than the previous riffle (Figure 2.6). Accretion of the distal margins of diagonal bars alters the flow pattern over the bar in future floods, altering the nature of bedload transport which controls the evolution of bars and the channel.

A similar process was observed by Lewin (1976) on the River Ystwyth in Wales, where following straightening of the channel, during subsequent flood events, bars became attached to alternate sides of the channel and erosion occurred on the opposite bank, re-establishing a meandering channel. Similarly, Laronne and Duncan (1992) describe the process of alternate bar formation during high stages of flow due to the advance of sheets or lobes on the gravel-bed Ashburton River, New Zealand.

Likewise, Coxon et al (1989) describe how the release of large particles into the stream through bank undercutting during a flash flood on the Yellow River led to the formation of a large mid-channel bar with an erosional chute to one side. Subsequently this led to further bank erosion on the opposite bank to the bar. Similarly, the formation of irregular point bars during a flood in the Plynlimon catchment in the upland of mid-Wales has been linked with bank undercutting on the opposite bank, resulting in increased channel sinuosity (Newson, 1980). However, Knighton (1998) suggests that in the long-term erosion of one bank is approximately compensated by deposition against the other.

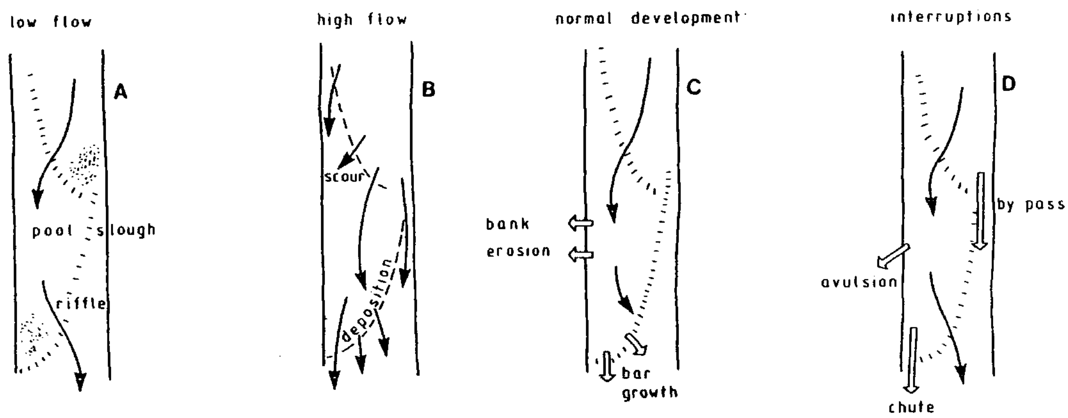


Figure 2.6 Schematic diagrams of flow over alternate diagonal bars and consequent morphologic development on the River Feshie, Cairngorms.
a) meandering thalweg at low flow with convergence over bar faces; b) divergence and deposition on bar face at high flow; c) normal pattern of episodic progradation in floods and consequent bank erosion; d) possible interruptions due to avulsion within channel or from it (Ferguson and Werritty, 1983).

On a smaller scale, deposition of groups of coarse particles either within the channel or as toe deposits at the base of the channel banks during a flood event can initiate a sequence of instability resulting in channel change. Within-channel coarse deposits can direct flow towards the channel banks causing bank erosion or scouring of the bed (Hitchcock, 1977).

River channel change in response to floods is largely governed by the detailed morphology of the existing channel and its floodplain, and the hydraulic processes associated with flood events (Werritty and Ferguson, 1980). The pool-riffle sequence, which is the characteristic reach-scale bedform of many gravel-bed channels of low to moderate slope, influences channel processes during flood events which has implications for the development of cross-sectional form and channel pattern. Although the pool, riffle, meander bend sequence can induce local variations in channel response during floods of high magnitude, equally the pool-riffle bed can be destroyed by large flood events leading to a change in channel pattern from meandering to braided (Harvey, 1986).

The link between bed topography and changes in cross-sectional morphology and bed material size during flood events are well documented (Milne, 1982b; Knighton, 1998; Carling, 1991; Clifford, 1993; Sear, 1996). Keller (1971) identified the 'velocity reversal hypothesis' whereby during high flow events velocity in a pool increases faster than that of a riffle, until at a relatively high discharge, the velocity of a pool exceeds that of a riffle. The net effect of this process is that during high flow events, pools are deeply scoured and coarse sediment is transported from one pool to another, with storage on the intervening riffle. It has been suggested that through a combination of high flow transport through pools and low flow storage on riffles, coarser material will be concentrated on riffles (Knighton, 1998). The 'velocity reversal hypothesis' offers a mechanism for a pattern of scour and deposition capable of producing areal sorting of stream bed material and of maintaining the pre-existing bed topography during high flow events. Riffles are the most stable bedform since their closely packed structure resists erosion during high flow events (Clifford, 1993; Sear, 1996). Although, mid-channel accumulations of coarse bed material or central bars at riffle sites deflect flow towards the banks leading to bank undercutting and channel widening, if flow is overbank during floods these channel deposits will be submerged thus inhibiting further bank erosion.

Although the velocity reversal hypothesis has been accepted as the cause of contrasting patterns of surface sediment size and sorting in pools and riffles (Keller, 1971, Andrews, 1979; Lisle, 1979), the existence of velocity reversal has been questioned by some authors (Carling, 1991; Clifford, 1993; Sear, 1996). For example, the maintenance of juxtaposed riffles and pools has been described in terms of sediment structure rather than sediment size (Clifford, 1993; Sear, 1996). Higher velocities in pools compared with riffles during high flow events (Keller, 1971) and loose bed structure in pools (Clifford, 1993) encourages deep scouring of the pool bed. It has been argued that, even without a velocity reversal, competence could still be higher at high flows in pools if their bed sediment has a more open structure (Clifford, 1993). Carling (1991), argues that, in the case of the River Severn, for a complete velocity reversal to occur, riffles need to be significantly wider than pools during high flows in order to accommodate lower velocities. Since bank strength on the River Severn prevents widening of riffles, compared with pools, there was no evidence for a velocity reversal at within-bank flows.

It is generally accepted that channel sinuosity is a major control over cross-sectional form (Milne, 1983a). The sinuosity of a meander bend influences resistance to flow during a flood event which may result in downstream variations in channel erosion. During high flow events, the majority of channel change occurs at meander bends. Milne (1982c) observed that along sinuous reaches of coarse bedload channels in upland Britain, the widest bedforms occurred in association with bank erosion at tightly curved pool sites around actively migrating bends. Pools associated with the bend apex help to focus erosion at the concave bank thus increasing the amplitude of the bend. Erosion of the concave bank of a meander bend is coupled with aggradation within the channel and on adjacent point bars as a result of transverse bed material transport and secondary circulations (Milne, 1982b).

2.5.2 Morphological changes in channel form associated with the occurrence of flood events in upland, gravel-bed streams

The downstream pattern of aggradation and degradation of the channel bed resulting from floods in upland gravel-bed streams has been identified (Werritty and Ferguson, 1980; Werritty, 1982; Ferguson and Werritty, 1983). Rates of scour and fill within the channel of the River Feshie are reported as being up to one metre in depth per year (Ferguson and Werritty, 1983). River channel changes are reported to result from the sudden switching of the rivers course due to the gradual aggradation of active channels combined with the degradation of nearby inactive channels during overbank flooding (Werritty and Ferguson, 1980; Laronne and Duncan, 1992). In the example of the River Nethy, a tributary of the River Spey in the Scottish Highlands, a sequence is described whereby aggradation leads to local ponding back of the flow which subsequently leads to overbank flow (Werritty, 1982). Similarly, Warburton (1992) describes how floods on the proglacial stream of Bas Glacier d'Arolla, Switzerland can lead to dramatic shifts in the channel from slightly aggrading to a state of considerable aggradation or degradation of the channel bed.

Carling (1986) describes a flash flood centered over Noon Hill in July 1983 which produced a very similar geomorphological response in Langdon Beck, Ireshope Burn and West Grain in the Northern Pennines as described for the Howgill Fells, Cumbria (Harvey, 1986;1991). In the upper reaches of these Pennine headwater streams (Carling, 1986) the shale bedrock and jointed limestone were eroded, the latter in the form of large boulders. In the lower reaches, aggradation occurred through the deposition of gravel up to one metre in depth. Carling (1986) describes how coarse sediment input from tributary channels formed boulder jams in the main channel with extensive channel scour occurring immediately downstream of the boulder jam. This process produced distinctive zones of bed erosion and deposition along the channel and is referred to by Carling (1986) as a 'stepped-depositional feature'. Similarly, a summer flash flood on the Yellow River, Country Antrim, caused aggradation in the lower reaches of the stream where the channel was wide and of low gradient, resulting in braiding of the channel in places (Coxon et al, 1989). In the upper reaches of the Yellow River, boulder jams, similar to those described by Carling (1986) for streams in the Northern Pennines produced steps in the stream profile.

A whole range of erosional and depositional features associated with the passage of flood events through upland stream channels are identified in recent literature. Channel widening as a result of erosion and bank undercutting (Anderson and Calver, 1977; Ferguson and Werritty, 1983; Acreman, 1983; Harvey, 1986; Coxon et al, 1989;) particularly within bedrock channels (resulting in block detachment), and streams with a layer of coarse glacial deposits on their beds (Newson, 1980), has been widely reported as resulting from high magnitude flood events. The excavation of completely new channels (Werritty 1982) and the reactivation of old channels through secondary anastomosis (Hitchcock, 1977) also occur during flood events. Carling (1987) demonstrates that stream response to overbank floods is largely by lateral migration induced by erosion of the gravel and silt banks, e.g. Eggleshope Beck, Teesdale, Northern Pennines. A possible link between localised bank undercutting during flood events and the formation of downstream braid bars which may develop into lateral bars has been identified for Langden Brook, Forest of Bowland (Hitchcock, 1977). In braided streams which have an abundant supply of coarse sediment, lateral bars may form downstream of localised bank undercutting if the upstream end of an adjacent channel becomes blocked by coarse sediment during a flood event (Figure 2.7).

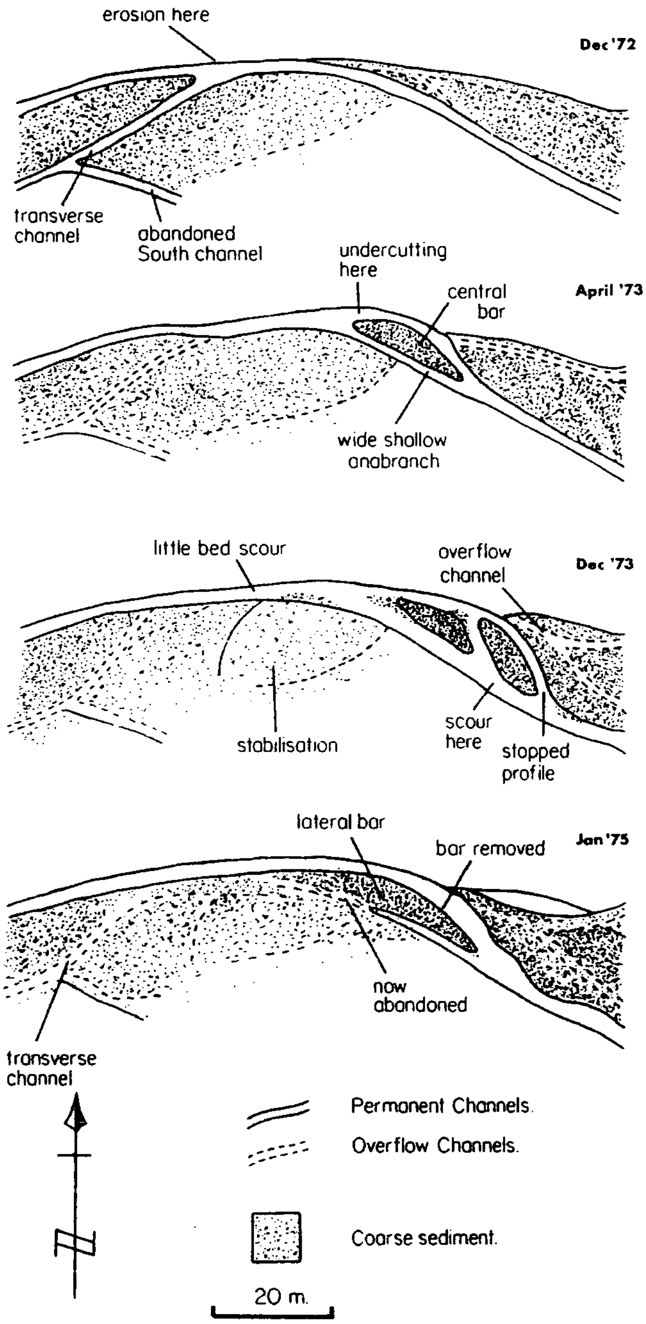


Figure 2.7 Sequence of channel change at Langden Brook (Hitchcock, 1977)

One of the major geomorphological effects resulting from the passage of high magnitude floods through upland streams is the straightening of the channel through initial channel blockage caused by the deposition of coarse material which can result in chute incision and bend cutoff (Harvey, 1986; Carling, 1986; Coxon et al, 1989). Channel change in streams in the Howgill Fells resulted directly from a combination of channel widening, straightening, bend cutoff and avulsion in response to a flash flood with a recurrence interval of 100 years. This altered many stream patterns from a meandering to a braided pattern (Harvey, 1986; 1991). Chute cutoffs associated with infilling of the channel with gravel similar to that described by Harvey (1986) in the Howgills is observed within the lower reaches of Langdon Beck, Ireshope Burn and West Grain in response to a flash flood in the Northern Pennines by Carling (1986). Laronne and Duncan (1992) suggest that moderate flow events can rework bars and bring about large changes such as cutoffs, avulsion of the main anabranch and channel switching.

The formation of mid-channel and diagonal bars are widely reported to result from flood events (Hitchcock, 1977; Werritty, 1982; Ferguson and Werritty, 1983). Progradation of alternate diagonal bars during flood events is shown to instigate avulsion both from and within the channel and high rates of downstream bank erosion (Werritty and Ferguson, 1983). Werritty (1982) describes the development of mid-channel bars in response to flood events for the River Nethy in the Scottish uplands. Hitchcock (1977) suggests a link between localised bank undercutting and the formation of downstream braid bars which subsequently develop into lateral bars. Also, the formation of boulder bars and boulders jams during high magnitude events are well-documented (Anderson and Calver, 1980; Acreman, 1983; Harvey, 1986; Coxon et al, 1989).

Other depositional features associated with major flood events have been identified as chute bars, large point bars and gravel splays on the inside of channel bends (Carling, 1986), the redistribution of gravel and cobble shoals (Newson, 1980) and the deposition of coarse debris over entire valley floors (Anderson and Calver, 1980).

Large quantities of bedload can be generated, particularly where tributary channels join the main stream often forming boulder jams and boulder bars (Carling 1986).

2.5.3 Sediment dynamics in upland gravel-bed streams during flood events

Gravel-bed rivers in the British uplands have long been associated with very high rates of erosion and deposition and the entrainment and transport of coarse bedload. Bedload transport has a tendency to occur episodically during major flood events (Klingeman and Emmett, 1982; Carling and Hurley, 1989). However, the study of bedload transport during flood events is problematic owing to the wide range of particle sizes found in natural gravel-bed rivers and the supply of material from outside the channel (Bathurst, 1987). Bed material heterogeneity leads to many problems in predicting bedload transport in a river (Klingeman and Emmett, 1982). The initiation of bed material motion may not occur simultaneously for all the available grain-sizes (Laronne and Carson, 1976). Likewise, sediment supply to the channel is highly episodic encouraging unsteady bedload transport rates during flood events.

This section identifies those factors influencing the rate and amount of bedload transport during flood events in upland gravel-bed streams. Secondly, patterns of downstream fining, a characteristic common to many upland gravel-bed streams, and the associated passage of sediment waves or 'pulses' along the channel during high flow events are discussed using examples from the British uplands.

Factors which influence bedload transport in upland gravel-bed streams

The incidence and nature of bedload transport in coarse alluvial channels during floods is not solely influenced by the magnitude of the flow event. Laronne and Carson (1976) argue that the transporting process is dependent on a combination of prevailing hydraulic conditions, bed topography and antecedent hydraulic conditions. Bed topography influences both the type of material being transported and bed roughness which is responsible for most of the stream energy to be dissipated through friction. Hassan and Church (1991) additionally identify the shape, roundness and size of particles as exerting a major influence over the movement of bedload in gravel-bed streams. Gomez (1983) questioned the existence of a single continuous relationship between the bedload transport rate and stream power during flood events, as a result of the process of progressive bed armouring. Reid et al (1985) also found a very poor relationship between bedload transport and water stage in Turkey Brook due to changes in the availability of sediment. Warburton (1992) found that variations in bedload transport over a sampling period of several hours in a proglacial mountain stream showed a poor relationship between discharge and bedload transport. Conversely, over a sampling period of 67 days, long-profile and cross-sectional surveys suggested that, apart from when the stream was in flood, steady-state transport is maintained.

The rate of bedload transport in gravel-bed streams is largely determined by the sequence of flood events (Gomez, 1983; Reid et al, 1985; Carling and Hurley, 1987). The channel bed becomes compacted during periods of low flow as the coarse framework gravels become infilled with a matrix of smaller particles. If the flood is the first of the season, the bed may have become stabilised and gradual loosening of the bed material only occurs as a result of a succession of floods which makes more material available for transport. Reid et al (1985) identified that after a considerable period dominated by low flows, the stream power at the threshold of initial motion is five times that of the overall mean. However, if there has been a sequence of relatively high magnitude flood events, the sediment supply, which is predominantly from within-channel sources in upland gravel-bed streams, may become exhausted causing a reduction in the rate of bedload transport (Carling and Hurley, 1987).

Church and Hassan (1992) and Hassan et al (1992), using evidence from a wide range of flow regimes including small, upland, gravel-bed rivers, suggest that the entrainment and distance of travel of particles is influenced by the magnitude of the preceding event. Martin and Church (1995) investigating the Vedder River, Vancouver, argue that summer and winter floods deliver different volumes of sediment for a given peak flow. In winter, snowmelt floods are likely to carry less sediment than a rainfall flood of the same magnitude since the latter is more heavily supplied by erosion caused by overland flow, influencing the bedload transport rate (Bathurst, 1987). However, Hoey (1992) argues that the order, inter-arrival time and absolute magnitude of events are all important for determining the impacts of increased water discharge and sediment supply during flood events.

The structure of upland stream gravel beds and in particular the interlocking and clustering of particles is found to inhibit their initial motion and hence reduce bedload transport rates (Brayshaw et al, 1983; Reid et al, 1985; Carling and Hurley, 1987; Carling, 1989; Richards and Clifford, 1991; Martin and Church, 1995). Reid et al, (1985) found that, when compared with flume experiments, the bedload movement in Turkey Brook commenced later as a result of a consolidated bed and continued for a longer period of time than had been previously predicted. Reid et al (1985) identified that the sporadic break-up of clusters of sediment particles results in fluctuations in bedload transport during flood events, often over the timescale of a few minutes. Hoey (1992) suggests that this particle cluster breakup mechanism is responsible for producing pulses of bedload movement along a channel. Brayshaw et al (1983) suggest that fewer clustered particles are entrained by flood flows than those on a plane bed as their incipient motion is delayed, which influences the incidence and timing of bedload transport. Clusters were found to influence both bedload transport and downstream fining through their addition of 'roughness elements'.

The attitude of an exposed particle in relation to surrounding embedded particles was found to be important when considering the entrainment of bed material in gravel-bed rivers (Brayshaw et al, 1983). Similarly, the uneven protrusion of particles on poorly-sorted, compacted channel beds of streams in the Pennine uplands had been found to affect bed roughness and particle entrainment, influencing bedload transport rates (Carling, 1983). Laronne and Carson (1976) suggest that the presence of particles protruding from the bed surface is strongly dependent upon modes of transportation and deposition. However, Richards and Clifford (1991) argue that an imbricated bed, in which there are relatively few exposed grains, bed material is stabilised against disturbance and transport.

Carling (1989) associated low bedload transport rates in the headwater tributaries of the River Tees to the packing characteristics of the bed material in which the undisturbed coarse framework elements of the bed become armoured. Carling (1988) suggests that a compacted bed combined with cohesive banks can both inhibit bedload transport and hence river channel change. Conversely, unstable banks, through bank collapse can increase the bedload transport rate (Carling and Hurley, 1987). Similarly, channel margins are believed to have a dominant role in acting as sources and sinks for small-sized bed material when the armour layer is stable at low flows (Klingeman and Emmett, 1982).

Variations in the rate of bedload transport due to progressive bed armouring has also been identified for gravel-bed streams in proglacial mountainous streams (Gomez, 1983; Warburton, 1992) and the mountainous Squaw Creek, Montana (Andrews and Erman, 1986; Carling 1998). Klingeman and Emmett (1982) suggest that an armour layer inhibits the incipient motion of bed grains and that the degree of inhibition varies with grain-size. Gomez (1983) argues that a coarse 'armour' layer forms when finer particles are winnowed from the surface layer during flood events of low or intermediate magnitude leaving behind the coarser particles which can only be entrained by a flood event. This eventually stabilises the surficial bed material, effectively preventing any further removal of underlying material. The net result is a coarsening of channel bed sediments (Carling, 1998). Conversely, Laronne and Duncan (1992) attribute large channel changes during moderate flow events to the lack of a coarse armour in the bed and banks of reaches along the gravel-bed Ashburton River, New Zealand.

Carling and Reader (1982) distinguish between the formation of an 'armour' layer as described by Gomez (1983) and a 'censored' layer observed in Pennine gravel-bed streams. Carling and Reader (1982) argue that during intermediate flows fine materials are winnowed from the interstices of the coarse surface framework, but that during a flood event initially material stored as Ostler lenses are mobilised, followed by matrix material winnowed from the framework gravel. Once the surface layer has been depleted of fines only coarse gravels remain and this is termed a 'censored layer' which protects the underlying finer material until a flood event of sufficient magnitude occurs and is capable of removing framework gravels. The resulting 'open-work' gravel at the surface has a similar size distribution to an 'armour' layer. However whereas an 'armour' layer is a coarse lag layer formed by deposition at the surface of gravel-beds on waning flows which prevents similar or finer material from entrainment, a 'censored' layer does not require any mobilisation of the coarse fabric. Church et al (1987) illustrate the differences between a 'framework gravel' and a 'censored gravel' (Figure 2.8).

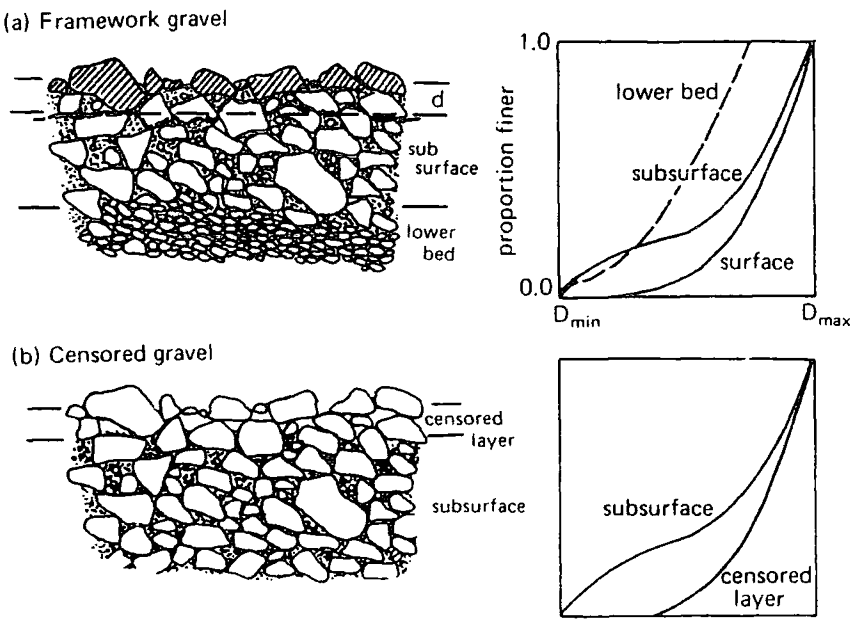


Figure 2.8 Typical structures of framework and censored gravels with grain-size distributions (Church et al, 1987)

Bed structure of upland gravel-bed streams can influence size-selective transport of bedload. For example, Ashworth and Ferguson (1989), concluded from tracer pebble experiments that entrainment is size-selective in upland gravel-bed streams. Klingeman and Emmett (1982) argue that if the bed material is bimodal in its grain size distribution, abrupt changes can be expected in particle availability for transport, the particle transport rate and in the size composition of bedload at times of armour breakup.

Particle mobility is greatest for material in open bed structures (Naden and Brayshaw, 1987) and smallest for sediment in tight structural arrangements, for example, in the gravel-bed channel of Seale's Brook, Quebec (Laronne and Carson, 1976). Carling (1987) argues that small particles become entrapped in the interstices of an immobile surface layer of cobble-bed material, whereas particles larger than the average grain-size of the bed material can easily roll over the underlying bed material, although they are not likely to travel very far owing to their mass. Similarly, Laronne and Carson (1976) suggest that as particle size increases, particle mobility and distance of transport will increase. Laronne and Carson (1976) argue that continuous rolling of larger particles is favoured due to a decrease in the probability of becoming trapped by a small unevenness of the bed surface. However using the example of Seale's Brook, Quebec, Laronne and Carson (1976) suggest that this trend would only continue up until the increased size of a particle encourages it to become embedded within closed structures of the channel bed.

Variations in bed topography also influence the amount and rate of bedload transport in gravel bed-streams (Laronne and Carson, 1976). Since some elements of the bed are stable and others are unstable, sediment transport during floods is discontinuous and channel storage is spatially variable (Brayshaw et al, 1983). Martin and Church (1995) argue that spatial variability in transport rates demonstrate that if measurements of bedload transport are taken at a particular site the results will not necessarily represent the reach length pattern of sediment transport.

However, the sequence of pools and riffles, the characteristic bedform of upland gravel-bed streams, exerts a major influence over bed structure which in turn influences bedload transport rates. The pool-riffle sequence is associated with spatial patterns in bed material size with riffle sediments being coarser and better sorted than adjacent pool sediments (Carling, 1991). Contrasting patterns in sediment size and bed structure of pools and riffles has been explained in terms of the velocity reversal hypothesis (Keller, 1971). Whereas Keller (1971) proposed that coarse sediment transport occurred from one riffle to another during high flow events, Clifford (1993) has suggested that scoured sediment is routed from one pool to another over the intervening riffles, explaining the filling of pools with coarse sediment during high flow events with scouring in pools occurring during lesser flows. However, there is continuing controversy as to whether infilling of pools occurs during high flow (Lisle, 1979; Carling, 1991; Clifford 1993; Sear, 1996) or during the intervening low flows (Keller, 1971; Milne, 1982b).

Clifford (1993) and Sear (1996) argue that turbulent flow over riffle surfaces during low flows, causing particle vibration and sporadic particle motion results in their bed structure being more compacted when compared with the more open structure of bed sediments in pools. Since the bed structure of pools is very loose, higher velocities in pools during high flow events quickly mobilises most of the sediment at the base of the pool encouraging transport of the largest particles through the pool rather than their deposition within it. Conversely, the more compact structure of riffle surfaces tends to trap the coarser fraction of the bedload and reduces erosion during high flow events. The net effect on sediment dynamics during flood events is that particles in pools travel further than particles in riffles with sediment transport being from pool to pool with storage on riffle surfaces. The occurrence of fine material drapes over scoured surfaces in pools provides evidence of the role of the velocity reversal hypothesis in sediment transport processes through the pool riffle sequence (Clifford, 1993).

The amount and calibre of sediment supplied to a stream from the banks and adjacent valley-side slopes influences bedload transport rates in upland gravel-bed streams and largely depends upon timing, order and magnitude of flood events. An inconsistent relation can be expected between bedload transport and water discharge if there is much spatial variability in the availability of transportable bed material along the length of a stream (Klingeman and Emmett, 1982). High magnitude discharge events and sediment supply events are not necessarily coincident in space or time (Hoey, 1992).

Carling and Hurley (1987) highlight the importance of within-channel sediment sources during flood events for streams in Teesdale, Northern Pennines, since the slopes adjacent to the channel are largely turf-covered. Bank collapse on the waning limb of the flood hydrograph and gradual depletion of fine matrix sediments from the channel bed provide the main sources of sediment. Carling (1998) identifies channel bars as the principal source of sediment supplied to the channel during moderate bedload transport events in the mountainous Squaw Creek, Montana. Richards and Clifford (1991) distinguish between relative and absolute sediment supply limitation, where the former is reliant upon supply restriction as a result of bed structuring and the latter is dependent upon the external supply for bank erosion or upstream influences.

In-channel storage of sediment can delay and attenuate the movement of sediment waves resulting from external supply events (Bathurst, 1987). Reach-scale sediment storage in the Allt Dubhaig in the Scottish Highlands has been demonstrated to influence bedload transport rates during floods (Wathen et al, 1997). Bed morphology, and in particular the presence of channel bars, influences within-channel sediment dynamics. Sediment stored in bars is less active than sediment in deep channel areas (Laronne and Duncan, 1992). For example, Warburton (1992) found that, for a proglacial mountain stream, the majority of bedload is transported in narrow bands corresponding with the maximum thread of flow. Also, Butler (1977) regarded the position of the particle in the channel cross-section as important in bedload movement during flood events. Bars which form in response to channel curvature and associated hydraulics become fixed within the channel and the sediment remains in storage for considerable periods of time.

Patterns of downstream fining and sediment waves

Although there is disagreement over the factors which influence the rate of bedload transport in upland gravel-bed streams the theory that bedload is transported through the channel as a 'pulse' or 'wave' of particles is quite widely accepted (Hoey, 1992). Bedload pulses can be produced by a range of processes in any given environment. Hoey (1992) has suggested that bedload pulses can be produced by both lateral and vertical grain sorting, either during the development of a degradational armour (Gomez, 1983) or by the episodic motion of large particles. Scour and fill sequences during flood events have also been associated with bedload pulses (Jackson and Beschta, 1982). However, many studies of external sediment input to the channel in the form of a sediment wave, or pulse, relate to the input of coarse mining waste in upland catchments (Macklin and Lewin, 1989).

Reid et al (1985) describe bedload as passing as kinematic waves in a 'slow moving traction carpet'. This is similar to Gomez (1983) who describes bedload as covering the entire active bed like a 'mobile patchy carpet'. Reid et al (1985), using a continuous bedload monitoring system on Turkey Brook, observed that particles move as groups identifiable as a slow moving layer one or two grains thick. Ashworth and Ferguson (1989) have argued that in upland environments, a wave of bedload finer than the existing bed material migrates along the channel and this initially promotes bed material fining with rapid aggradation of the bed. This theory of the existence of a sediment wave which progressively passes through the channel is supported by Lane et al (1996) in glacierised catchments and Ferguson et al (1996) on the Allt Dubhaig in the Scottish Highlands. Ferguson et al (1996) explain the process of downstream fining of bed material, a characteristic common to many gravel-bed streams in the British uplands (Ashworth and Ferguson, 1989) in response to a decline in slope imposed by a local base level which occurred in the form of a wave of fine bedload migrating along the channel. This produced a shift in channel pattern from near-braided to sinuous. Ashworth and Ferguson (1989) attributed pronounced downstream fining on the River Feshie, Scottish uplands, to a decrease in the percentage movement of and mean distance moved by progressively coarser classes of pebbles.

A decline in slope has long been associated with downstream fining of bed material in upland gravel-bed streams (Milne, 1979; Knighton, 1980). Laronne and Carson (1976) argue that local bed slope is a significant factor influencing the probability of particle entrainment and particle mobility. Knighton (1975) found that in the upper reaches of the River Bollin in the Pennine uplands, particle size changed systematically downstream with channel gradient. This pattern of downstream fining was particularly clear in headwater zones since the input of coarse sediment from tributaries was relatively small. The relationship between slope and mean grain-size distribution is highly dependent upon the size of material initially supplied to the channel and the distance over which the stream can effectively operate on the material (Knighton, 1975).

Summary

Although it is clear that various factors are significant in determining the rates of bedload transport, stream power appears to be a critical factor in achieving high rates of bedload transport in gravel-bed rivers. It is only during high flow events that the coarser framework gravels are disrupted, and near capacity sediment transport results (Carling and Reader, 1982; Carling and Hurley, 1987). When stream power is high and when almost all bed material is moving, the influence of bed structure on bedload transport rates will be minimal (Klingeman and Emmett, 1982). However, although the majority of particle sizes present in the stream bed surface are transported during sustained large discharges (Andrews and Erman, 1986) it has been argued that, in streams with shear stress only slightly above the threshold of particle movement, most particles move a relatively small number of steps, probably one or twice per flood event (Hassan and Church, 1991).

Carling (1987; 1988) stresses the importance of overbank flows in the entrainment of the whole range of bed material and the instigation of river channel change through positive feedback mechanisms in the case of headwater tributaries of the River Tees in the Northern Pennines. Laronne and Carson (1977) argue that larger floods associated with larger discharges activate the bed to a greater depth, resulting in higher transport rates. In contrast Reid et al (1985) suggest that during flows of higher than 0.9 metres turbulent eddies become remote from the bed failing to disturb underlying particles leading to a reduction in the bedload transport rate.

2.6 The role of climate and land-use change in determining flood regime and river channel change

The interaction of large floods, changes in land-use and climatic change and their influence on water and sediment yields must be assessed in order to explain river channel changes over the past 200 years. Subtle changes in climate and land-use can influence the hydrological regime within a catchment and significantly alter the magnitude and frequency of floods (Macklin, 1986, McEwen, 1989). This in turn may lead to river channel changes, often in the form of erosion or aggradation of the channel bed (Macklin and Lewin, 1986; 1989) and in some cases channel switching from a single meandering stream to a multi-channel braided system may occur (Harvey, 1986). However, until recently, little attention was paid to changes in land-use over the historic timescale, despite the escalating human impact on fluvial systems.

Since periods of land-use change within a catchment often coincide with periods characterised by climatic fluctuations, the influence of each on flood regime and historic river channel change is often difficult to identify (Hooke, 1977; McEwen, 1989). Therefore it is not surprising that research has tended to collectively examine the interaction between land-use change, climatic conditions and large floods and their effects on river channels and floodplains (McEwen, 1989a; Macklin et al, 1991).

The timescales over which changes in land-use and climatic fluctuations alter flood regime is discussed with particular reference to the lag times between land-use and climatic change and channel planform change. The influence of changes in both land-use and climatic fluctuations, and a combination of both factors, on historic channel planform change is then explained, using examples from the British uplands.

The timescales over which changes in land-use and climate affect flood regime are variable. Direct linkages between a change in the discharge and/or sediment supply rarely produces an immediate response. There is usually a lag time between a change in land-use within a catchment and change in water and sediment yields affecting the channel (Gregory and Madew, 1982). For example, an increase in braiding on the River Dee, Aberdeenshire, pre-1900 was attributed to increased inputs of sediment caused by deforestation in the 18th century (McEwen, 1989a). Similarly, a change in land-use resulting from the onset and cessation of mining activity in a catchment may only temporarily alter the input of sediment to the channel (Aspinall et al, 1986; Macklin, 1986). Longer term fluctuations in climate, such as the onset of the Little Ice Age, led to a cooler, wetter climate and increased flood frequency in Northern Britain over a period of centuries (Macklin et al, 1992c). The occurrence of the 'Little Ice Age' is particularly relevant in establishing a link between the late historic climatic change and river regimes (Tipping, 1994). McEwen (1989b) found that an increase in snowmelt associated with the cessation of the Little Ice Age in mid 1800's may explain the increase in moderate to extreme flooding in the Whiteadder catchment, a tributary of the River Tweed, Scottish Uplands. However, the onset of the Little Ice Age is viewed by many as producing a highly spatial response.

2.6.1 Land-use change

Changes in land-use immediately adjacent to the river channel and over large areas of the catchment can induce changes in water and sediment discharge which may initiate a wide range of channel adjustments. The most extensive changes affecting streams in upland catchments are land-use changes attributable to agriculture, forestry, mining and grazing. Hooke and Redmond (1989) report that almost 35% of upland channels had shown some change of planform between 1870 and 1950 and that at least some of these changes were due to either channelisation, deforestation or urbanisation.

The considerable importance of deforestation within a catchment, with associated increases in sediment input to the fluvial system, is identified in many upland gravel-bed streams (McEwen, 1989a; Macklin and Lewin, 1989). Deforestation at a catchment scale accelerates hillslope and gully erosion which increases sediment inputs to rivers, sometimes causing a switch from a meandering to a braided channel pattern. Equally, deforestation of the floodplain greatly increases its erodibility during floods (McEwen, 1989a).

Conversely, agricultural improvement associated with land drainage is linked to an increase in flood peaks and channel incision (Macklin and Lewin, 1989; Macklin et al, 1992a; Rumsby and Macklin, 1994; Macklin et al, 1994). However, direct modification of channels in the form of flow diversions and localised embankments (McEwen, 1989a) and reservoir construction (Rumsby and Macklin, 1994) can reduce the severity of flood flows and floodplain inundation, but can also increase in-channel flows and promote channel incision. For example, the development of Kielder Reservoir reduced the magnitude of flood peaks and the severity of recent floods in the Lower Tyne (Rumsby and Macklin, 1994). However, because reservoirs trap up to 95% of bedload and suspended sediment this encourages scour immediately below the dam (Brookes, 1994). Channelisation can lead to a decrease in flow resistance and changes in the sediment availability and sediment dynamics which generally increases peak flows and reduces sediment transport (Gregory and Madew, 1982). Lewin (1987) describes a reach of the River Trannon, a tributary of the River Severn, as changing from a stable, low gradient, meandering channel to one in which straight and sinuous segments alternate and where channel erosion is prominent due to a channelisation scheme.

Many streams in the British uplands and particularly the Northern Pennines have been heavily polluted by metal mining during the 18th and 19th centuries. The consequences of heavy metal extraction on downstream channel and floodplain development in Pennine streams is well-documented, for example, the River South Tyne at Garrigill (Aspinall et al, 1986) and the River Nent, near Alston, Cumbria (Macklin, 1986). Macklin and Lewin (1986) identify the role of historic metal mining on channel development in the Rheidol Valley, Wales. It has been clearly demonstrated that historic metal mining increases sediment load and the input of fine- grained metaliferous waste impairs the growth of riparian vegetation (Aspinall et al, 1986;

Rumsby and Macklin, 1994). This leads to aggradation in the channel which is usually followed by a period of severe channel incision with the cessation of mining activity (Macklin and Lewin, 1986, Macklin et al, 1991). Terraced fills of Holocene age in the Rheidol valley, Wales, showed evidence of man-induced accelerated deposition of fine-grained alluvium with subsequent incision associated with historic metal mining in the late 19th century and which is still continuing (Macklin and Lewin, 1986).

When periods of mining activity coincide with the occurrence of large floods in small, upland, gravel-bed streams, dramatic changes in channel planform can result. Macklin (1986) observed that a coincidence of mining activity on the River Nent, Northern Pennines, with several large historic floods transformed the channel from a single to a multi-thread channel. The initial transformation was caused by either a single major flood or a series of floods which introduced large quantities of coarse sediment to the system. Once the initial flood-related transformation had begun coarse sediment provided by mining waste effectively perpetuated the process.

In cases where changing use of agricultural land, inputs of mining waste and extreme flood events all interact over the same timescales, high rates of channel change result (Macklin and Lewin, 1989). Macklin and Lewin (1989) identified the existence of five 'sedimentation zones' along a stretch of the River South Tyne, Northern Pennines, which are thought to have resulted from inputs of coarse mining waste, episodic erosion of Quaternary valley fills and from coarse sediment inputs from tributary streams. However, whereas the mobilisation of these sediments by major flood events in the nineteenth century led to substantial channel erosion and aggradation, in recent years, flood events without the influence of mining activity produce only localised channel change.

However, the role of land-use change on historical changes in channel planform must not be overestimated. McEwen (1989a), identifies a threshold in rainfall magnitude over which the effects of deforestation on catchment sediment yields are negligible. Macklin and Lewin (1986) demonstrated that the river channel in the Rheidol valley in Wales, which had undergone a period of aggradation followed by incision associated with historic metal mining, had nevertheless remained stable over a period of 150 years. Rumsby and Macklin (1994) having examined the interaction between climatic fluctuations, a number of changes in land-use and flood regime and their effect on channel changes in the River Tyne basin over the past 250 years, suggest that land-use has a limited effect on flood frequency. Although mining waste increased downstream supplies of suspended sediment, maximum rates of deposition occurred during a period of reduced mining activity. Rumsby and Macklin (1994) conclude that the main effect of land-use change is to increase the sensitivity of the Tyne to climatically induced changes in flood magnitude and frequency.

2.6.2 Climatic fluctuations

If channel change occurs in response to climatic fluctuation, it might be expected that whole regions would be affected. However, the effects of climatic change depends on runoff response, which will in turn vary with lithology, soil, and land-use characteristics (Hooke and Redmond, 1992). Climatic fluctuations can influence both the speed of storm runoff and periodicity of sediment mobility in river channels. McEwen (1989a) found that in the case of the River Dee, Aberdeenshire, periods in which there was an increase in moderate to extreme rainfall peaks and snowmelt associated with severe winters coincided with periods which had an increased frequency of moderate to extreme flood events. In the 1870's an increase in the number of moderate to extreme flood events coincided with an increase in braiding along the River Dee between 1869 and 1902. Likewise, periods of increased precipitation have been linked with increased flooding on Bowmont Water, Scottish Borders (Tipping, 1994) which led to accelerated bank erosion and transport of gravels, sufficient to remove evidence of pre-existing valley floor deposits. A phase of deep channel incision on Bowmont Water, commencing in the late 18th century is comparable with observed channel incision on Thinhope Burn, upper Tyne valley, Nothumberland (Macklin et al, 1994), but is not comparable with channel incision observed in many upland valleys in the Northern Pennines since many were affected by historic mining activity (Macklin, 1986). Limited channel incision post 1856 on Bowmont Water corresponds with channel incision in the Howgill Fells and Harthope Burn, in the Cheviots (Milne, 1982a). Hooke and Redmond (1992) suggest that increasing evidence of similar sequences of channel planform change in several valleys and the increasing number of dates showing a high degree of coincidence suggests a greater influence of climatic change in the British uplands than had hitherto been estimated.

The importance of climatic fluctuations on flood regime and river channel change have been demonstrated for the River Tyne basin (Rumsby and Macklin, 1994) and for headwater streams in the Northern Pennines (Macklin et al, 1992a; Macklin et al, 1994). Rumsby and Macklin (1994) discovered that modest variations in climate, and in particular average annual rainfall, over the past 300 years, had resulted firstly in phases characterised by a higher frequency of large floods, with a return interval of more than 20 years and secondly in phases of more moderate floods with a return interval of five to twenty years. Consequently, during periods characterised by large floods, there was widespread channel bed incision on the Tyne, in comparison with periods of more moderate floods which resulted in lateral re-working and sediment transfer in the upper reaches and channel narrowing and infilling in the lower reaches. Macklin et al (1994) argue that in some upland basins, such as Thinhope Burn, Northumberland, and in the Howgill Fells (Harvey, 1991) several metres of channel 'fill and cut' could be accomplished during one large flood. Periods of increased rainfall, and clustered periods of flooding in the upper South Tyne (Macklin et al, 1992a) and throughout the Tyne basin (Rumsby and Macklin, 1994) were found to coincide with major hydroclimatic fluctuations over the same period in Western Europe.

Therefore, although the influence of changing land-use and climatic fluctuations can be separately identified as the main cause of an alteration in flood regime and river channel metamorphosis, in many cases the two combine and the effects of one are indistinguishable from the other (McEwen, 1989a; Macklin et al, 1992; Rumsby and Macklin, 1994). For example, Macklin et al (1991) investigating the influence of historic land-use change on Coe Burn, Callaly Moor in Northumberland argue that episodes of valley floor alluviation were linked to episodic coal mining activity. However, subsequent stream incision through the valley fill to a depth of 1.7m may have resulted from either nineteenth century abandonment of farmsteads and field systems, causing a reduction in sediment supply, or a climatically-related increase in flood frequency and magnitude or both.

However, Werritty and Ferguson (1980), concluded that major river channel changes on the braided River Feshie in Scotland do not imply changes in climate or other long-term controls and they stress the need to appreciate significant short-term and small-scale variabilities in channel planform change. Although Hooke and Redmond (1992) argue that both human activities and climatic fluctuations have both had an impact on river channel change over the historical period they suggest that increasing rates of erosion and channel instability observed on Devon rivers is related to autogenesis in meander development in which increasing meander sinuosity leads to increasing channel instability. It is suggested that although changes in land-use and climate provide the impetus for channel change, physical conditions limit planform development.

CHAPTER 3

STUDY SITE AND FIELD METHODS AND TECHNIQUES

This chapter describes the characteristics of the Swinhope Burn study site and the main techniques used to collect the field data.

3.1 STUDY SITE

3.1.1 Location

The study site, at an altitude of 400 m O.D, comprises a 1.4 km long reach of Swinhope Burn, a south bank tributary of the upper River Wear in Weardale, Northern Pennines, County Durham (Figure 3.1). Swinhope Burn joins the main valley 0.5 km downstream of Westgate (NY 912 381) (Figure 3.2). The study site lies 2.5 km south-east of Daddry Shield, the stream flows north to north-east, between a field boundary downstream of Swinhope Bridge (NY 897 348) on the Westgate to Newbiggin road to a point adjacent to Swinside House (NY 900 355).

Swinhope Burn is a small, upland, gravel-bed stream flowing in a predominantly shallow channel along an irregular meandering course. The channel bed material covers a wide range from fine gravel to small boulders. The stream lies in a small, partly enclosed elongate basin (900 by 150 m), with high gradient slopes in the upper reaches and a steep morainic ridge, the Greenly Hills moraine (NY 912 359, 430m) 0.75 km downvalley of the study site (Figures 3.3 and 3.4).

3.1.2 Geology

The Swinhope Burn basin is underlain by Carboniferous Millstone Grit and Limestone with shales, sandstones and clays. This is overlain by glacial drift deposits and a thick layer of alluvium in the valley bottom. The outcrops of Great Limestone are marked by lines of sink holes. The Westernhope Old Vein, mainly composed of coarse fluorite crosses Swinhope Burn from Black Hill (NY 908 353) to the spur of High Pike (NY 895 360), but the fault is concealed by extensive drift deposits.

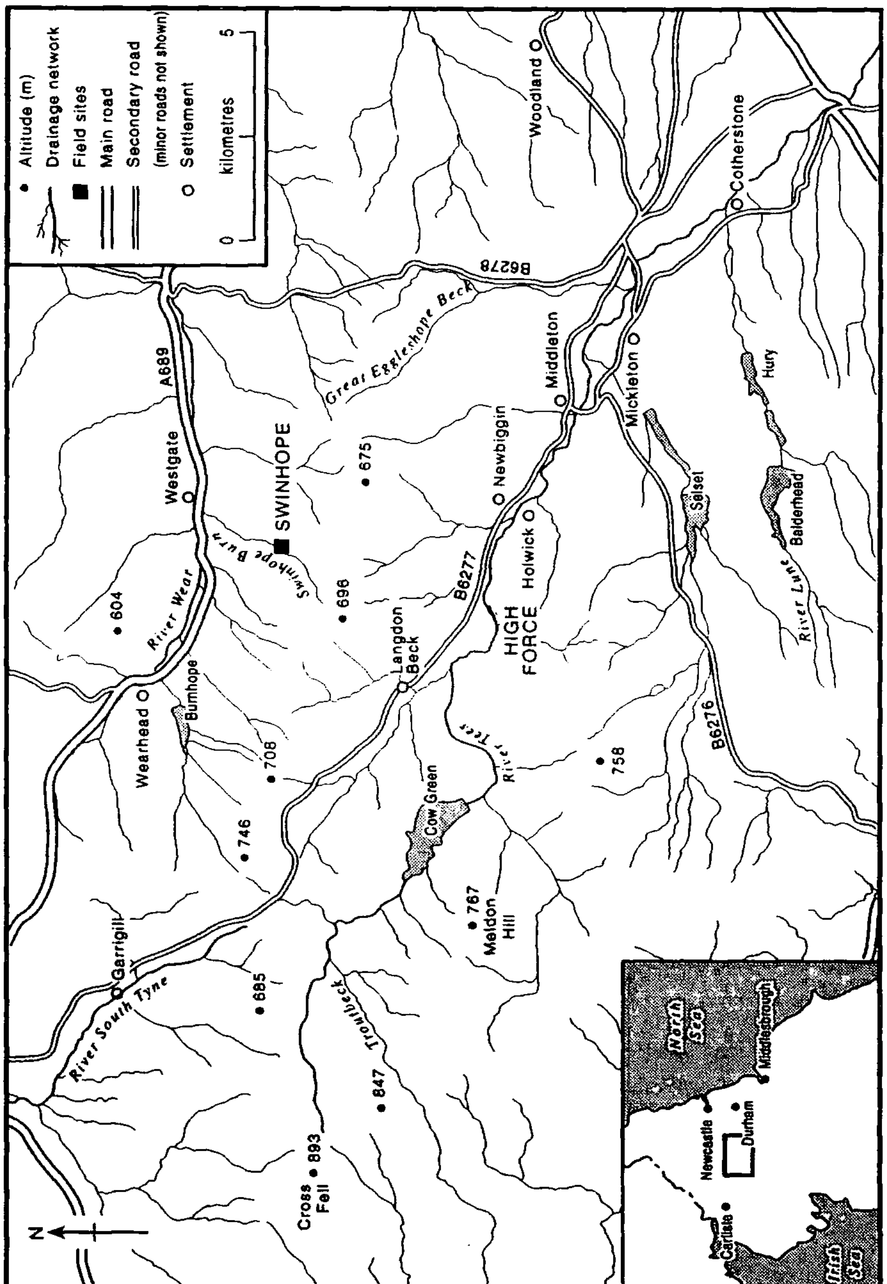


Figure 3.1 Location map of Swinhope Burn

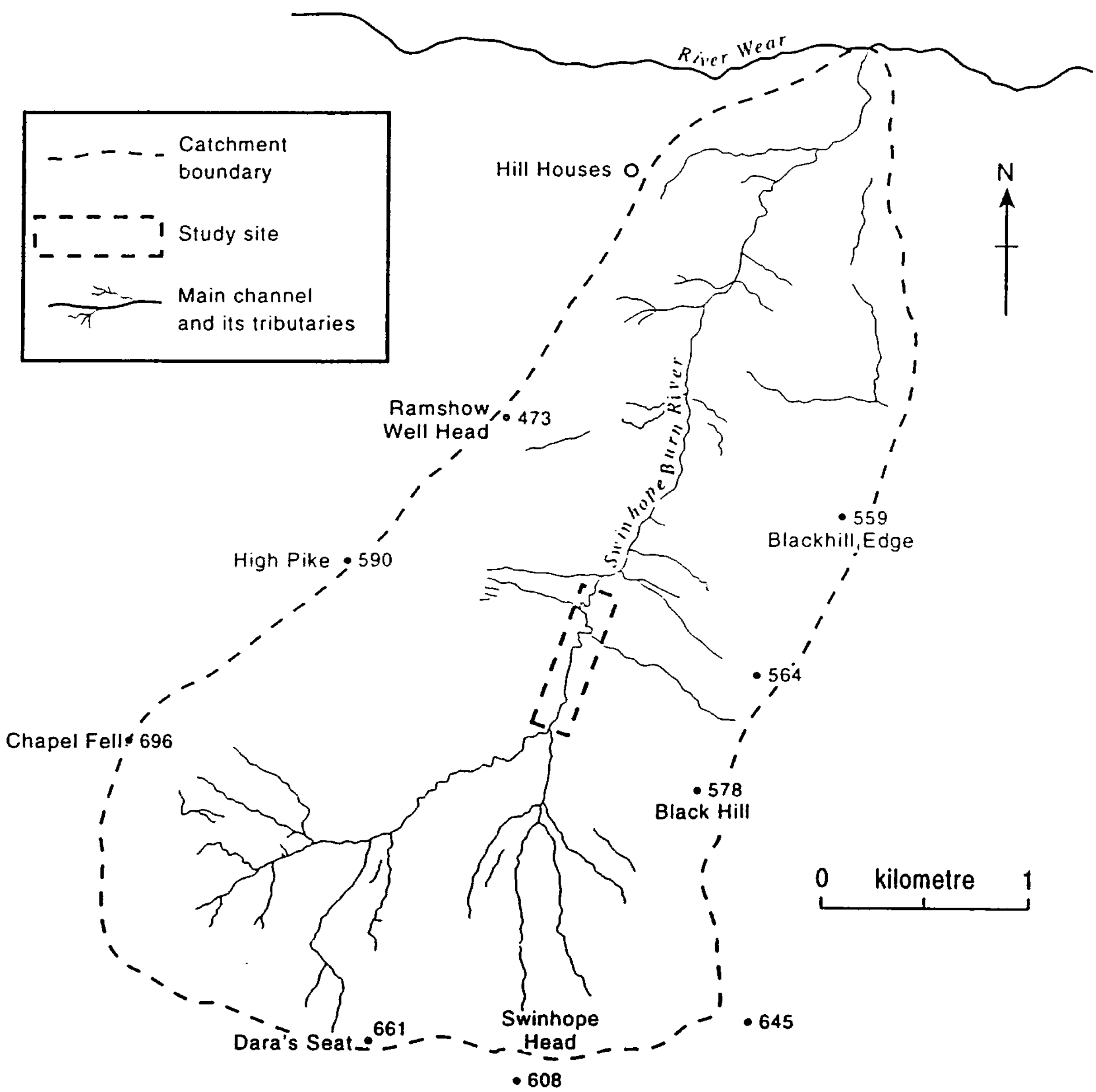


Figure 3.2 Swinhope Burn catchment area and study reach

Figure 3.3 Swinhope Burn basin looking upstream towards Swinhopehead.

Figure 3.4 Swinhope Burn basin looking downstream towards Greenly Hills Moraine (GHM)

3.1.3 Glacial history of Weardale and Swinhope Burn

The development of rivers in Weardale throughout the post-glacial has been largely determined by glacial erosion and sedimentation resulting from the last and earlier glaciations. In upper Weardale there is evidence of reworking of thick meltwater deposits of sands and coarse gravels in the valley, resulting from the retreat of ice at the end of the Devensian glaciation. The present day River Wear has deeply entrenched its valley floor forming a sequence of well-developed terraces.

Recent work on the glacial geology and geomorphology of Weardale (Moore, 1994) has identified the Greenly Hills Moraine within the Swinhope Burn basin (NY 896 363 to NY 903 361), 0.75 km downvalley of the study site, which appears to have been deposited by Pennine ice moving into Swinhope Burn from the north-west during the Late Devensian glaciation. The ice removed blocks of strata from the hillslopes and formed a 'push' moraine across the valley at Swinhope, creating the upland basin which now forms Swinhope Bottoms. The core of the Greenly Hills Moraine is made up of a series of imbricate and overthrust beds several metres thick, which are tilted. At a height of 430 m O.D., it forms a steep, arcuate, north-facing, till covered slope which rises to between 30m and 60m above the valley floor.

The unusually level section in the long-profile of Swinhope Burn basin (Figure 3.6) may be partly explained by a temporary lake which existed at this location during the late Devensian glaciation (Moore, 1994). This was formed by ponded water marginal to ice occupying the lower part of the valley below the present day Greenly Hills moraine. Although Moore (1994) found no evidence of lake clays, Moore suggests that incision through the moraine and formation of the 60m deep gorge may have resulted from the escape of meltwater from the basin as reverse drainage became established.

Moore (1994) identified sands and gravel on the valley slopes of Swinhope Burn which appear to have been deposited by meltwater. On distal slopes, deposition appears to have been in shallow channels which braid between mounds of till. These grade into a level spread of sand and gravel across the floor of Upper Swinhope Burn. On the proximal valley slopes of Swinhope Burn, slump deposits were found to cap and be interbedded with glaciofluvial gravel, the deposits including blocks of Great Limestone interbedded with slumped shale and sandstone from the underlying strata.

3.1.4 Climate

The climate of the Northern Pennines can be described as 'Upland Maritime' which is characteristically cloudy, cool and wet. The upper Wear tends to be sheltered from the full strength of the predominantly rain-laden westerly winds by Cross Fell (NY 687 344, 893m), the highest summit of the Pennines which has an annual rainfall total of 2200mm. Swinhope Burn basin, at an altitude of 700 metres, receives relatively high winds and an annual rainfall of approximately 1400 mm (Archer, 1992).

There are often long periods of cyclonic rainfall during the winter months in addition to brief, intense convective summer storms which can produce sudden flash floods owing to the flashiness of the river regime. In winter, precipitation frequently falls as snow and Swinhope Burn basin and river channels become partially frozen. Although Swinhope Burn is ungauged a combination of snowmelt and rain tend to produce floods.

3.1.5 Vegetation

The low gradient peaty floodplain through which Swinhope Burn flows mainly consists of coarse gravels and cobbles, particularly in the upper reaches, which is overlain by a thick layer of finer alluvium. The vegetation provides rough grazing for cattle and sheep with parts of the basin comprising open moorland. Throughout the basin there is some evidence of palaeochannels which are sometimes reoccupied at high flow. These are highlighted by reeds which pick out the former river channels and are more evident in the lower part of the basin where the floodplain morphology becomes increasingly uneven. The channel banks are predominantly turf covered, although in places the channel boundary is either formed by dense reed beds or is tree-lined, the latter occurring at the head of the study reach. In places the banks have been destroyed by grazing cattle (poaching), which is particularly apparent during the summer months.

3.1.6 Streamflow and rainfall

The River Wear, with a catchment area of 171.9 km^2 is currently gauged by the Environment Agency at Stanhope in Weardale (NY 984 391) at an altitude of 202 m OD which is the nearest gauged flow site to Swinhope Burn (Figure 3.5). A mean daily flow of $3.67 m^3s^{-1}$ and a mean annual flood of 121.9 m^3s^{-1} are recorded at Stanhope. The highest mean daily flow on record of 155.1 m^3s^{-1} was recorded at the Stanhope gauging station on 31st January 1995, with a peak flow of 286.97 m^3s^{-1}. Geomorphological evidence suggests that this event produced a flood of considerable magnitude which passed through the basin of Swinhope Burn.

Flow records are also available for the adjacent catchment of Rookhope Burn gauged at Eastgate, (NY 952 390) 6.5 km north-east of Swinhope Burn basin and at an altitude of 241m OD, although records have been discontinued since 1980. Although Swinhope Burn is ungauged, it can be described as having a 'flashy' hydrological regime with an estimated bankfull discharge of 2.5 m^3s^{-1} (Warburton and Danks, 1998). However, overbank flows occur approximately two to three times per year, particularly during the winter months, in response to frontal rain and rain combined with snowmelt.

Rainfall is currently recorded at Burnhope Reservoir (NY 850 391), Allenheads (NY 858 455) and High Greenwell, the nearest rain gauge to Swinhope Burn, (NY 860 383) with an annual rainfall of 1800 mm recorded on the adjacent Burnhope Seat (NY 850 391). A mean monthly rainfall total of 1301 mm is recorded for the Stanhope station at Tunstall Reservoir (NZ 064 408) at an altitude of 220 m.

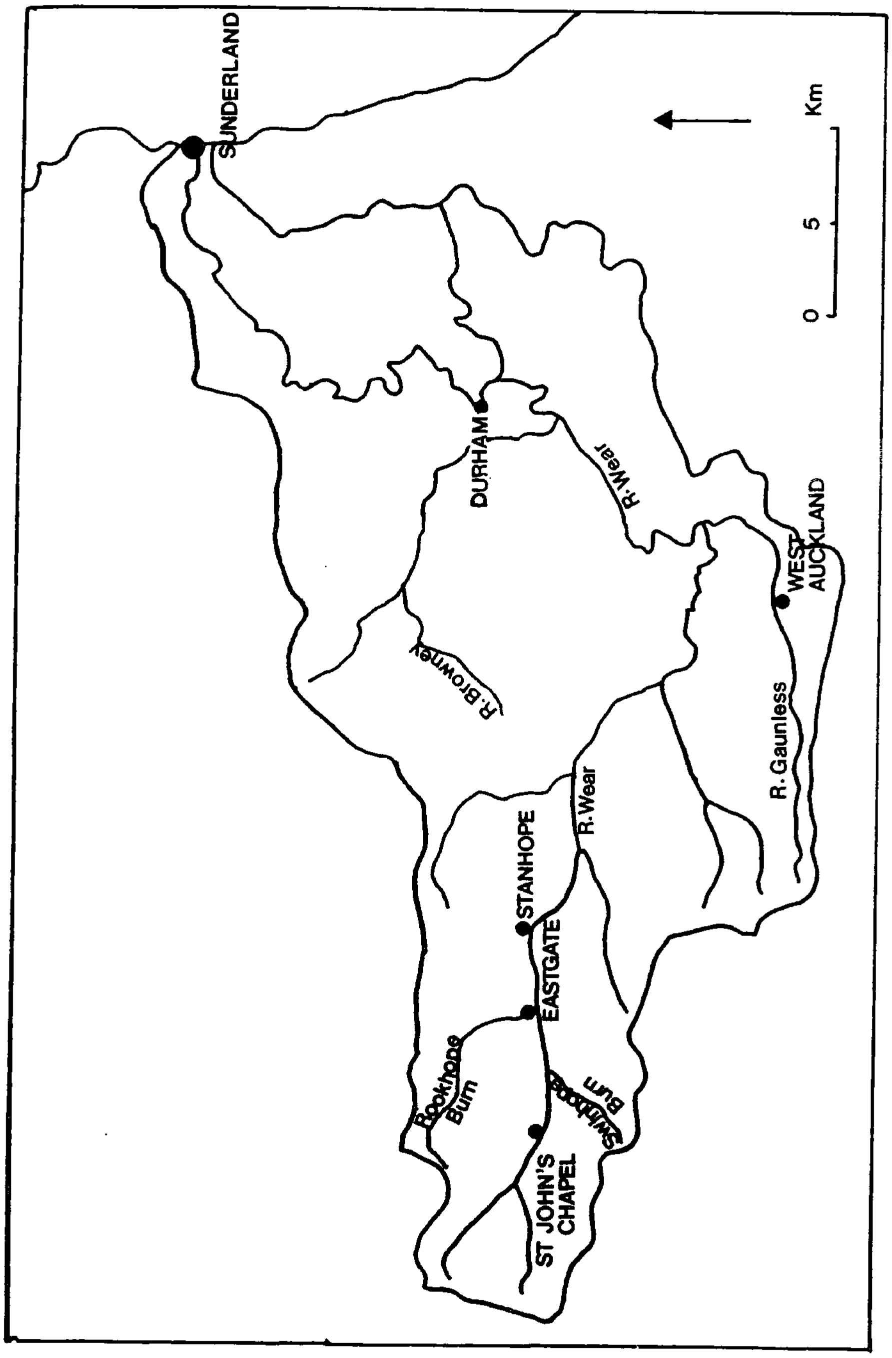


Figure 3.5 Map of River Wear catchment (based on Archer, 1992)

3.1.7 Long profile of Swinhope Burn and adjacent catchments

Swinhope Burn basin is unusual in that it does not have the characteristic convex-concave longitudinal profile exhibited by other upper Weardale tributary streams (Figure 3.6). Swinhope Burn is the exception because it has a distinct step in its middle reaches, formed by partial valley closure by the Greenly Hills moraine. The mean channel slope for Swinhope Burn is steeper than the other four catchments (Table 3.1) which is particularly evident in the upper reaches above the study site. The study site is located where the profile begins to level out at an altitude of 410 m, dropping to 400 m, 0.5 km downstream. The channel slope remains relatively flat for the remaining 1 km of the study reach and for a further 0.5 km before the channel enters the incised gorge of the Greenly Hill moraine (NY 896 363 to NY 903 361) (Figure 3.7). Although Swinhope Burn has incised through the Greenly Hills moraine it retains a small floodplain area above this local base level control.

Burn	Catchment area (km^2)	Stream length (km)	Mean channel slope	Total stream length (km)	Drainage density ($km\ km^{-2}$)	Basin Relief
Swinhope	10.5	5.9	0.07	21.7	2.07	0.09
Snowhope/ Horsley	8.2	5.78	0.06	24.3	2.97	0.09
Westernhope	13.9	6.45	0.05	34	2.45	0.06
Bollihope	30.3	12.48	0.04	93.5	3.09	0.05
Rookhope	37.6	15.1	0.03	100.6	2.68	0.04

Table 3.1 Catchment characteristics - Swinhope Burn and adjacent catchments

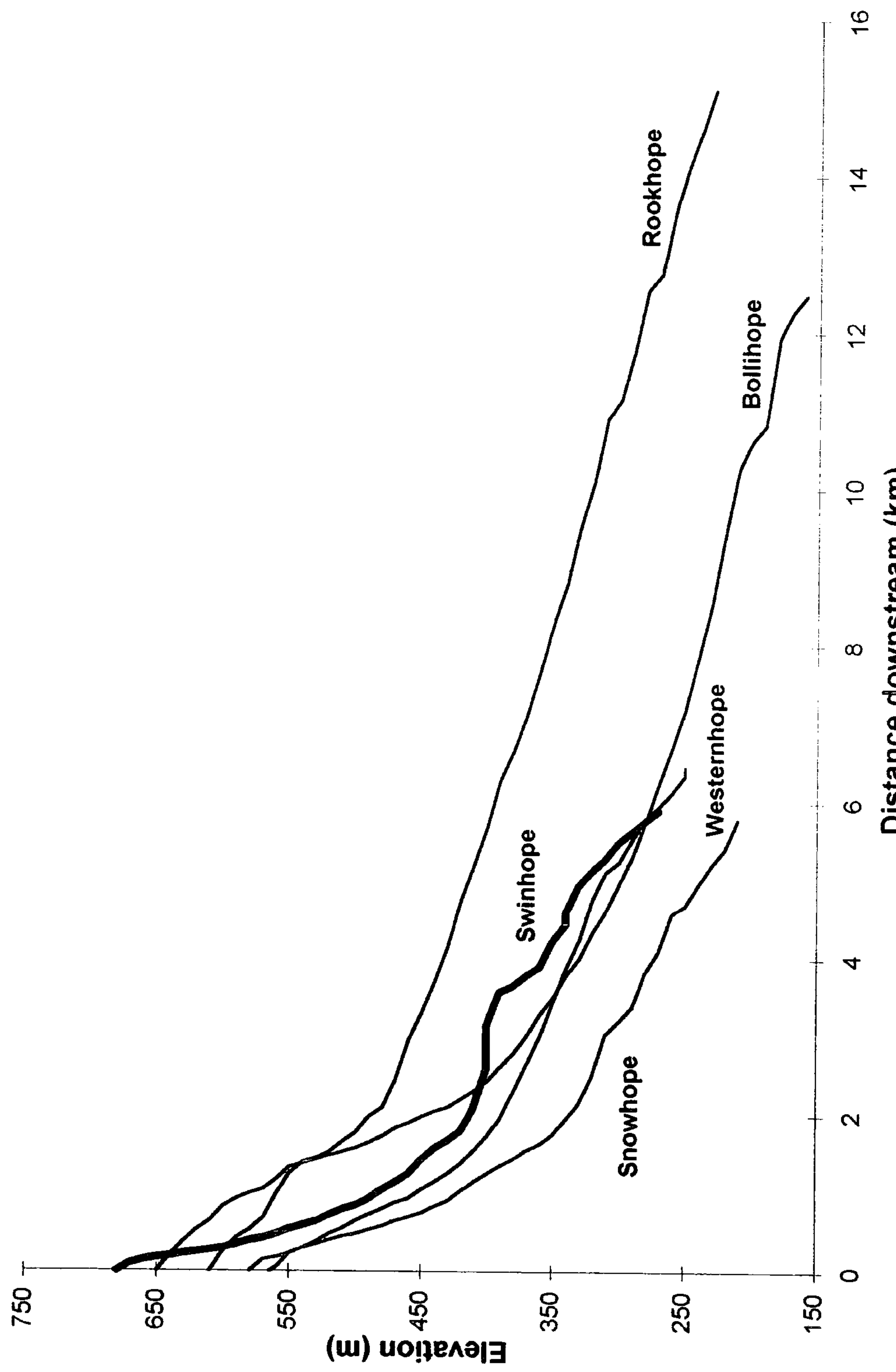


Figure 3.6 Long profile of Swinhope Burn and adjacent catchments

Figure 3.7 Swinhope Burn entering a gorge through the Greenly Hills Moraine showing upstream development of a small floodplain area. Flow is from right to left.

3.1.8 History of land-use change and mining activity in the Swinhope Burn Basin

Over the past decade, research has clearly identified the importance of anthropogenic changes and, in particular, large-scale mining activity on river channel and floodplain development within the Northern Pennine Orefield (Aspinall et al, 1986; Macklin, 1986, Stimpson, 1987; Macklin and Lewin, 1989; Passmore and Macklin, 1993). Increased inputs of coarse mining waste into many Pennine streams during the 18th and 19th centuries, and the associated process of hydraulic mining termed 'hushing', has been linked to historic floodplain sedimentation and the switching of channels from a meandering to a braided pattern. Inputs of fine metal-rich wastes were phytotoxic which impaired vegetation growth reducing bank and bar stability which also encouraged braided channel patterns to develop (Macklin and Rose, 1986). Headwater streams in the mining areas of upper Tynedale, Weardale and Teesdale were most affected.

For 200 years the Beaumont Company and the London Lead Company developed the lead mining industry in the Northern Pennines. The Beaumont company was responsible for the substantial development of lead mining and smelting in Allendale and most of Weardale (Dunham, 1948; Burt, 1984). The 18th century saw the greatest development of lead mining and smelting with over 7500 tonnes of lead being mined from the Northern Pennines Orefield in 1772. The end of the 19th century saw the decline of lead and iron mining in this area and The Weardale Lead Company, which took over from the Beaumont Company in 1884, finally ceased production of lead and iron ore in Weardale in the 1930's.

Swinhopehead Mine (NY 877 342) is situated where the Swinhope Cross Vein crosses the headwaters of Swinhope Burn, 1.2 km south-west of Swinhope Bridge on the Newbiggin to Westgate Road, 2 km upstream of the study reach. The mine is currently disused, but there is evidence that lead ore was mined from the main fault (Dunham, 1948). The Beaumont Company records state that the Swinhope Vein was about three and a half feet wide, with 'free spar thinly mixed with ore', with large amounts of silaceous fluorspar present. The only record of production is 326 tons of galena (sulphide of lead) in 1823-46, although a trial for fluorspar was made by G.G. Blackwell & Son in 1905 but no production is recorded.

Although some streams within the Northern Pennine Orefield have undergone substantial change as a result of large-scale mining activities, for example, the River Nent near Alston (NY 748 467) and Black Burn, a tributary of the River South Tyne (NY 699 427), it appears that the Swinhope Burn basin has not been affected by mining activities to the same extent.

Field evidence of mining activity within the study site is limited. However, approximately two-thirds of the distance along the study reach, a very clear exposure of alluvium shows a deep layer of clay material present at the top of the sequence overlying a layer of older river gravels which may be associated with local historic lead mining. Additionally, an abandoned channel preserved in the floodplain in the upper reaches of Swinhope Burn, identified on a topographic map dating from the period of mining activity, may provide additional evidence of anthropogenically-induced river planform change.

The present land-use in the Swinhope Burn basin is rough sheep grazing, and cattle grazing during the summer months, and it is unlikely that this has altered over the historical period. According to research on the recent history and land-use in Weardale, based on two small peat bogs proximal to the Swinhope Burn basin, (Roberts et al, 1973), during the early nineteenth century extensive tracts of the lower portions of the moorlands and pastures in Weardale were enclosed. After 1816 pine and spruce woodlands were planted. Since the end of the Second World War, woodlands have been planted by both the Forestry Commission and private ventures. Evidence from Tithe Maps suggests that the majority of the land in the Stanhope Parish was either meadow and pasture or uncultivated moor. It therefore seems probable that the history of agricultural land-use change in the Swinhope Burn has not changed significantly over the past 200 years.

3.2 FIELD METHODS AND TECHNIQUES

This section describes the methods and techniques used to collect field data. The main methods include establishment of 101 cross-sections along the length of the study reach and survey of channel form; survey of channel gradient and planform; sampling of bed and bank material; and a sediment tracing experiment using magnetic and painted tracer pebbles.

3.2.1 Survey of cross-sectional form

The repeated levelling of transects across the channel, plane-table mapping and low angle photography are common methods of establishing changes in the cross-sectional form and channel planform in response to flood events (Werritty and Ferguson, 1980; Newson, 1980; Werritty, 1982; Werritty and Ferguson, 1983). These methods identify the extent of aggradation and degradation within the channel and widening of the channel due to bank erosion resulting from flood events. Similar techniques were used in this fieldwork.

In order to examine downstream variations in cross-sectional form in response to floods, a network of 101 monumented cross-sections were pegged on the right and left banks of the 1.4 km long reach of Swinhope Burn (Figure 3.2). The end points of the sections were marked with wooden pegs and a metal pin at every 10th section.

The cross-sections are evenly spaced approximately every 15 metres along the length of channel. The cross-sections were evenly spaced by measuring with a tape from one peg to the next along the left bank, following the channel boundary as close as possible. This section interval (15 m) is six times the average channel width of 2.5 metres although there is significant local variation in channel width. Although this spacing of cross-sections potentially corresponds with the pool-riffle spacing of five to seven times channel width proposed by Milne (1979) and Harvey (1975) for upland streams, the spacing of cross-sections at 15 m at Swinhope Burn, in practice resulted in the sampling of a complete range of channel sites. Close spacing of cross-sections enables the effects of localised variations in bed configuration, bank material and sinuosity on changes within cross-sectional form following a flood to be determined.

The study reach starts at an altitude of 406 metres and is located where the step in the long profile of Swinhope Burn begins (Figure 3.6). The reach ends at an altitude of approximately 389 metres at the point at which the gradient begins to fall steeply as the stream incises through the Greenly Hills moraine (NY 912 356) (Figure 3.4). The average overall slope of the channel is 0.0118. The 101 cross-sections cover almost the entire step within the profile, therefore the study reach encompasses a distinctive segment of Swinhope Burn.

The geometry of the 101 cross-sections was initially determined by levelling and tied into a local benchmark. The sections were re-levelled following a major flood event in order to identify the extent and nature of changes in bankfull width and bankfull depth as a result of erosion and deposition within the channel. Typical leveling errors are estimated at +/-1 cm for vertical dimensions and +/-5 cm for horizontal dimensions.

3.2.2 Variables used to describe cross-sectional dimensions

The primary variables used to describe the cross-sectional geometry are bankfull width, maximum bankfull depth and bankfull width to depth ratio (Figure 3.8). Re-levelling of cross-sections following the flood allowed the areas of erosion and deposition within the channel to be calculated for both the bed and banks.

Bankfull channel dimensions are used as a measure of cross-sectional geometry since river channels are on average adjusted to the 'dominant' flow which just fills the available cross-section and which is largely responsible for shaping the channel. However, it is appreciated that this approach has its limitations, since there is no consistent method for specifying the bankfull channel. In addition, bankfull flow may not be the most effective flow regarding sediment transport. Difficulties in definition of bankfull width arose due to inequality in the elevation of opposite banks at some sections. In these cases bankfull width is taken as that of the lower bank since lateral spillage from the channel would take place at this level leading to an abrupt increase in the width of the channel. The presence of bars within the channel and riparian vegetation helped to establish bankfull width. Bankfull depth was recorded at the deepest part of the channel on the line of the cross-section.

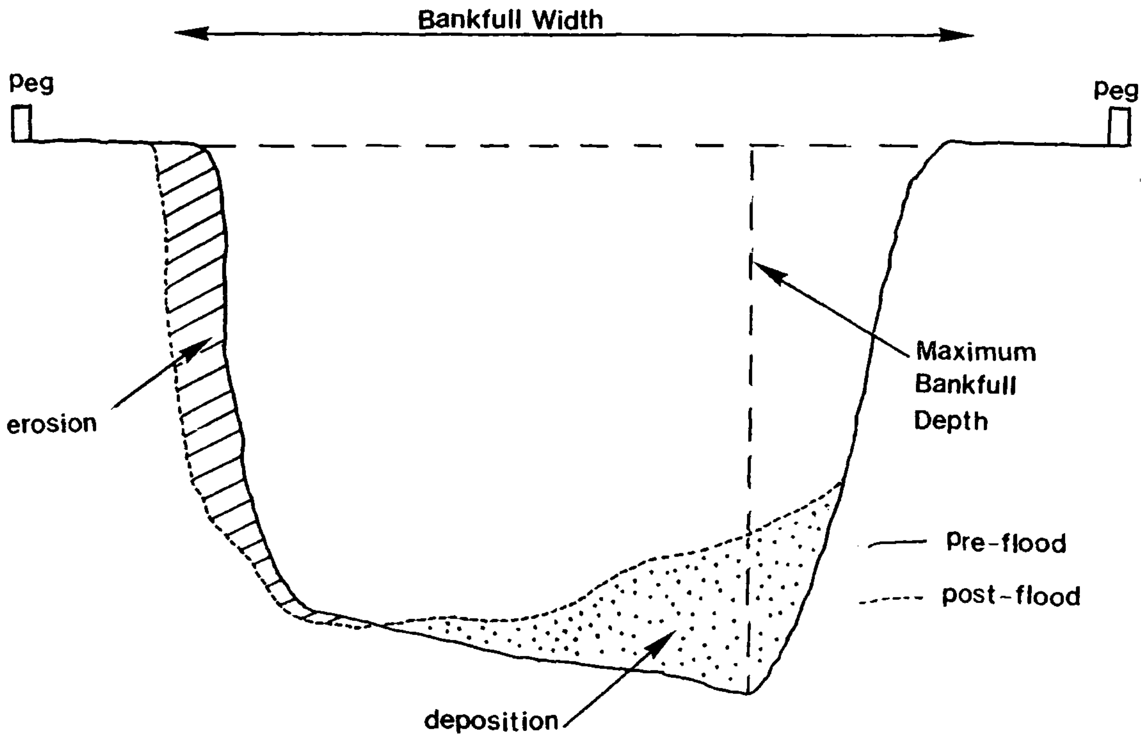


Figure 3.8 Primary variables used to describe cross-sectional geometry

3.2.3 Survey of channel gradient and planform

A detailed long profile of the study reach of Swinhope Burn was levelled and tied into a local benchmark. At the majority of cross-sections right or left pegs were additionally surveyed in order to provide reference points along the channel. The lowest points along the channel bed were surveyed and survey points were chosen in order to provide a detailed representation of variations in bed configuration and in particular the pool-riffle and meander bend sequence downstream.

Channel planform was mapped using an air photograph of Swinhope Burn taken on 7th September, 1991 (Aerofilms). Using the air photograph, each bend was identified and labelled (Figure 5.13, Chapter 5). In order to identify downstream variations in bend sinuosity a bend sinuosity index (P_B) was calculated as the ratio of channel length/meander wavelength.

3.2.4 Grain-size sampling of bed material

An initial grain-size survey was completed in order to identify the existing composition of the bed material and to examine whether there is a significant downstream trend in grain-size. The purpose of the initial survey was to provide a benchmark from which it would be possible to record any changes in grain-size resulting from subsequent high flows. The initial survey identified six grain size classes which were representative of the bed material of Swinhope Burn (Figure 3.9). The bed material in the channel covers a wide range of grain-sizes from fine gravel to small boulders. Re-sampling after a major flood event allowed downstream variations in grain-size to be identified.

To provide a measure of change in the coarsest grain-sizes present on the surface of the channel bed, the b-axis of the coarsest 20 particles was measured in a m^2 quadrat placed along the centreline of each of the 101 cross-sections in a mid-channel position. The metre square grid was placed approximately mid-channel. The method of sampling used in the grain-size survey is a variant of Wolman's (1954) grid sampling technique. The grid method of sampling as identified by Wolman (1954) is the most commonly employed surface method. The 'Wolman method' is highly adaptable in that it can be easily modified to the scale of the sampling programme so can be used for small-scale sampling on headwater tributaries, for example on the Kingledoors Burn (Milne, 1982b) or for large-scale bed material sampling projects such as for the Ashley River in New Zealand (Mosley and Tindale, 1985). However, the method can be very time consuming and 'operator errors' in sampling can result in significant differences between sample and population parameters, particularly when more than 100 pebbles are sampled or if a number of 'operators' are sampling. However, other methods can be used such as photography (Newson and Leeks, 1987), areal sampling (Gomez, 1983) or volumetric sampling (Church et al, 1992; Carling and Reader, 1982) depending on the level of accuracy required and the calibre of bedload to be sampled.

From the data mean grain-size was calculated for each of the cross-sections. It is appreciated that a sample of 20 particles may not be a good estimate of the mean grain-size. However, this size of sample was chosen because it allows a rapid survey to be undertaken in the immediate post-flood period and by selecting the coarsest particles ensured a consistent sampling method.

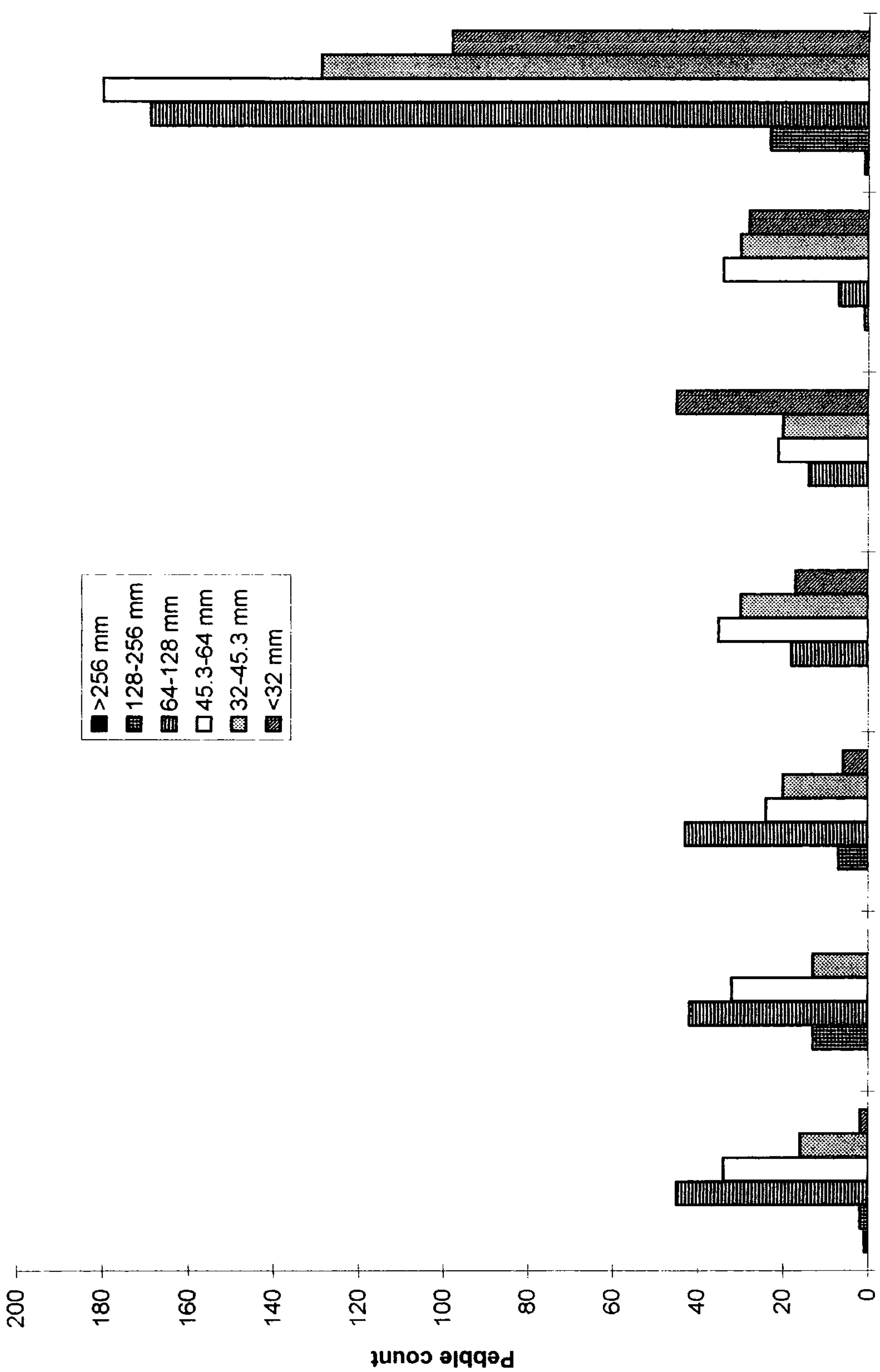


Figure 3.9 Grain-size classes representative of bed material at Swinhope Burn

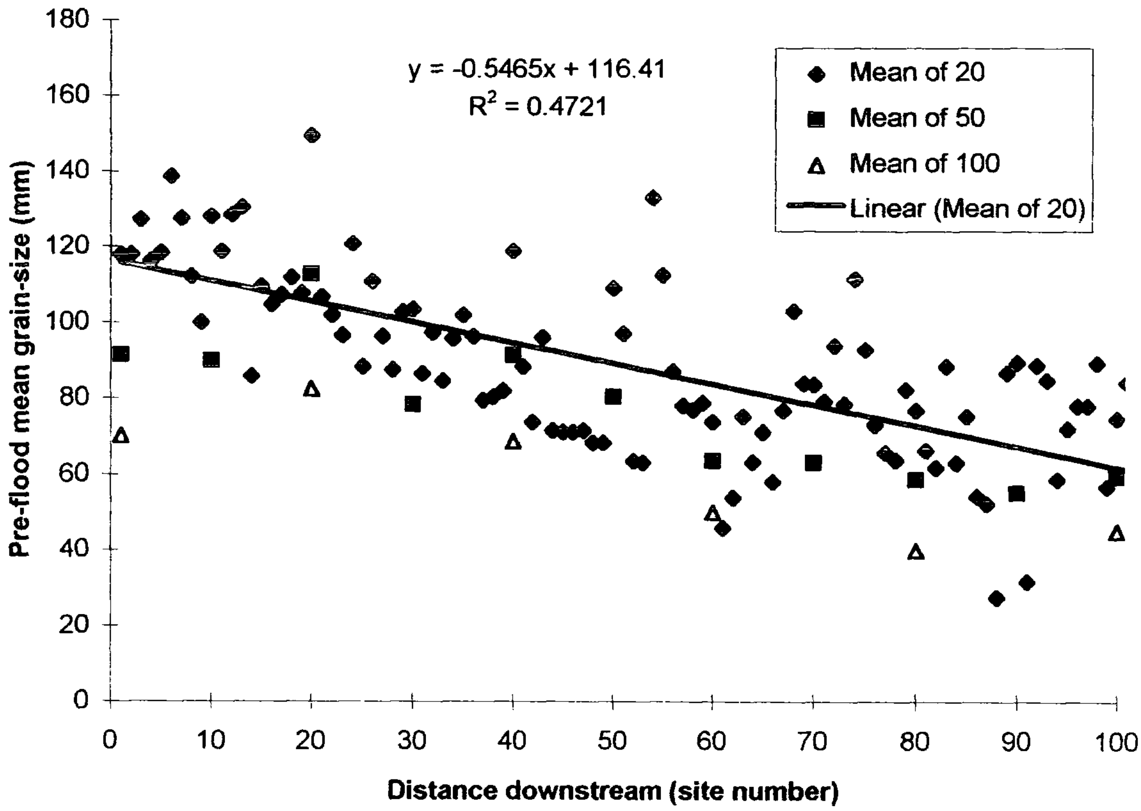


Figure 3.10 Downstream variations in mean-grain size of bed material at Swinhope Burn

In order to evaluate the importance of sample size on the results of the survey, at 11 sections the coarsest 50 particles, and at six sections the coarsest 100 particles, were sampled. It was found that in a narrow channel like Swinhope Burn, and particularly in the lower reaches, a sample of 100 particles can represent 20-50% of the channel width. The smaller samples of 20 particles showed a dominance of cobble size material whereas the largest samples of 100 had a mean grain-size of coarse gravel. The samples of 50 particles plot mid-way between the two. This is illustrated in Figure 3.10 where the sites at which 100 particles were sampled show lower mean grain-sizes than at sites at which the coarsest 20 or 50 particles were sampled. In other words, smaller samples tend to be biased towards the coarser particle sizes when selecting the largest particles in each section.

3.2.5 Grain-size sampling of bank material

In order to determine the relationship between bank cohesion and changes in cross-sectional form resulting from major flood events, bank material was sampled and classified according to the Wentworth grain-size classification for the right and left banks of each of the 101 cross-sections. The Wentworth grain-size classification was adapted (Table 3.2) in order that the bank material data could be analysed and presented in graphical form. The grain-size of bank material sampled ranged from clay to very large boulders. In cases where there were one or more grain-size classes present in the sample, for example, in composite banks, all were recorded. Although this method of classifying bank material lacks precision when measuring fines, it allows rapid field survey and its accuracy is sufficient for the purposes of this study.

Class	Bank material type
1	clay
2	silt clay
3	sandy silt clay
4	sandy silt
5	fine sand
6	medium sand
7	coarse sand
8	fine gravel
9	medium gravel
10	coarse gravel
11	small cobble
12	mixed cobble
13	large cobble
14	small boulders
15	medium boulders
16	large boulders
17	very large boulders

Table 3.2 Grain-size classification for bank material based on Wentworth grain-size classification

3.2.6 Sediment Tracer Experiment - laboratory and field techniques

Tracer pebbles were sampled randomly from the surface of the riverbed at five experimental sites and were grouped into three-grain size classes (32-45.3, 45.3-64, 64-128 mm). The pebbles were initially scrubbed clean in order to remove any fine sediment and algae and allowed to dry in order to ensure that the paint was less easily abraded once the pebble was returned to the stream. The b-axis of each pebble was re-measured and placed into one of the three grain-size classes. Pebbles taken from each of the five experimental locations were kept separate to ensure that the natural bed material was returned to its former stream location.

Initially, all 700 tracer pebbles were covered in water-resistant white paint. Subsequently, groups of tracer pebbles sampled from the four cut-bank locations were each painted with a distinct colour in order to identify their original location subsequent to transport (Figure 3.11). In the laboratory, the tracer pebbles selected for the magnetic tracing experiment were drilled and a ferrite magnet (Alcomax) measuring 20mm in length with a diameter of 6mm was inserted. The cavity in each pebble was refilled with epoxy resin. Finally, all 700 pebbles were sequentially labelled with a number in black paint to aid the re-identification of each pebble following transport (Figure 3.12)

At the beginning of March, 1997, all the tracer pebbles were re-introduced to the channel in their original location. The magnetic tracer pebbles were firmly lodged within the channel bed. They were placed in rows across the width of the channel and extended up to four metres upstream of section 1. Each of the four groups of painted tracer pebbles was lodged at the base of each experimental cut-bank. The painted tracer pebbles were partially embedded within the base of the cut-bank in a similar way to the existing pebbles.

The movement of both painted tracer pebbles and magnetic tracer pebbles within and through the study reach were monitored in order to identify the importance of the relative size of a flood event in the entrainment, transport and deposition of the tracers.

Figure 3.11 Tracer pebbles painted in four distinct colours

Figure 3.12 Tracer pebbles sequentially labelled with a number in black

A 'Magna-Track 100' magnetic locator was used to survey the position of the magnetic clasts (Figure 3.13). Although many magnetic tracers were still visible on the surface of the bed, using the magnetic locator it was possible to locate magnetic tracer pebbles buried within the bed. Painted tracer pebbles which had been transported downstream by the flood were visually re-identified according to their colour.

The new positions of the tracer pebbles within the channel bed, following the flood, were located and recorded by measuring the distance from two of the monumented cross-section pegs present on the right and left banks of the channel.

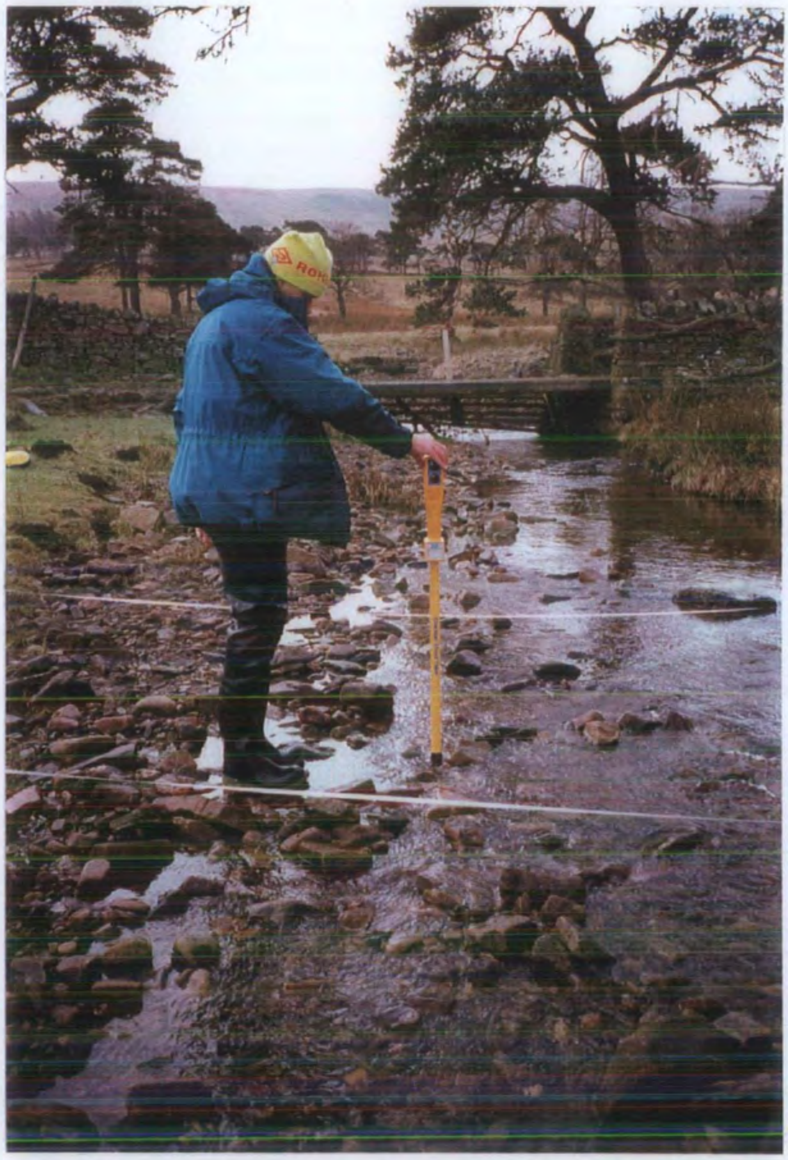


Figure 3.13 'Magna-trak' 100 magnetic locator used to survey the position of magnetic clasts following the January, 1998 flood

CHAPTER 4

HISTORICAL RIVER CHANNEL CHANGE

4.1 Introduction

The assessment of channel planform change at the reach scale, over the historical period, is useful in understanding contemporary channel processes as it is important to place present rates of channel adjustment into a longer term context. Upland gravel-bed streams have often been viewed as highly active with planform changes in response to floods being regarded as both frequent and ubiquitous (Ferguson and Werritty, 1980). However, although a contemporary stream channel may appear to be highly dynamic in its response to flood events, suggesting that over longer periods of time substantial changes in planform are likely to have occurred, scrutiny of historical maps and air photographs may prove the opposite (Warburton et al, 1993). Likewise, it is imprudent to assume that the past fluvial system is a simple extrapolation of the present one (Ferguson and Werritty, 1980; McEwen, 1989a), due to changing relationships between the channel and its environment over different timescales. Climatic fluctuations, a change in flood magnitude and frequency or land-use changes may result in significant changes in channel planform. This being the case, the study of river response to flood events must combine an understanding of both historical and contemporary river channel change within a reach. Although timescales over which historical river channel changes is examined varies according to the oldest available cartographic map, in this study the 'historical' timescale is defined as the last 180 years, commencing with the Inclosure Plan of 1815.

4.2 Scope of chapter

The objective of this chapter is firstly to summarise recent research into historic river channel change in the British uplands paying particular attention to the methods used to identify the nature, extent and causes of historic planform change and the timescales over which research has been undertaken. Secondly, evidence of historic channel planform change in Swinhope Burn, Weardale over the period of 180 years will be examined. Reasons behind the inherent stability of the study reach will be suggested. Thirdly, three hypotheses to explain the probable cause of a period of dramatic historic channel planform change will be discussed and tested with reference to field evidence, historic documentation and recent literature. Fourthly, the recent flood history of Swinhope Burn will be outlined. The chapter concludes with an explanation of the reasons for the inherent stability of the study reach of Swinhope Burn and identifies the most probable explanation for the period of river channel change at some point during the period 1815 to 1856.

4.3 Background

Research into the rates and patterns of historic channel change in the upland rivers of north-east England began with the pioneering work of Petts (1979) in the Rede valley and Milne (1982a) on Harthope Burn, Northumberland. This paved the way for further work in the Northern Pennines on the River South Tyne and its tributaries (Macklin, 1986; Macklin and Aspinall, 1986), which assessed the role of flood events and historic metal mining in channel and floodplain metamorphosis.

In the Northern Pennines, as in many other parts of upland Britain, river patterns were disrupted at the onset of glaciation on a number of occasions. Valley floors have a tendency to be narrow with floodplains and channels frequently being confined by bedrock or drift bluffs (Macklin and Aspinall, 1986). If a channel is located in a basin, historic lateral channel migration can be substantial incising through glacial deposits forming terraces and an alluvial valley floor. Although glacial sediments of basins and valley bottoms are currently being reworked by the present streams, the historic and contemporary development of channel planform is highly dependent upon their glacial legacy.

In recent centuries, however, the development of metal mining in the Northern Pennines has dramatically affected both the quantity and quality of sediment being introduced into local streams and in many cases has had a major effect on historic planform development (Macklin, 1997). However, since metal mining in the Northern Pennines is very well documented (Raistrick and Jennings, 1965) it is possible through trace metal dating to precisely determine the role of metal mining on river planform change (Macklin and Dowsett, 1989).

Historic river channel change has been the topic of research throughout the British uplands, often with the emphasis being on the role of historic flood events (Anderson and Calver, 1977, 1980; Tipping, 1994). In the Scottish uplands, where there is marked regional diversity in rates and periodicity of channel pattern change (McEwen, 1994) the reconstruction of the history of channel disruption and recovery in response to floods began with Werritty and Ferguson (1980) on the braided, gravel-bed River Feshie in the Cairngorms. Since then, the importance of flood events in determining river channel change in the highly active channels of the Scottish uplands has been demonstrated for the River Dee (McEwen, 1989a), Allan Water, Stirlingshire (Rowling, 1989), the River Coe (McEwen, 1994), and the upper Bowmont valley (Tipping, 1994). In Wales, the role of historic metal mining on river planform change has dominated research (Lewin et al, 1977, 1983). Pioneering work on historical river channel changes on the River Bollin (Mosley, 1975; Hooke, 1977) and on the River Dane, in Cheshire (Hooke, 1987) and more recently throughout the U.K. (Hooke and Redmond, 1992) are examples of the development of research in the piedmont zone.

Timescales for identifying the nature and extent of historic river channel changes varies according to oldest cartographic map available for a particular area. For example, in the Scottish uplands, the use of Roy's Military Survey of Scotland dating back to the mid 1750s can extend the record back 200 years (Werritty and Ferguson, 1980) provided its reliability can be confirmed by old estate plans (McEwen, 1989a). In the Northern Pennines channel planform has been well documented over a time period of 200 years for the River Nent (Macklin, 1986), which has the longest record of historic river channel change in northern England and over a period of almost 200 years at Black Burn, Cumbria, (Macklin, 1997a) since maps were drawn up in the mid to late 18th century for mining purposes. In many cases, studies of historic river planform change date back to the 1st Edition of the Ordnance Survey Map in the mid 1850's providing a timescale of around 100 to 150 years (Mosley, 1975; Milne, 1982a; Macklin and Aspinall, 1986; Tipping, 1994; McEwen, 1994). Anderson and Calver (1980) on Hoaroak Water, Exmoor and Werritty (1982), on Dorback Burn, Cairngorms, limited their timescales to the last 30 years solely utilising aerial photography to identify changes in channel pattern resulting from flash floods.

Maps and air photographs provide valuable detailed information on historic river planform changes which cannot be obtained from any other source. In the majority of cases air photographs are used for the post-1940 period with historic maps being used to extend the timescale up to 200 years. In studies where the aim is purely to determine the nature and extent of historic river channel change, maps and air photographs may be exclusively used (Mosley, 1975; Werritty and Ferguson, 1980). However, McEwen (1989a) identifies a problem with using maps and air photographs to determine whether there has been historic channel pattern change in that they only provide a set of 'stills' at any particular time and it is not known whether change in the intervening period had occurred progressively or abruptly. Therefore, in order to reconstruct the history of flood events and identify possible causes of channel planform change a wide variety of methods have been identified (McEwen, 1987; Trimble, 1998). Ideally, gauged discharge data or an established flood chronology used in conjunction with maps and air photographs is likely to provide the clearest indication of flood-induced river channel change (Milne, 1982a; McEwen, 1989a; Macklin et al, 1992a, 1992b; Rumsby and Macklin, 1994).

McEwen (1990), effectively used short-term gauged discharge records in conjunction with meteorological publications and local newspapers to establish a historical flood chronology for the River Tweed catchment, Scotland. However if gauged discharge data is not available or if a historic flood chronology can not be established, the study of rates of channel planform change over a historical time period in relation to fluctuations in the magnitude and frequency of flooding and anthropogenic changes is limited (McEwen, 1994). In these circumstances, the use of other methods to supplement map and air photograph material is needed. Rainfall data is an important source of information in the establishment of historic flood records (Rowling, 1989; McEwen, 1989a, 1989b; Rumsby and Macklin, 1994), whereas lichenometric analysis is particularly useful in dating flood deposits (Milne, 1982a; Macklin, 1986; Macklin and Aspinall, 1986; Macklin et al, 1992a, 1992b; Macklin et al, 1994; McCarroll, 1994; McEwen, 1994; Macklin, 1997(b)). The utility of trace metal dating of fine grained overbank flood sediments for dating flood events has been illustrated throughout the

Northern Pennines (Macklin and Aspinall, 1986; Macklin, 1986; Macklin and Dowsett, 1989; Macklin et al 1992a, 1992b; Passmore and Macklin, 1994), the Yorkshire Dales (Macklin, 1997b) and in Wales (Lewin et al, 1977, 1983). However, in many remote upland catchments gauged discharge data or rainfall data may not be available. If the catchment has been largely unaffected by historic mining activity, trace metal dating is inappropriate. In these cases local flood documentation may be the only method from which it is possible to identify the timing of large flood events which may or may not have been responsible for observed changes in channel planform over the historic period . However flood documentation is often used in conjunction with other methods in order to establish historic flood chronologies (Lewin et al, 1977; McEwen, 1989a, 1989b; Rowling, 1989; Macklin 1992a, 1992b; Rumsby and Macklin, 1994).

4.4 Historic river channel change on Swinhope Burn, upper Weardale, Northern Pennines

This chapter has two main objectives:

1. to determine the extent and nature of river channel change in Swinhope Burn over 180 years using historic maps and air photographs.

2. to identify the cause/s of river planform change using field evidence and local historic flood documentation.

4.4.1. Description of the Swinhope Burn study site

The study reach depicted in Figure 4.1 is a 1.4 km reach of Swinhope Burn, a right bank tributary of the upper River Wear in the Northern Pennines. This is a small, upland, gravel-bed stream flowing in a predominantly shallow channel along an irregular meandering course. The channel bed material covers a wide range from fine gravel to small boulders . The stream lies in an enclosed valley with high gradient slopes in the upper reaches and a steep 'push' moraine ridge downvalley of the study site. The basin occupied by the study reach is of low gradient (0.02) due to the presence of the Greenly Hills moraine which has produced a distinct stepped long profile. The channel banks are highly cohesive particularly in the lower reaches. There is very little

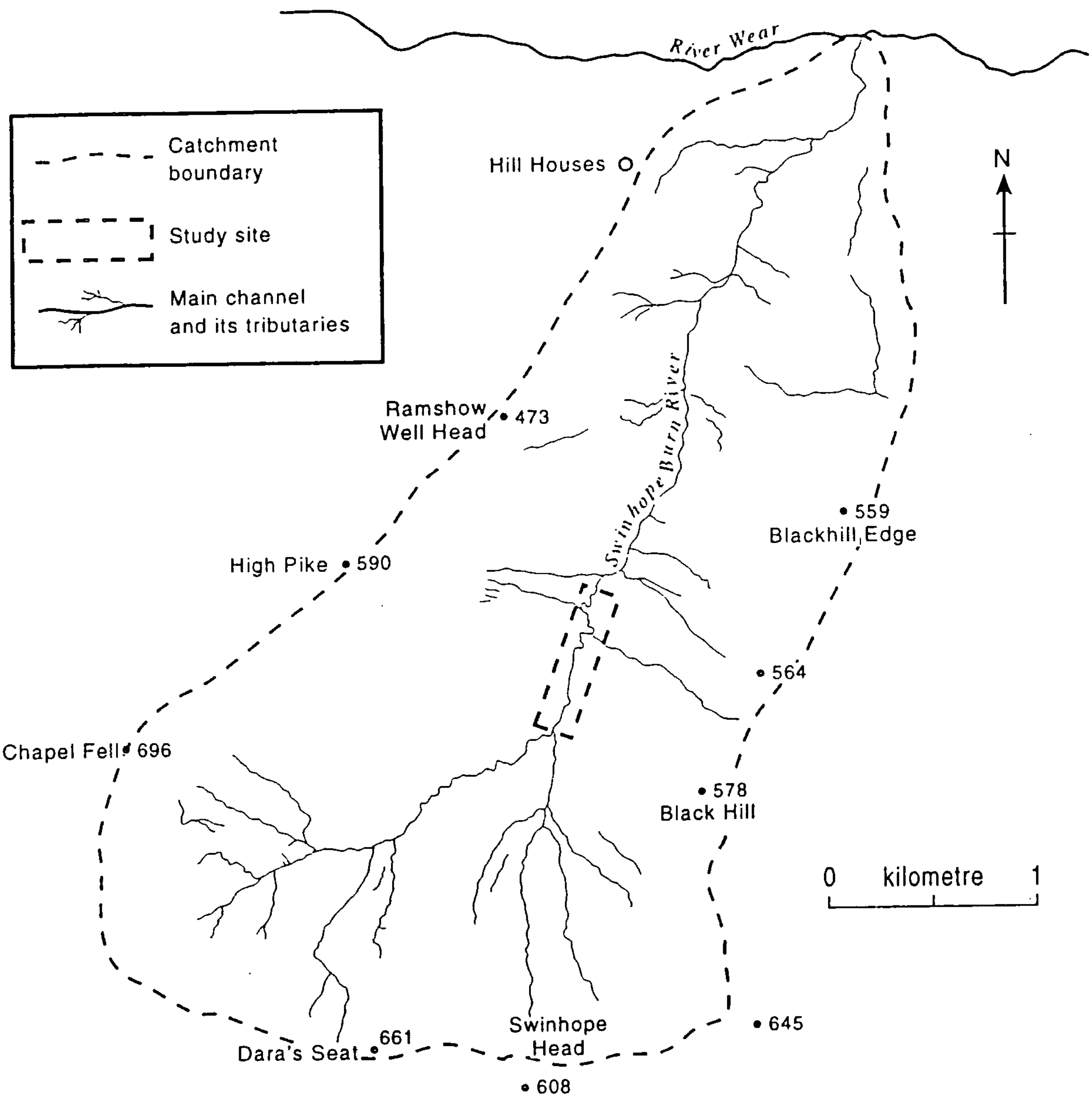


Figure 4.1 Swinhope Burn catchment and study area map

coupling between valley-side slopes and the channel, apart from coarse sediment input from occasional, small, valley-side bluff erosion scars.

4.4.2. Historical Records of Channel Planform Change

Figure 4.2 (Appendix A) shows the planform of Swinhope Burn between 1815 and the present day and shows variations in channel sinuosity. The lines represent the traces of the centrelines of the main channels as depicted on historical maps and air photographs. The eight historical maps and one air photograph enable changes in channel position and planform to be documented over a time period of 180 years. In addition to the 1991 air photograph planform shown on Figure 4.2 eleven other air photographs covering the study reach were taken in various formats between 1951 and 1995, but these were omitted since they showed no major evidence of river planform change in Swinhope Burn.

The earliest map, the Inclosure Plan of 1815 shows a meandering channel planform similar to the present planform of Swinhope Burn, although it is slightly less sinuous. However, by 1844, a dramatic metamorphosis of the channel planform had taken place. At the top of the reach there is a braid bar with a single thread channel of very low sinuosity downstream. By 1856, the First Edition of the Ordnance Survey County Series shows that the channel planform has reverted to a meandering pattern again, the main elements of which have persisted to the present day. Maps and air photographs published over the subsequent period of 100 years all have the same basic meandering pattern with only minor differences which may have resulted either from changes in cartographic techniques or the angle of air photography. Other minor changes in channel planform may have resulted from the natural evolution of meander bends which over the past 100 years has produced an increasingly sinuous channel.

In contrast to many dynamic gravel-bed rivers of the Northern Pennines which over the historical time period have undergone significant changes in channel planform as a result of large flood events (Macklin et al, 1992), large-scale historic mining (Macklin, 1986), land-use change (Macklin et al, 1994) or climatic fluctuations (Macklin et al 1992b) historic planform change in the Swinhope Burn basin has been insignificant. The study reach has been remarkably stable over the last 180 years with

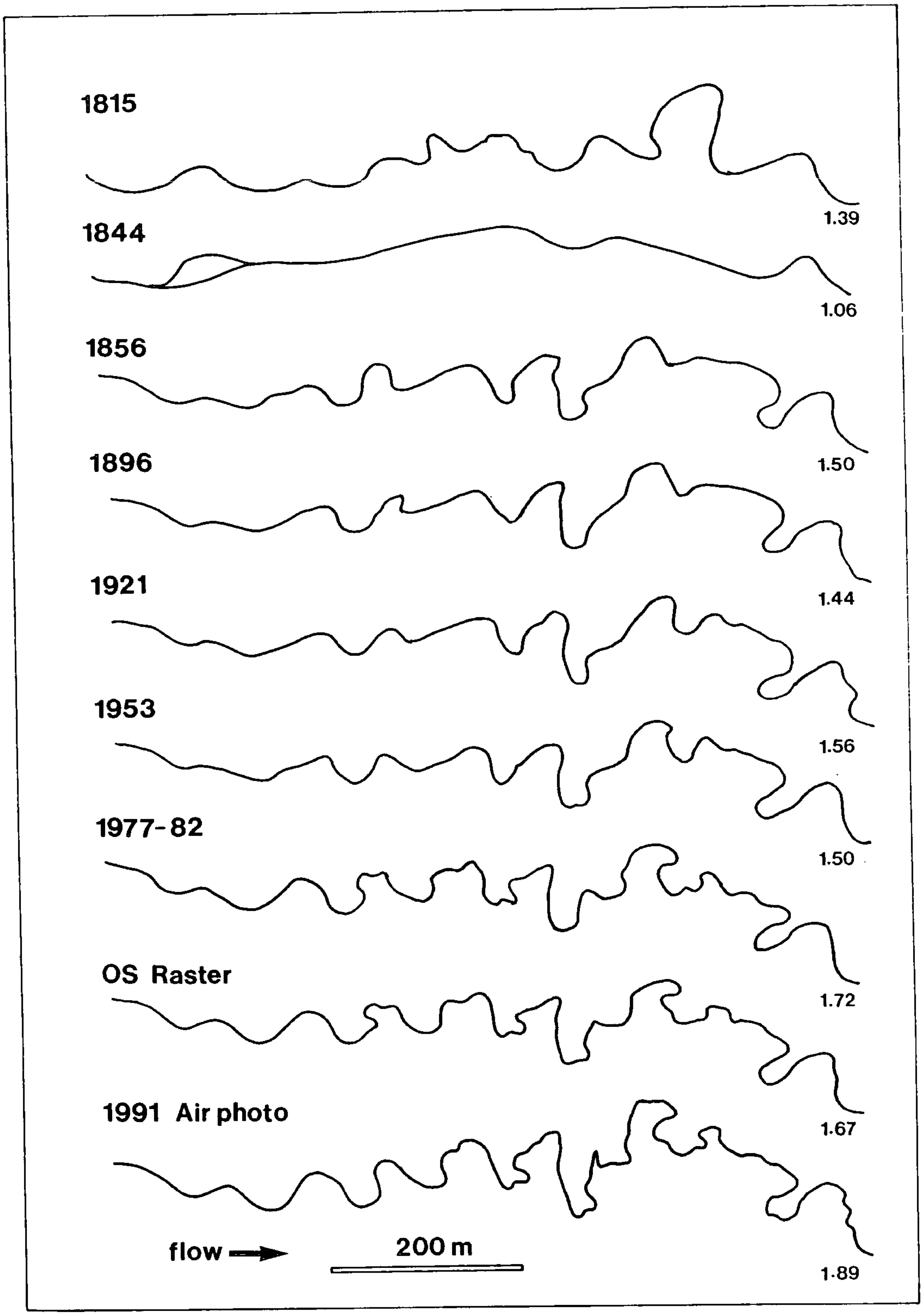


Figure 4.2 Historical evidence of river planform change at Swinhope Burn, Weardale. A full source of references is given in Appendix A.

only one documented occasion when, for a relatively short period of time, a dramatic change in channel pattern occurred.

4.4.3. Reasons for channel planform stability

The explanation for this remarkable stability in channel planform may lie in the rather unique geomorphic setting of the study reach. In comparison with other upper Weardale tributary streams, Swinhope Burn has a very distinct long profile with a step in its middle reaches (Figure 4.3). The presence of the Greenly Hills moraine situated at the lower end of the basin partially closes the valley system and has led to the development of a small upstream floodplain, where the study reach is located. Consequently the study reach is essentially a sedimentation zone.

Since the channel barely impinges upon the valley-side slopes, except at a few streamside scars, coupling between them is minimal. This reduces sediment supply to the channel and therefore the potential for river channel change. Equally important is the shallow nature of the channel slope within the study reach, which reduces the ability for sediment transport and has led to a significant reduction in the downstream grain-size distribution.

The banks are highly cohesive in the lower reaches which inhibits lateral erosion of the channel, and which may partly explain its historic lateral stability identifiable on Figure 4.2. On the other hand, lateral stability encourages vertical instability. Since the study of historic channel planform change using maps and air photographs is limited to a two-dimensional profile whereby erosion or aggradation of the channel bed cannot usually be detected, unless terraces are present (Lewin et al, 1983) it is impossible to determine whether the lateral stability of the channel is compensated by vertical instability.

One of the most important causes of river planform changes throughout the British uplands, and particularly within the Northern Pennines is changing land-use (Macklin et al, 1991). Today, land-use within the catchment is predominantly rough sheep pasture and open moorland. However, there is no evidence to suggest that land-use has changed in the Swinhope Burn catchment over the historic period. Research on the recent forest history and land-use in Weardale (Roberts et al, 1973) suggests that during

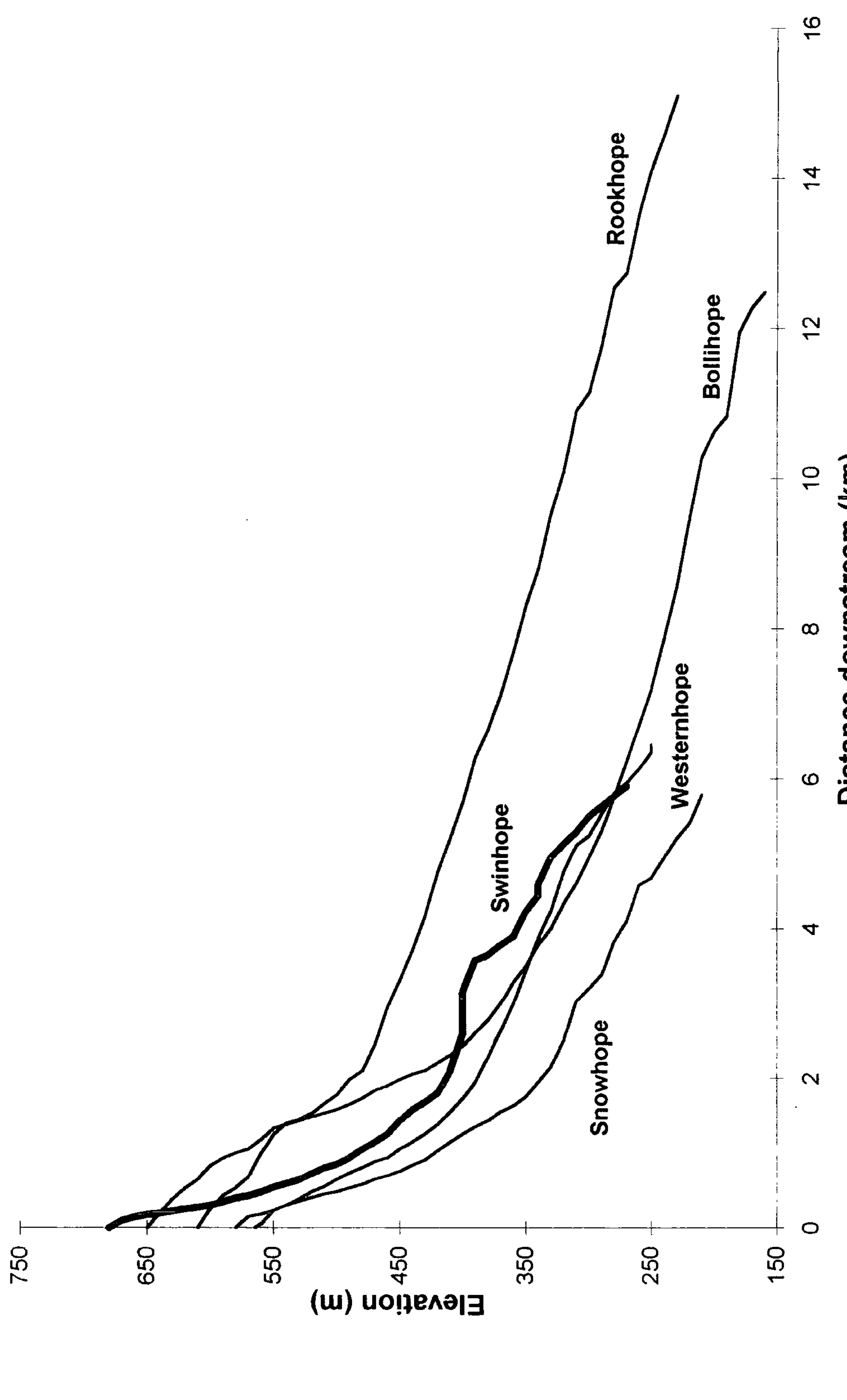


Figure 4.3 Long profile of Swinhope Burn and adjacent catchments

the Iron Age and Romano British period, 200 BC to AD 300, the woodlands were cleared. In the early 14th century settlements had exploited the upland pasture reserves of Weardale up to 270 metres and on the moorlands large intakes of land were ploughed up to 420 metres. It is possible that the lower part of Swinhope Burn was cultivated, but it is more likely that the upper reaches were reserved for hunting. By the mid 17th century very little woodland would have remained. In the early 19th century much land was inclosed, which corresponds with the Inclosure Map of 1815 (Figure 4.2). After 1816 pine and spruce woodlands were planted. In 1844, the Tithe records which accompanied the Tithe Map (A.R. Reed, 1842) state that land-use was grassland throughout the Swinhope catchment with small areas of woodland which correspond with the present day distribution of trees.

This lack of change in land-use helps to explain the remarkably stability in channel planform over the period. However, the picture is complicated by evidence that during the first half of the 19th century, small-scale metal mining took place in the uppermost reaches of the Swinhope catchment.

4.4.4. Causes of river planform change in Swinhope Burn

Although the channel planform of Swinhope Burn has been shown to be extremely stable over a period of 180 years, the reason for the dramatic change in channel pattern from meandering to essentially straight with braiding, distinguishable on the Tithe Map of 1844 (Figure 4.2), must be identified. Although it is difficult to pinpoint the actual cause of a historic change in channel pattern without the benefit of a well-established flood chronology for the basin, trace metal dating or lichenometry, it is possible, using other field evidence and local historic flood documentation to identify a probable cause. There are three hypotheses to explain the cause of such a dramatic change in channel pattern in Swinhope Burn from the Inclosure Map in 1815 to the Tithe Map in 1844:-

1. Mining activity within the catchment

2. The occurrence of a large flood event or a sequence of large flood events.

3. A combination of 1 and 2.

In order to evaluate the likely cause of river channel change each hypothesis will be examined in turn with reference to both local historic documentation and field evidence.

4.4.5. Historic Metal Mining in the Swinhope Burn Catchment

Historic documentation and field evidence

In recent centuries the most important change in land-use was the development of metal mining. The role of mining on river channel change is well documented for the streams in the Northern Pennines (Macklin, 1997). Over a period of 200 years the Beaumont Company was responsible for the substantial development of lead mining and smelting in Allendale and most of Weardale (Dunham, 1948).

Although some streams within the Northern Pennine Orefield have undergone substantial changes as a result of large-scale mining activity, for example, the River Nent near Alston (Macklin, 1986) and Black Burn, a tributary of the River South Tyne (Macklin, 1997a), it initially appeared that Swinhope Burn had been mainly unaffected by large inputs of metaliferous waste generated by mining activity.

Swinhopehead Mine (NY 877 342) is situated where the Swinhope Cross Vein crosses the headwaters of Swinhope Burn. Although the mine is currently unused, there is documentary evidence to show that lead ore was mined from the main fault (Dunham, 1948). The Beaumont Company records state that the Swinhope Vein was about three and a half foot wide with 'free spar thinly mixed with ore', with large amounts of silaceous fluorspar present. The only record of production is 326 tons of galena in 1823-46.

When compared with the output of lead ore of 127,785 tons from Boltsburn Mine, one of the largest Weardale mines (Raistrick and Jennings, 1965), mining activity at Swinhopehead mine was small. Around 1829 many mines in the Northern Pennine Orefield were abandoned due to foreign competition and reduced activity of lead-using industries, such as the building industry (Raistrick and Jennings, 1965). However, Weardale mines were particularly badly hit. Mining for lead ore in the Swinhopehead mine began before and persisted during the post-depression period whereas other mines in the area were abandoned, which may suggest that the mine operated sporadically or was a relatively small operation. A trial for fluorspar was made by G.G. Blackwell and Son in 1905, but no production is recorded. Historical documentation suggests that mining in the uppermost reaches of Swinhope Burn was relatively short lived, the mine being operational for a period of 23 years.

Likewise, there is very little field evidence to suggest that historic mining activities have influenced historic river channel change. The only visible signs of the impact of mining are the remains of a few disused buildings just downstream of the entrance to Swinhopehead Mine. There are still a few small streamside waste heaps where the lead ore was 'sorted and dressed' using hammers. Immediately downstream of the main level the channel is braided in places and there are a series of small berms running parallel to the river and a few terraces whose surfaces exhibit impaired vegetation growth. It is likely that the impaired vegetation on bar surfaces has resulted from heavy metal pollution from the Swinhopehead mine as is the case on numerous streams draining the North Pennines Orefield (Macklin, 1997). Further downstream, within the study reach, impaired vegetation growth is conspicuously absent. Sediment within the study reach does not exhibit the angularity and black iron manganese coating characteristic of streams polluted by metal mining.

Using field evidence and a well-documented knowledge of the lead mining industry in the Northern Pennines (Dunham, 1948; Raistrick and Jennings, 1965) it appears very unlikely that channel pattern change in Swinhope Burn over the historical period has been significantly influenced by metal mining in the same way as other Pennine streams.

However, the dramatic change in channel pattern from a meandering stream to one which is straight and with a braid bar in the upper reaches evident on the 1844 Tithe Map (Figure 4.2), corresponds with the period of mining activity from 1823 to 1846 suggesting a causal link. The abandoned channel (NY 897 348) is still evident on the 1991 Air Photograph (Figure 4.4) and has been preserved in the floodplain (Figure 4.5). However, it must be stressed that braided channel patterns so often associated with the onset of metal mining on streams within the Northern Pennines (Macklin and Aspinall, 1986; Macklin, 1997) and Wales (Lewin et al, 1983) are quite poorly developed on Swinhope Burn. Also, the existence of a braided channel pattern does not necessarily prove that mining activity is the cause.

Since the Tithe Map of 1844 shows a channel with much lower sinuosity than that recorded on the Inclosure Map almost 30 years before and with a braid bar in the upper reaches, there is reason to suggest that mining activities in the catchment have affected the historical development of channel planform. It can be suggested that upstream inputs of coarse mining waste were temporarily transported downstream to the study reach causing blockages within the channel. This may have caused meander cutoff through channel avulsion which resulted in a considerable reduction in sinuosity and the formation of a braid bar in the upper reaches. This would help to explain the channel pattern detailed on the Tithe Map of 1844 (Figure 4.2).

However it seems quite surprising that a mine of such small size and one which was operational for such a short period of time could have induced such a dramatic change in channel pattern. This may, however, be countered by the small size of the Swinhope Burn basin. The actual amount of mining waste introduced into the channel would largely depend upon the geology of the basin, the methods of working and coupling between the channel and the slopes in the uppermost reaches immediately downstream of the mine.

The only clear evidence to suggest a causal link between mining in the catchment and river channel change is the coincidence of the period of mining activity, 1823 to 1846, with the observed change in channel pattern around 1844. Therefore, although it is possible that mining within the catchment was solely responsible for the change in channel pattern it seems more likely that either a major flood or a series of major floods in conjunction with inputs of coarse mining waste may have been sufficient to induce such rapid channel transformation. Confirmation of the causal link between mining activity and channel pattern change can only come from the measurement of high metal concentrations in the abandoned channel.

Figure 4.4 Air photograph showing study reach at Swinhope Burn taken 7/9/91 (reproduced courtesy of Durham County Council and Aerofilms)

Figure 4.5 Abandoned channel preserved in the floodplain utilised by present day flood flows in the upper reaches of Swinhope Burn

4.4.6. Historic flood events in the Swinhope Burn catchment

Historic documentation and field evidence

Although the period of channel change corresponds with that of mining activity within the catchment, it also corresponds with a period when a succession of large floods occurred on the River Wear and its tributaries.

Historical documentation detailing flood events in remote areas of the Northern Pennines is very scarce. Although the earliest newspaper, The Durham Advertiser, began in 1814, most of the others date from the mid to late 19th century and the dates and details of flood events frequently refer only to the more populated areas of County Durham.

Since the most dramatic period of river channel change, as identified in Figure 4.2, occurred sometime from 1815 to 1844, this research concentrates on identifying recorded flood events within this timescale. The aim is to identify either one large flood or a series of floods which may have been responsible for the switching of the channel from a meandering to a straight channel with braid bar.

Using local historical publications (John Sykes, 1833; T. Fordyce, 1867; W.M. Egglestone, 1874), contemporary newspapers (The Durham Advertiser and The Durham Chronicle) and current accounts (Archer, 1992) it has been possible to identify the occurrence and details of particular flood events which likely affected the Swinhope Burn catchment. Table 4.1 identifies historic flood events which have been reported to have affected the region between 1815 and 1844. Figure 4.6 locates the main rivers and towns mentioned in the flood reports and Figure 4.7 shows details of the River Wear catchment and its tributaries.

Date of flood	Areas/rivers affected	Source
30th December, 1815	R. Tyne, R. Tees and R. Wear	Durham Advertiser (1816); Sykes (1833); Archer (1992)
2nd February, 1822	R.Wear (Stanhope, Durham), R.Tees (Yarm), R. Tyne (Newcastle)	Durham Advertiser (1822); Durham Chronicle (1822); Sykes (1833); Archer (1992)
2nd September, 1824	Swinhope Burn, St. John's Chapel, Middleton-in-Teesdale, Hexham and Berwick (Nothumberland)	Durham Advertiser (1824); Sykes (1833); Egglestone (1874)
10th & 11th October, 1824	R. Wear, R. Tyne (Newcastle), R. Gaunless (West Auckland) and R. Browney.	Durham Advertiser (1824) Sykes (1833); Archer (1992)
13th & 14th July, 1828	R. Wear, R. Tees and R. Tyne	Durham Advertiser (1828); Sykes (1833); Archer (1992)
10th January, 1837	R. Wear and R. Tyne	Durham Advertiser (1837); Fordyce (1867)
20th December, 1837	R. Wear and R. Browney	Durham Advertiser (1837); Fordyce (1867)

Table 4.1 Dates and details of historical documented floods on the River Wear, River Tyne, River Tees and their tributaries

During the period between 1822 and 1828 a series of four major floods are recorded on the River Wear. Archer (1992) identifies the three major floods (February, 1822, October, 1824 and July, 1828) as occurring in quick succession after 'an era of comparative quiescence' and documents no other major flood events until 1852. Two other floods affecting the River Wear are recorded in 1837, on the 10th January and

20th December (Table 4.1). Although the damage caused by the flood of January 1837 was described as 'comparatively trifling', it appears that the rivers were 'greatly flooded', particularly on the Wear as a result of the flood in December 1837 (Fordyce, 1867). The floods of 1837 are recorded in the Durham Advertiser and may have passed through the Swinhope Burn basin.

Documentation relating to the flood of September 2nd, 1824 (Durham Advertiser, 1824; Sykes, 1833; Egglestone, 1874) is of most importance in reconstructing the history of flood events within the study reach since Swinhope Burn is identified by the flood report as being the most severely affected area. W.M. Egglestone, in the Weardale Nick-sticks (1874) records the event as follows:-

'A terrible thunderstorm during which a great number of people and cattle were killed in the North of England. The storm was most terrific at Swinhope in Weardale, a torrent of rain having fallen which swept away Swinhope stone bridge and brought down several large pieces of peat moss. A quantity of hay was carried away out of the 'well field' at stone-heap, near St. John's Chapel and the low lands hereabouts were completely flooded'.

Likewise Sykes (1833) in his 'Historical register of remarkable events' similarly records a 'severe thunderstorm' in the vicinity of Middleton-in-Teesdale and Hexham and Berwick in Northumberland on the 2nd September, 1824. The Durham Advertiser (4th September, 1824) records that on the 2nd September, Durham City and its neighbourhood were visited with a 'storm of thunder, lightening and rain'.

Although the late summer storm of 1824 appears to have been localised around the Swinhope Burn basin it is possible that flooding was widespread throughout the region, as far north as Berwick. However, it is certain that the heavy rain resulting from a thunderstorm had produced a flood within the Swinhope basin which was of sufficient magnitude to demolish a stone bridge and erode large pieces of 'peat moss'. Although it is unclear as to which 'stone bridge' at Swinhope is being referred to as there are two, one at the head of the study reach and the other further downstream where Swinhope Burn joins the River Wear, the flood must have originated at the basin head.

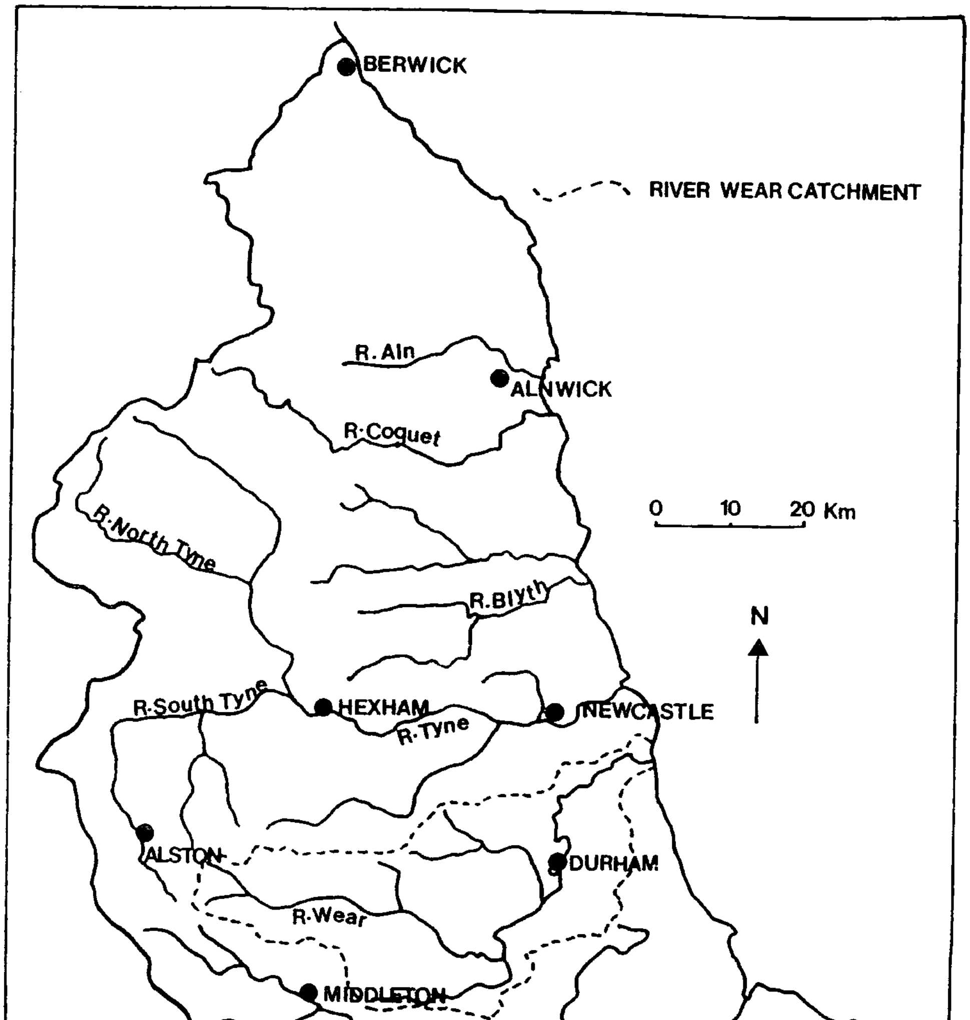


Figure 4.6 Location map of rivers and towns in north-east England (based on Archer, 1992)

The flood which occurred on 10th and 11th October, 1824, a little more than a month after the aforementioned flood, is recorded as being 'one of the severest storms of wind and rain ever remembered' and affected both the River Tyne and the River Wear (Sykes, 1833). It is recorded that 'the rain fell in torrents, and laid considerable tracts of land under water. Much damage was done on the banks of the River Wear by its rising to a tremendous height' (Sykes, 1833). It was described by the Durham Advertiser as a 'perfect hurricane' and the papers agreed that the river was at its highest at Durham since 1771 (Archer, 1992).

Apart from the two floods which occurred on the Wear during September and October, 1824, two other floods are recorded to have affected the Wear, the first in February, 1822 and the other in July 1828. The February, 1822 flood was reported to have caused the most significant problems at Stanhope in Weardale. Field walls are documented to have been demolished and fields were covered with sand and stones (Archer, 1992).

The second flood of July 1828, the magnitude of which was not equalled for a further 25 years on the Wear, was reported to have resulted from 'a heavy rain having fallen' and 'the rivers in the northern counties' were said to have 'swollen to such a degree as caused much loss and damage to the lands and crops on the low grounds and banks of the rivers' (Sykes, 1833). It is stated that 'great damage was done on the grounds contiguous to the rivers Tees and Wear' and 'the low lands adjacent to the River Wear were for a long time under water'.

The occurrence of a succession of four major floods affecting the River Wear from 1822 to 1828 is highly significant as this coincides with the channel pattern change in Swinhope Burn identifiable on the 1844 Tithe Map. Although there is evidence that one of these major flood events occurred within the Swinhope Burn catchment, the flood of September, 1824 (Egglestone, 1874) it is more than possible that the other three flood events caused a degree of flooding along the study reach. The flood of 1822 and the second major flood in 1824 both occurred during the winter months which makes it more likely that Swinhope was affected as they were probably due to regional weather systems rather than localised convective summer storms.

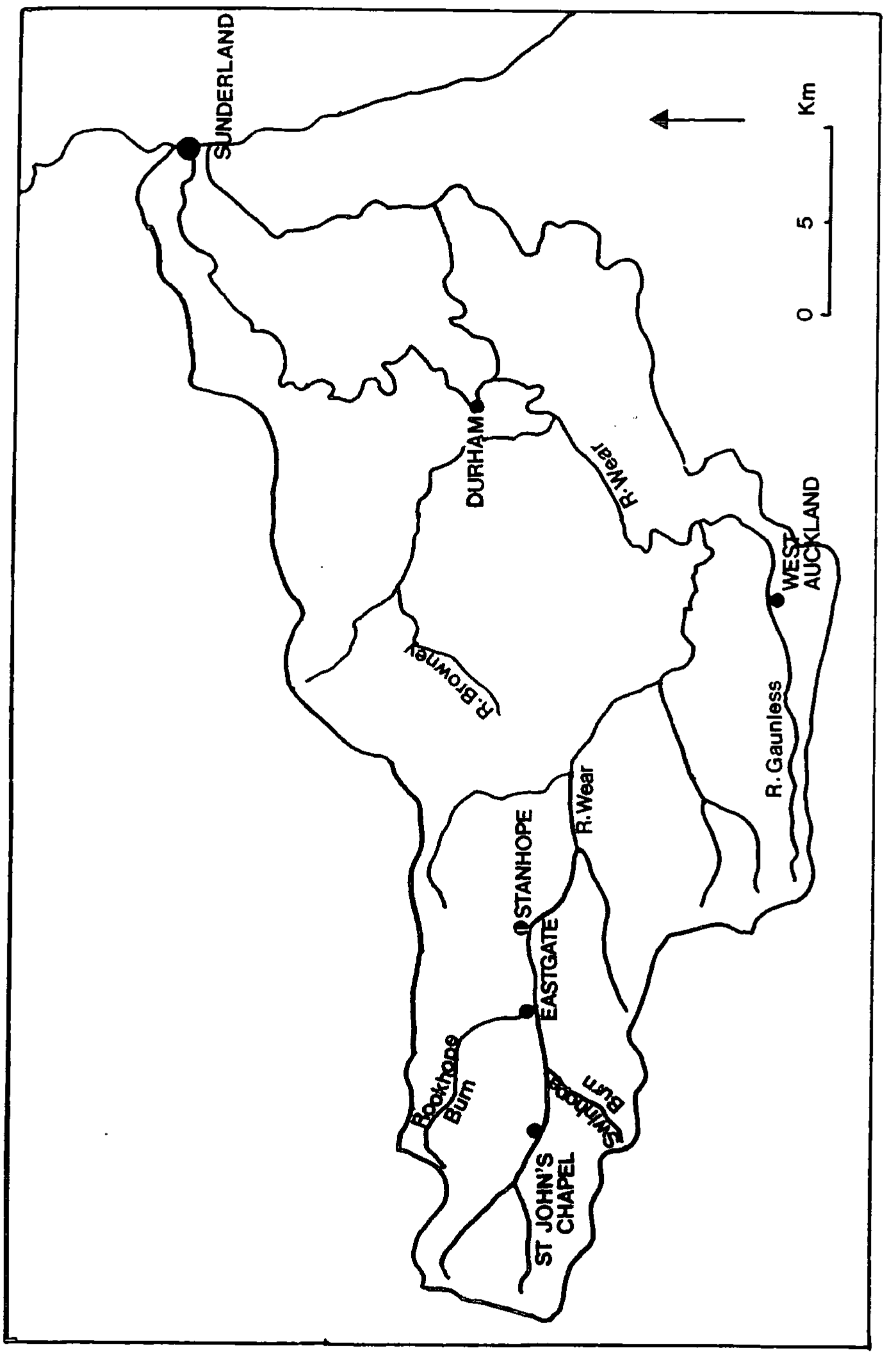


Figure 4.7 The River Wear and its catchment (based on Archer, 1992)

Although the 1822 flood appears to have been restricted to the River Wear, the October 1824 flood was widespread affecting both the River Wear and the River Tyne. Even though the 1828 flood occurred during the summer months it also appears to have been regional in nature affecting both the River Wear and the River Tees.

Field evidence of contemporary floods in the Swinhope Burn basin suggests that one high magnitude flood is not of major significance in terms of channel change. Since changes in channel form resulting from floods are small, the channel rapidly recovers to its pre-flood state. Therefore, it is likely that a number of major floods occurring in quick succession would be necessary to produce a significant and prolonged change in channel planform.

Other field evidence for the role of flood events in determining river channel change in Swinhope Burn mainly takes the form of numerous palaeochannels crossing the floodplain which is visible from the 1991 air photograph (Figure 4.4) and from a photograph taken during a recent flood event in which the palaeochannels are reopened (Figure 4.8). The palaeochannels, which are more evident in the lower part of the basin where the general morphology is more uneven, appear to be occupied by higher present day flood flows. The network of palaeochannels crossing the basin are evidence that historic floods were of sufficient magnitude to cause channel avulsion right along the reach. However, the braiding shown on the Tithe Map of 1844 (Figure 4.2) in the upper reaches of Swinhope is the clearest evidence of river channel change. The abandoned channel is still preserved within the floodplain (NY 897 348), and is visible on the 1991 air photograph (Figure 4.4) and on the ground (Figure 4.5). The thick layer of fine fluvial deposits throughout the basin bounding the study reach provides some evidence of overbank flooding over the historical period as do a number of steep cut banks and bluffs which occur along the channel. Conversely, the presence of thick organic deposits on the floodplain suggests the development of a valley-bottom bog, resulting from relative channel stability and prolonged wet conditions.

Documented flood records and field evidence strongly suggest a causal link between the occurrence of a succession of major floods on the River Wear during the 1820's and the change in channel pattern recorded on the 1844 Tithe map. It seems likely that these floods affected the Swinhope Burn basin.

Figure 4.8 Palaeochannels visible in the Swinhope Burn basin utilised by present day flood flows

The change in planform from a single thread, meandering channel to a low sinuosity channel with a braid bar in the upper reaches probably resulted from the passage of these floods through the study reach.

The likely effect of these floods would have been to straighten the channel through avulsion as the channel became blocked due to the increased sediment supply (Macklin, 1986). For example during the flood of September, 1824 'large pieces of peat moss' were reported to have been transported downstream at Swinhope Burn (Egglestone, 1874). Likewise, the formation of a braid bar at the top of the study reach may have resulted from the increased supply of coarse sediment.

It is likely that the succession of four major floods occurring over a period of six years has been a critical factor in governing channel planform change along the study reach, evident on the 1844 Tithe map (Figure 4.2). However, over the historical period it is likely that the Swinhope Burn basin has been flooded on many occasions, given the documented flood history on the River Wear and evidence of palaeochannels traversing the floodplain. It is therefore very interesting that the channel planform has remained so stable over the last 180 years. This suggests that the dramatic change in channel planform recorded in 1844 may have resulted from a combination of factors, not reproduced in the last 180 years.

The reasons behind the dramatic change in channel pattern shown on Figure 4.2 may lie in coincidence between the period of mining activity, from 1823 to 1846 and the period during the 1820's which saw a succession of major floods on the River Wear which both correspond with the change in channel planform on the 1844 Tithe Map.

Since inputs of coarse sediment to the channel through mining activities and during high magnitude flood events can produce similar patterns of river channel change (Macklin, 1986), in the form of braiding and straightening of the channel, it is very difficult to determine the effects of each on historic channel planform change. In order to determine the most probable cause of channel planform change it is useful to compare Swinhope Burn with other sites in the British uplands which have undergone historical channel change as a result of either a high magnitude flood event or through historic metal mining.

4.4.7. Examples of historic river channel change from recent literature

River channel change in response to mining activities

The role of historic metal mining in channel planform development has been well documented for the Northern Pennines (Macklin, 1986; Macklin, 1997) and Wales (Lewin et al, 1977, 1983). Certain similarities can be identified between channels within catchments where historic metal mining has influenced channel planform change, and the changes in channel pattern identifiable on the Tithe Map of 1844 in Swinhope Burn (Figure 4.2).

Changes in channel pattern from one which is single thread and meandering to one which is braided is generally associated with the downstream movement of a sediment wave generated by historic metal mining (Macklin, 1997a). Abandoned, braided channel systems found above present stream levels provide evidence of channel adjustment to high inputs of coarse sediment associated with metal mining (Macklin, 1986). Braiding has been shown to have occurred initially in the upper reaches of the River Nent (Macklin, 1986) and was subsequently followed by the intermittent downstream movement of coarse gravel. The braid bar found at the head of the study reach at Swinhope Burn on the 1844 map (Figure 4.1 and Figure 4.5) may provide similar evidence of an increased supply of coarse sediment from upstream, generated by mining activities. Likewise, Lewin et al (1983) describe the transformation of the Afon Ystwyth in Wales from a single-thread meandering channel to one with multiple channels in response to prolonged lead mining in the catchment. However, this was accompanied by significant aggradation of the valley floor with incision of the channel during the post-flood period leading to the formation of a series of terraces. Historical maps record an increase in sinuosity due to the multiplicity of channels whereas at Swinhope Burn a decrease in sinuosity is identified.

The Tithe map of 1844 shows evidence of a change in channel pattern with a dramatic decrease in sinuosity and the presence of a braid bar in the upper reaches suggesting the cause may be historic metal mining. However, examples of other streams in the British uplands which have undergone transformation through historic mining suggest that it is unlikely that inputs of coarse sediment through metal mining is solely responsible for channel planform change in Swinhope Burn. The examples describe the development of multiple channels and large mid-channel bars, vertical aggradation within channels, large volumes of bare gravel covering valley floors and the impairment of vegetation growth on berms and bars which are all diagnostic features of a prolonged period of metal mining within a catchment.

However, it has been suggested that metal mining in contrasting fluvial environments will lead to contrasting patterns of channel modification (Lewin et al, 1977). This will largely depend on the nature of mining activities and the local fluvial process system. In some cases, as on the River Ystwyth, Wales (Lewin et al, 1977) inputs of toxic fine sediment are of more significance than inputs of voluminous waste in terms of river channel change.

River channel change in response to flood events

The alternative hypothesis to explain the dramatic change in channel planform of Swinhope Burn from meandering to relatively straight with a braid bar (Figure 4.2) is the occurrence of a major flood or a series of floods (McEwen, 1994). Flooding has been frequently identified as the cause of historic river channel change throughout the British uplands (Anderson and Calver, 1980; Ferguson and Werritty, 1980; Milne, 1982a; McEwen, 1989, 1994; Macklin, 1994; Macklin, 1997c). The immediate response of channels to a flood of high magnitude has been identified as braiding (Anderson and Calver; 1980, Werritty and Ferguson, 1980; Werritty, 1982; McEwen, 1989a). This is accompanied by a decrease in channel sinuosity either through meander loops becoming blocked with coarse debris leading to channel avulsion (Anderson and Calver, 1980; Werritty and Ferguson, 1980; Werritty, 1982) or through the development of an expanded channel network which ultimately leads to the development of a sweeping flood by-pass channel (McEwen, 1994). McEwen (1994) identifies major channel switching triggered by a major episodic flood or a series of floods on the River Coe. However the critical factor in the occurrence of channel change is an abundant supply of coarse sediment from valleyside slopes. Similarly, Anderson and Calver (1980) observed a reduction in the sinuosity of Hoaroak Water on Exmoor following the 1952 flood through existing channels becoming blocked with debris and overtopped. Once again an abundant supply of sediment was the key factor in causing river channel change, but unlike Swinhope where the channel had regained its meandering pattern within 12 years, the constant supply of sediment has meant that 25 years later the planform of Hoaroak Water essentially dates from the early post-flood period.

The hypothesis that channel planform changes evident on the 1844 Tithe map are due to the occurrence of a major flood or a series of floods is supported by research on Harthope Burn, Northumberland (Milne, 1982a). Historical maps showed that the present channel was straighter and more braided than it had been 100 years earlier. The cause of reduced channel sinuosity was identified as occurring through blockage of the channel during the transport of coarse material during floods which had led to low-sinuosity cutoffs. Similarly, braiding of the channel had resulted from inputs of coarse sediment from upstream during flooding of the basin.

However, the role of flood events in determining river channel change can vary from one catchment to another. Anderson and Calver (1980) identify large drainage area and a wide valley floor as critical criteria if a large runoff event is to cause channel planform change. If the channel is confined by terraces or valley-side slopes, has a low-gradient perhaps with a local base level and has restricted availability of coarse sediment, lateral migration of the channel is limited and it is likely to remain stable (McEwen, 1989a). In cases such as these the immediate effects of a high magnitude flood may be minor channel avulsions. If there are large readily available sources of coarse sediment a similar high magnitude flood in another catchment may have a lasting geomorphic impact (Werritty and Ferguson, 1980).

Recent literature on historic channel planform response to high magnitude flood events strongly suggests that the change in channel pattern in Swinhope Burn is the result of either a major flood or a series of floods. The decrease in channel sinuosity and braiding in the upper reaches identifiable on the 1844 Tithe Map (Figure 4.2) are very similar channel changes to those described as resulting from flood events on many other streams in the British uplands. Equally, field evidence of historic floods in the form of palaeochannels crossing the basin and historic documentation of flood events within Swinhope Burn basin and on the River Wear during the 1820's also suggest that it is highly likely that flood events are responsible for river channel change. It is possible that since there was a sequence of large flood events on the River Wear during the 1820's that channel planform had not recovered its meandering pattern by 1844 when the river was surveyed for the Tithe Map.

However, the critical factor in the maintenance of a low sinuosity channel would have been the continued supply of coarse sediment. It may be that high inputs of coarse sediment introduced to the channel during these large documented flood events in the 1820's combined with inputs of coarse mining waste from 1823 to 1846 from the upstream Swinhopehead Mine. Mining began in the catchment in 1823 which coincided with major floods in 1822, 1824 and 1828. It is suggested that the channel pattern was initially transformed either as a result of one or a series of flood events during the early 1820's and that small inputs of coarse mining waste helped to maintain the low sinuosity channel. This would explain the decrease in sinuosity of the study reach and the presence of a braid bar in the upper reaches. By 1846, mining had ceased and by 1856 at the latest, the channel had regained its meandering pattern evident on the 1815 Inclosure Map.

There is evidence from other studies of historic river channel change in the Northern Pennines to suggest that inputs of coarse sediment generated by major flood events and inputs of coarse mining waste can combine to cause both a reduction in channel sinuosity and an increase in braiding similar to that shown on the 1844 Tithe Map of Swinhope Burn. Examples are the River Nent, Blagill (Macklin, 1986), The Islands, River South Tyne (Macklin, 1997c) and River West Allen (Macklin and Aspinall, 1986). Using historical maps, Macklin (1986) identified that pre-1861, the River Nent in the Northern Pennines was a single thread meandering channel. During the 1840's there was a series of major floods on the River Nent which brought about an initial transformation to a braided river with multiple active channels. Mining was more intensive and lasted for a longer period on the River Nent, and the channel did not return to its pre-1861 meandering pattern until the 1940's. However the evidence suggests that metal mining initially provided a ready source of coarse sediment making the channel inherently unstable. Initially inputs of coarse sediment generated by large floods combined with the mining waste to bring about the initial channel transformation, but this pattern was subsequently perpetuated because of the continuing supply of coarse mining waste.

A general increase in flood frequency during the latter part of the nineteenth century in Northern England combined with an increased sediment supply in many streams in the Northern Pennine Orefield due to gross mining pollution has been identified as being responsible for river metamorphosis throughout the region (Macklin, 1997c).

However, it must be stressed that it is often very difficult to distinguish between the effects of inputs of coarse mining waste and inputs of coarse sediment generated through major flood events on river channel change. Lewin et al (1983), noted the change from a single-thread meandering channel to a system of multiple channels and aggradation on the valley floor of the River Ystwyth in Wales. Mining related gravel splays found on the valley bottom of the River Ystwyth is likened to the flood-induced cobble sheets described by Ferguson and Werritty (1983) on the River Feshie. Lewin et al (1983) concluded that channel planform had responded to change in sediment supply. Earlier work on the River Ystwyth had suggested that there was little evidence that large quantities of coarse sediment had been fed locally into the river during the mining period although braiding is evident on historical maps dating from this period. They identified the braided channels as short-term bedform changes which were dissimilar to the multiplicity of low stage channels observed in other braided rivers affected by large inputs of coarse mining waste. On the River Ystwyth they concluded that high magnitude flood events were more important in river channel change through the reactivating of abandoned channels than mining activities had been. However, they confirm the difficulties associated with distinguishing between mining-induced braiding and flood-induced braiding on historical maps.

4.5 Recent Flood History of the Swinhope Burn catchment

As the study reach at Swinhope Burn is located in a remote upland area of the northern Pennines, very little is known of its recent flood history. Information detailing the effects of flood events outside major towns and cities was limited until the establishment of local newspapers in the second half of the nineteenth century (Archer, 1992). Although the first gauging station in the Wear catchment (Figure 4.7) was set up in 1954 on the River Browney at Burn Hall, like many catchments of similar size the streamflow of Swinhope Burn has remained ungauged. However flow records are available for the River Wear at Stanhope in Weardale (16 km north-east of the study reach at an altitude of 202 m, NY 984 391) commencing in 1958 and a right bank tributary of the River Wear, Rookhope Burn, Eastgate, Weardale (6.5 km north-east of the study reach at an altitude of 241m, NY 952 390) commencing in 1969. Although flow data from gauged adjacent catchments have successfully been used to provide an insight into the size of flows generated within a study reach in order to evaluate their efficiency in potential channel change events (Milne, 1982a), in this case the data available may be slightly less representative of likely flows on Swinhope Burn. This is due to the large size of the catchment area of the River Wear (171.9 km^2) compared with Swinhope Burn (10.5 km^2) and the fact that Rookhope Burn at Eastgate is a left bank tributary of the River Wear and flow records have been discontinued since 1980.

Although there are raingauges in adjacent catchments to Swinhope Burn, the nearest being at High Greenwell (5 km north-west of the study reach at an altitude of 390m, NY 860 383) owing to the localised nature of storms and associated flood events within this area it would be unwise to use this data to estimate flood frequency within the study reach.

However, using the flow records for Stanhope and documented reports of large flood events on the River Wear, generalisations can tentatively be made regarding the likely occurrence of flooding in Swinhope Burn basin since 1958. The evidence suggests that over the past three decades there have been a number of large floods on the River Wear, which are likely to have caused flooding within Swinhope Burn basin. Table 4.2 shows mean daily flows and peaks over a threshold recorded at the gauging site at Stanhope and Rookhope in Weardale, which coincide with documented large flood events. Flow records for Rookhope are available from 1960 to 1980 only.

Table 4.2 identifies six major documented floods which have occurred within the River Wear catchment over a period of 28 years. Four out of six of the floods identified as having both a high mean daily flow and peak over threshold value recorded on the River Wear at Stanhope also rank highly in the peak over threshold records for Rookhope Burn. The largest flood event on record, which occurred on the 31st January, 1995 is not recorded at Rookhope Burn since the gauging station at Eastgate was discontinued after 1980.

Since high mean daily flows and high peak over threshold values have been recorded at both Stanhope and Eastgate and since large floods within the Wear catchment have been documented it seems likely that flooding would have occurred within the Swinhope Burn basin at these dates. Certainly, the largest flood on record which occurred on 31st January, 1995 passed through Swinhope Burn basin (pers. comm. farmer), as did the floods of February, 1997 and January, 1998, which occurred during the field monitoring period.

Peak flow for Stanhope (m^3s^{-1}) ***(Rookhope Burn data in italics)***	**Mean Daily Flow for Stanhope (m^3s^{-1})**	**Flood Date**	**Documented flood record**	**Comments**
287.97	155.06	31- 1- 1995	Pers. Comm. farmer. Swinhope Burn.	Flood recorded at Swinhope Burn. Ranked 1st on peak flow record
160.39 119.5	49.09 51.39	19 - 2 - 1997 & 20 - 2 - 1997	Personal observations - morphological and grain-size survey carried out to identify channel change.	Occurred during monitoring of field site on Swinhope Burn. 19-2-97 (ranked 4th highest on peak flow record). 20-2-97 (ranked 26th highest on peak flow record)
149.66 *(33.57 - ranked 4th on record)*	84.27	5 - 11- 1967	Archer (1992)	Ranked 9th on peak flow record. More than 75mm rain recorded over the greater part of the Wear catchment.
129.85 *(38.64 - ranked 1st on record)*	63.21	11- 09 - 1976	Archer (1992)	Ranked 18th on peak flow record
111.08 *(26.19 - ranked 11th on record)*	43.89	16 - 10 - 1967	Archer (1992)	Ranked 33rd on peak flow record
105.65	76.64	3 - 1- 1982	Archer (1992)	Ranked 36th on peak flow record
92.84 *(25.64 - ranked 12th on record)*	50.25	15 - 10 - 1976	Archer (1992)	
91.35	32.54	8 - 1 - 1998	Personal observations - sediment tracing experiment.	Occurred during monitoring field site on Swinhope Burn. Ranked 58th highest on peak flow record.

Table 4.2 Documented floods in River Wear catchment using mean daily flows and peak over threshold values for the River Wear at Stanhope (1958-1998) and Rookhope Burn (1960-1980) at Eastgate. (Data supplied by the Environment Agency, Newcastle).

However, although four large floods are documented for the period 1967-1976, the map showing the channel planform of Swinhope Burn surveyed between 1977 and 1982 (Figure 4.2) shows very little change when compared with the channel planform surveyed in 1953. Likewise there is very little change in channel planform of Swinhope Burn from the 1991 Air Photograph (Figure 4.4) to the current (1998) Ordnance Survey Raster Data, which again suggests that even though the largest flood on record, which occurred in January, 1995, affected the Swinhope Burn basin, there was very little actual channel planform change.

Assuming Swinhope Burn basin was affected by flooding at these dates, it can be suggested that either the channel is highly stable and the channel planform was unaffected by these flood events or it may be that the channel rapidly recovered to its pre-flood state. Alternatively, channel response to flooding may have been through changes in its vertical cross-sectional form, rather than through lateral adjustment of the channel.

River response to flood events in the Swinhope Burn basin has been monitored since October, 1995. Nine months after the largest flood on record in the Wear catchment had passed through the Swinhope Burn basin, evidence of the flood was still widespread. This took the form of extensive bank undercutting, failure of slopes bounding the channel and deposition of large cobbles on the outside of meander bends. A large fan of fine back water deposits in the upper part of the study reach (Figure 4.9) was evidence of the magnitude of the flood as water was ponded up behind a stone wall which subsequently collapsed along with fences which had formed the field boundaries (Figure 4.10, personal communication, farmer). However, in the past three years there have been at least two other sizable flood events (Table 4.2), although not of the magnitude of the January, 1995 flood, which have reworked these flood deposits. However, very little channel change has resulted. Analysis of contemporary river channel change in Swinhope Burn over the period of two years (Chapter 6) will confirm that, for floods of this size, the channel is highly stable and that although flood events pass through the basin, their role in determining river channel change is minimal.

Figure 4.9 Large fan of fine backwater deposits immediately above the study reach at Swinhope Burn resulting from the January, 1995 flood

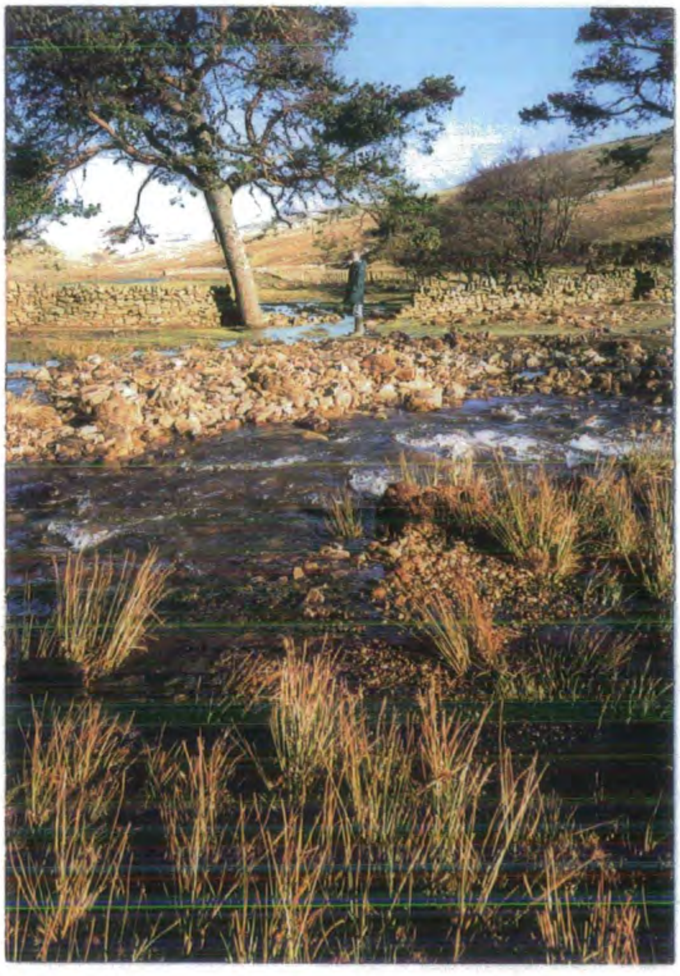


Figure 4.10 Structural damage caused by January, 1995 flood at Swinhope Burn

Therefore, in line with historical evidence of planform change in Swinhope Burn, dating back to the Inclosure Map of 1815, a brief analysis of the flood history of Swinhope Burn since 1958 has confirmed that the channel is indeed stable with the current channel planform having persisted for at least the past 140 years.

4.6 Summary

Although upland gravel-bed streams have often been viewed as being highly dynamic and frequently undergoing considerable changes in channel planform, evidence of historic channel planform development from historic maps and air photographs suggests that Swinhope Burn is an exception. Over a period of 180 years, the river planform at Swinhope Burn has remained remarkably stable, with only one apparent major change which occurred sometime between 1815 and 1844. Likewise, an examination of the recent flood history of Swinhope Burn has identified a number of high magnitude events which have passed through the study reach, but which have caused very little change in channel planform. The reason for the inherent stability of channel planform relates to its unique geomorphic setting. The study reach is located at the point at which Swinhope Burn has a distinct step in its long-profile. The presence of a local base-level, the Greenly Hills moraine, and associated reduction in stream channel slope has created a sedimentation zone, which has led to a significant reduction in stream power and downstream grain-size. Highly cohesive banks and limited coupling between channel and slopes have reduced the supply of coarse sediment to the stream which has inhibited changes in channel form. Unlike many other catchments in the British uplands, where significant changes in land-use have occurred with resulting changes in channel planform, the Swinhope Burn catchment has remained under pasture or open moorland over the historical period.

Although the evidence from historical maps demonstrates that the channel planform of Swinhope Burn has essentially remained stable over the past 180 years, in the form of a single-thread meandering channel, a dramatic change in channel pattern to a much less sinuous channel with a braid bar in the upper reaches is distinguishable on Tithe Map of 1844. It has been suggested that this change in channel pattern, which probably occurred at some point during the 1820's, initially resulted from an increase in coarse sediment supply generated by either one large flood or a series of major floods on the River Wear, one of which certainly affected the Swinhope Basin in September, 1824. However small-scale lead mining production recorded for Swinhopehead mine coincides with the period during which the channel underwent this transformation. It is possible that additional inputs of coarse mining waste to the stream combined with flood-generated inputs of coarse sediment effectively perpetuated channel instability. Since there are no other historic maps available for the period 1815 to 1856, apart from the 1844 Tithe map, it is impossible to determine the length of time the channel was transformed from its stable, meandering state to one which was relatively straight with a braid bar in the upper reaches. Therefore from historical map evidence and flood and mining documentation it can be concluded that a combination of increased inputs of coarse sediment generated by high magnitude flood events in the 1820's and upstream mining activities during the period 1823 to 1846 probably led to the observed changes in channel pattern evident on the 1844 Tithe map.

Evidence of recent floods in Swinhope Burn show that over the past 40 years it is very likely that flood events of high magnitude have passed through the Swinhope Burn basin. During the period of field monitoring, two large flood events have occurred, one of which ranked the fourth highest on the peak over threshold record from the River Wear at Stanhope. Very little change in channel planform was recorded following this flood. Likewise, the flood of January, 1995 was recorded as having severely affected the study reach at Swinhope Burn (personal communication, farmer), with large areas of the basin being submerged with fences and walls being demolished. Similarly, there was very little lasting channel change. This suggests that given the geomorphic setting of the study reach, the occurrence of a single major flood is unlikely to result in lasting channel change. It is therefore possible that an abundant and prolonged supply of coarse sediment, generated either by a series of major floods over the period of a few years or through intermittent mining activity, is the critical factor in determining channel instability and ultimately channel change.

Finally, examples of other upland gravel-bed streams which have undergone historic changes in channel planform as a result of flood-generated inputs of coarse sediment combined with inputs of coarse mining waste (Macklin, 1986; Macklin, 1997c) support the hypothesis that the temporary change in the channel planform of Swinhope Burn from a meandering pattern to one of low sinuosity may be due to the combination of these two factors.

CHAPTER 5

STRUCTURE OF THE STREAM CHANNEL

5.1 Introduction

The channel form of upland gravel-bed streams is adjusted to accommodate bankfull discharge and sediment load but is constrained by channel gradient, bank cohesion, sinuosity and bed topography. For instance, downstream variations in the width:depth ratio largely result from variations in bank cohesion, where bank material consisting of gravel and cobbles in a fine silt matrix increases width:depth ratio and highly cohesive silt clay banks produce the opposite effect (Milne, 1983a).

Equally, in upland environments, spatial variations in floodplain sediments can cause irregularities in channel planform, with a meandering pattern developing in response to pockets of highly cohesive material and a relatively straight channel developing in coarser non-cohesive material (Milne, 1983a). The pool-riffle sequence, the characteristic bedform of upland gravel-bed streams, can produce spatial variations in surface sediment sorting and the location and amount of channel erosion and deposition during floods. This has been largely explained in terms of the velocity reversal hypothesis (Keller, 1971; Carling, 1991; Clifford, 1993). Once a riffle has formed it largely governs changes in channel cross-section since mid-channel accumulations of coarse material can deflect flow towards the banks increasing the width:depth ratio (Milne, 1982c). However, channel slope exerts a major control over both the structure of upland streams and downstream fining of bed material (Knighton, 1980), and has been identified as partly an independent variable with respect to contemporary channel processes particularly where an imposed local base level is inherited from the deglacial conditions (Ferguson and Ashworth, 1991).

The purpose of this chapter is to demonstrate how the channel and mean grain-size of bed material at Swinhope Burn are adjusted to downstream variations in channel slope and channel planform, at a reach scale, and bank cohesion and bed topography at a more localised scale.

5.2 Scope of chapter

Initially, this chapter provides a background for recent research into the structure of upland gravel-bed streams. Downstream variations in cross-sectional form and mean grain-size are identified. The following sections examine individually the effect of variations in bed topography, bank cohesion, channel sinuosity and channel gradient on the structure of the channel at Swinhope Burn.

5.3 Background

The riffle-pool sequence is the characteristic reach-scale bedform of upland gravel-bed channels of low to moderate slope. Riffles are topographic highs which may extend partially or completely across the channel and are marked at low flow by steep water surface slopes and rapid divergent flow. Pools are the intervening lows with more gentle water surface slopes and less rapid, convergent flow patterns. In gradually curving or straight reaches, pools are not specifically related to bends. However, deep pools are especially associated with meander bends and are formed by scour beneath the concave bank and in particular at the bend apex (Ferguson, 1981).

The pool-riffle sequence influences channel processes which has a bearing on both channel shape and channel pattern. A meandering channel pattern is developed with riffles at inflection points and pools at bend apices where bank erosion is concentrated. In upland gravel-bed streams, the meander wavelength is 10 to 14 times bankfull width whereas pool and riffle spacing is five to seven times bankfull width (Milne, 1979) or up to 10 times channel width (Ferguson, 1981). However, the mean spacing of pools and riffles is fairly constant when expressed as a multiple of mean channel width. The pool riffle sequence is also associated with spatial patterns in bed material size with riffle sediments being coarser and better sorted than adjacent pool sediments (Carling, 1991). Riffle-pool beds tend to be relatively stable with riffle positions tending to remain fixed through a process of particle replacement whereby the largest fraction of the bedload moves from riffle to riffle during high flow events (Knighton, 1998). However, during extreme flows a pool-riffle bed may be destroyed leading to a change in channel pattern from meandering to braided (Harvey, 1986).

Local variations in cross-sectional form are controlled by pool-riffle and meander development in the reach. Riffles in straight reaches tend to be wider and shallower at all stages of flow than pools. The fact that riffles are wider and shallower than pools within straight reaches is initially related to lack of bank cohesion at riffles. However, once a riffle has formed the increase in bed height or the presence of central bars within the channel tends to deflect flow towards one or both banks leading to bank undercutting and increased channel widening (Milne, 1982c; Knighton, 1998). Meander bend apices are usually the sites of deep pools. The location of the pool helps to focus erosion at the bend apex which perpetuates the line of development (Milne, 1982c). At tightly curved and deep pool sites, unidirectional bank erosion around actively migrating bends increases bend amplitude and can produce wider bedforms (Milne, 1982c). However, where pools and riffles impinge on valley-side bluffs the greatest width:depth ratios are observed since scouring of bed and banks is inhibited by the coarse nature of the sediment supply (Milne, 1982c).

Bed topography, bed material size and cross-sectional form are highly interrelated (Milne, 1982b; Knighton, 1998). Milne (1982b) identifies mean grain-size as the most useful criterion for distinguishing between pool and riffle sequences. Riffles are generally associated with coarser surface sediments and pools with finer sediments. Along Kingledoors Burn, a small coarse-bedload upland stream channel, riffles and pools have been found to possess relatively similar size distributions (Milne, 1982b). Very deep pools, and particularly those which are tightly curved, often situated at meander bends, tend to contain very fine sediments which results from greater infilling during low flow periods (Milne, 1982b). The distinction between bed material in pools and riffles has been made in terms of better sorting of particles on riffles than in pools (Carling, 1991). Riffles and pools can also be differentiated in terms of sediment structure, with riffles having tight imbricated structures and pools having a loose structure. This contrast has been identified as providing an explanation for the maintenance of elevation contrasts between juxtaposed pools and riffles (Clifford, 1993).

The composition of the banks in upland streams is an important factor determining cross-sectional form of the channel. Where banks are non-cohesive, consisting often of gravel and cobbles in a silty matrix, a wide shallow and often straight channel will develop. Extensive pockets of coarse floodplain sediments often coincide with long riffles (Milne, 1982c). Conversely, where the banks are highly cohesive, being predominantly silt-clay a narrow, deep and often highly sinuous channel will develop. Particularly deep pools will form on meander bend apices which become persistent features through erosion of the concave bank and deposition on adjacent point bars (Milne, 1982c, 1983a). The cohesive sediment ensures that the concave banks maintain a steep angle while retreating. Milne (1982b) suggested that the scouring of the concave banks and transverse bed material transport is responsible for bar and pool locations being linked at bend sites in one single sedimentary unit. The presence of channel bars, slumped and cut banks and toe deposits at the base of banks will also influence the width:depth ratio and the mean grain-size of the adjacent and downstream bed material. The presence of a slumped bank or substantial channel deposits at the base of banks also locally reduce or prevent bank erosion and redefine local flow patterns.

Equally, it is generally accepted that channel sinuosity is a major control over cross-sectional form (Milne, 1983a). A meandering channel pattern can increase resistance and reduces channel gradient. Although channel pattern is primarily controlled by slope, discharge and sediment supply, the actual ability of a stream to shift laterally depends on the resistance of the banks (Knighton, 1998). For example, if the bed material is very coarse, as in the upland gravel-bed streams studied by Milne (1982b; 1983a) a relatively low sinuosity channel will result.

A reduction in bed material size with distance downstream is a characteristic common to many alluvial streams and influences the downstream behaviour of hydraulic and morphological variables. Downstream variations in mean grain-size and in particular 'downstream fining' have been demonstrated for a number of gravel-bed streams in the British uplands, for example, the River Feshie (Ashworth and Ferguson, 1989), and the Allt Dubhaig (Ferguson et al, 1996). Knighton (1980) identifies downstream fining in four English streams in the piedmont zone whereas Werritty (1992) assesses the roles of lithology and abrasion on downstream fining in an upland gravel-bed river in southern Poland. Although stream gradient depends upon the balance between grain-size and discharge, which are highly dependent upon geology and catchment area, it can alter significantly downstream and this has consequences for both stream power and channel pattern (Ferguson, 1981). A decline in slope has long been associated with downstream fining of bed material in upland gravel-bed rivers (Milne, 1979; Knighton, 1980; Ashworth and Ferguson, 1989). For example, Ferguson and Ashworth (1991) and Ferguson et al (1996) identified downstream fining of bed material in relation to declining slope on the Allt Dubhaig in the Scottish Highlands, and Knighton (1975) in the Pennine hills and Cheshire plain on the rivers Bollin-Dean. The correlation of slope and mean grain size is highly significant. However, the input of sediment from tributaries and the effect of basin geology complicates this relationship.

Knighton (1975, 1980) found that in the upper reaches of the River Bollin in the Pennine uplands, there was an initially rapid decrease in mean grain-size, but as gradient became less steep, the rate of change became much smaller. Particle size changes systematically downstream with channel gradient and in headwater zones this is particularly the case since the input of coarse sediment from tributaries is likely to be relatively small. Knighton showed that channel slope decreases very rapidly where bed material was in the pebble size range but remained almost constant for mean grain diameters of 4mm or less. Consequently the relationship between slope and mean grain-size is dependent upon the size of material initially supplied to the stream and the distance over which the stream interacts with the material. Knighton (1975) argues that for a given bed material size and discharge at a point along a stream channel slope will be adjusted to provide the shear stress necessary for bed material transport at high flows.

The effects of a local base level on channel changes within the Allt Dubhaig (Ferguson and Ashworth, 1991; Ferguson et al, 1996) have been identified as rapid downstream fining of bed material and change in channel pattern from near-braided to sinuous. The downstream changes described are thought to have resulted from profile concavity inherited from de-glacial conditions. Grain-size decreased approximately exponentially with distance along the Allt Dubhaig (Ferguson et al, 1996). Ferguson and Ashworth (1991) argue that local valley slope is partly an independent variable with respect to modern channel processes. Changes in channel gradient along the study reach induce differences in channel morphology. Ferguson and Ashworth (1991) found that as channel gradient became more gentle, meander bends became increasingly stable. They conclude that slope-induced channel change is widespread in gravel-bed streams.

5.4 Downstream variations in cross-sectional form and mean grain-size in an upland stream

In order to examine downstream variations in cross-sectional form of the channel, 101 cross-sections were initially surveyed (Figure 5.1). A grain-size survey identified downstream variations in the existing composition of the bed material. The horizontal axis (distance downstream) in the scatter diagrams presented in this and the following chapter are labelled with site numbers rather than distance downstream in metres. However, since each site is spaced 15 metres apart evenly along the channel, a direct conversion can be made between site number and metres downstream. This allows the findings to be expressed as a rate of change downstream and facilitates direct comparison with other studies.

5.4.1 Downstream variations in width:depth ratio

A downstream decrease in the width:depth ratio of Swinhope Burn is demonstrated in Figure 5.2. The study reach is wider and shallower in the upper reaches and narrower and deeper in the lower reaches. In the upper reaches, bank material is coarser, consisting of gravel and cobbles in a fine silt matrix which encourages lateral adjustment of the channel. In the lower reaches, the banks are highly cohesive silt clay, which had led to the development of a deep, narrow channel.

However, there is considerable local variation in the width:depth ratio along the channel which is in response to local variations in bed topography, composition of bank material and the presence of bars and streamside scars. The effect of downstream variations in channel sinuosity on width:depth ratio is illustrated by the coincidence of peaks in width:depth ratio with meander bends, for example at cross-sections 26, 47, and 60. Peaks in width:depth ratio are smaller in the lower reaches since the depth of the pools is greater than in the upper reaches, owing to the cohesive nature of the bank material, and this offsets the increase in channel width at meander bends. The general assumption that width:depth ratio should increase downstream with discharge is complicated by local factors which help to explain the downstream decrease in width:depth ratio.

Figure 5.1 **Air photograph of Swinhope Burn showing selected cross-section locations (Courtesy of Aerofilms and Durham County Council). Length of reach is approximately 1.4 km.**

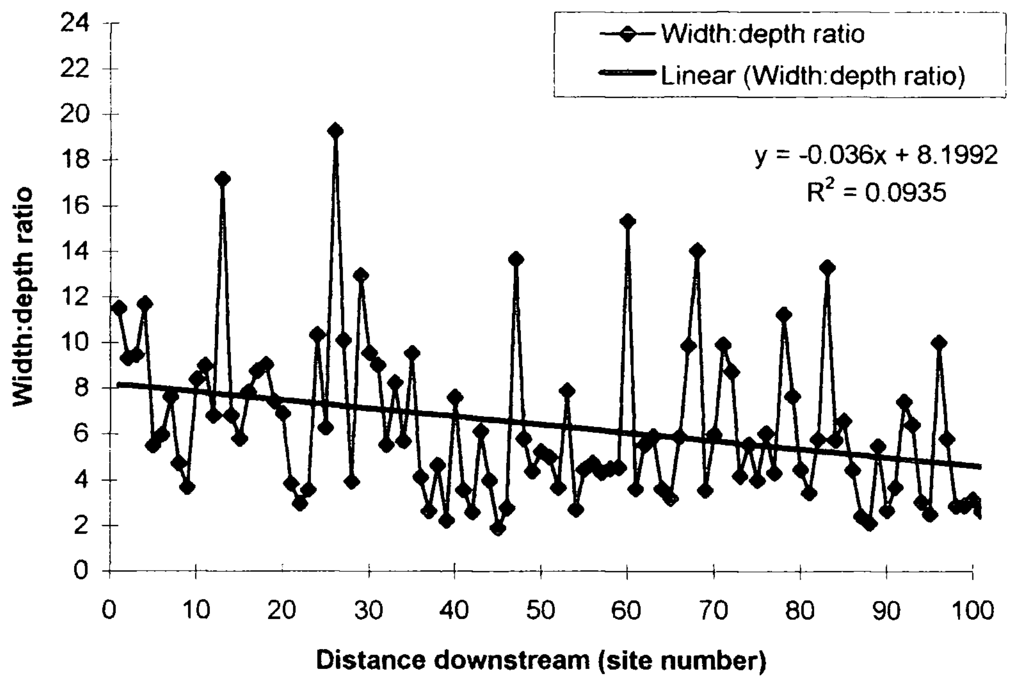


Figure 5.2 Downstream changes in width:depth ratio

5.4.2 Downstream variations in mean grain-size distribution

Figure 5.3 shows downstream changes in the mean grain-size distribution of the bed material of Swinhope Burn. Downstream fining of mean grain-size is shown most clearly from the sites at which 100 particles were sampled. Scatter in the data results partly from inputs of sediment from streamside scars and from erosion of pockets of coarse floodplain sediments. This pattern of downstream fining has been widely demonstrated for many upland gravel-bed streams (Ferguson et al, 1996).

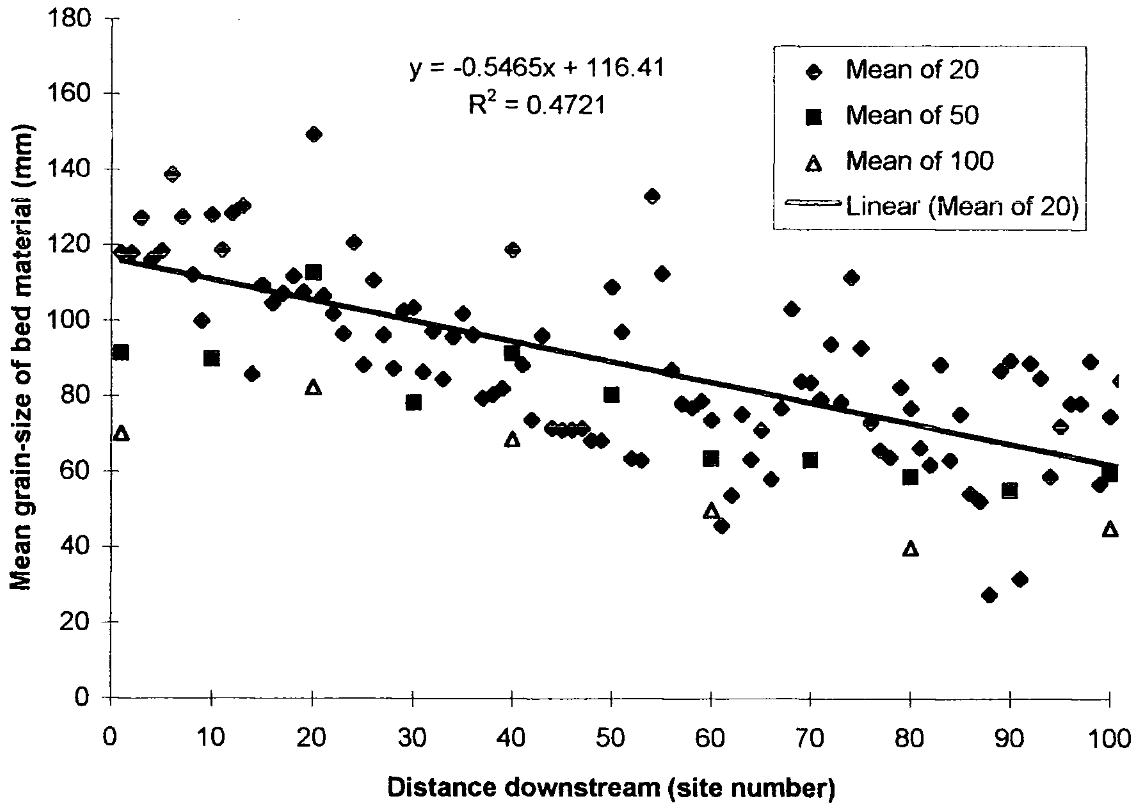


Figure 5.3 Downstream changes in mean grain-size of bed material

5.5 The relationship between cross-sectional form, mean grain-size and the sequence of pools, riffles and bends

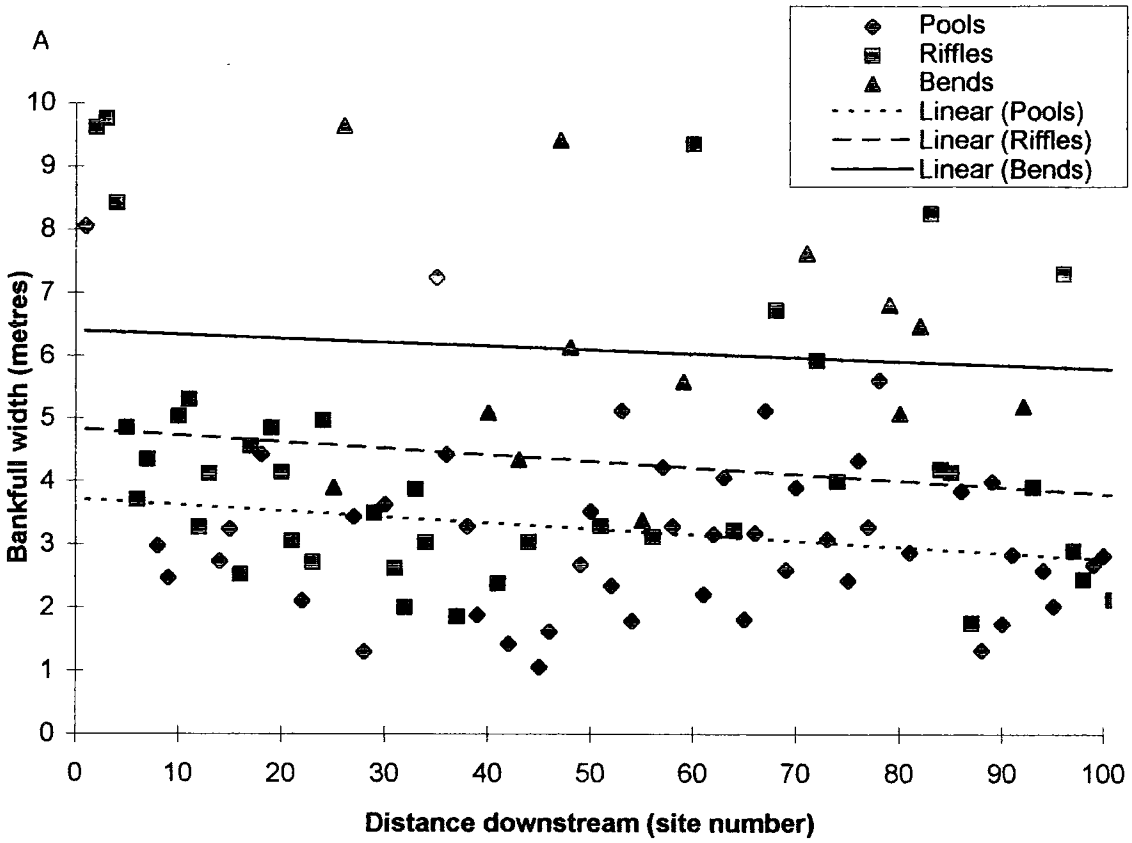
In order to establish the relationship between localised changes in cross-sectional form and bed topography, each cross-section was classed as either a pool, riffle or bend. In this study 47 sections are categorised as pools in straight reaches, 41 as riffles in straight reaches and 13 on bends, (10 of which are pool sites). However, the definition of individual pools and riffles within a sequence can be problematical, particularly where the cross-section is located on the boundary between one of the three bedform types. Riffles were identified as those bedforms which generally coincided with a sequence of coarse sediment within a relatively wide, shallow section of channel whereas pools were identified as those bedforms which showed the opposite. The differentiation between pools and riffles was clearest where bend curvature was high and where bedforms were well developed. Variations in bed topography based on a division of the channel into pools, riffles and bends allows differences in cross-sectional response of the channel to the passage of a flood to be clearly identified.

Downstream changes in bankfull width in relation to the pool, riffle and bend sequence are illustrated in Figure 5.4(a). In all three cases there is a slight downstream decrease in bankfull width although the slopes of all three regression lines are the same. The upper regression line indicates that meander bends occupy the widest channel locations. Bend apices are usually sites of deep pools, i.e. 10 out of 13 meander bends sampled in the cross-sectional survey are occupied by pools. Since pools associated with the bend apex help to focus erosion at the concave bank thus increasing the amplitude of the bend (Milne, 1982c) this explains the increase in bankfull width at meander bends in comparison with pools and riffles in straight reaches.

The linear regression lines plotted for pools and riffles in the relatively straight reaches indicates that riffles tend to be wider than pools but not as wide as meander bends. This may be explained by the coincidence of riffles with areas of coarse floodplain sediments and pools with finer grained cohesive sediments which tend to occupy a narrower and deeper cross-sectional form (Milne, 1982c; Knighton, 1998).

Downstream changes in bankfull depth in relation to the pool, riffle and meander bend sequence are illustrated in Figure 5.4(b). In the case of pools, riffles and meander bends there is a downstream increase in bankfull depth. The limitations of using measurements of bankfull width to identify bankfull depth are demonstrated in Figure 5.4(b) since the regression line suggests that riffles are deeper than pools upstream of section 40 which is not the case. However, the two regression lines are so close and the scatter so great that any differences would not be significant.

The greatest downstream increase in bankfull depth is shown by the linear regression line for meander bends. This may be explained by the occurrence of some very deep pools in the lower reaches where the channel cuts through cohesive, silty clay banks leading to excessive deepening. These very deep pools have been shown to occur at sites on the floodplain where the main flow approaches fine-grained cohesive banks at a



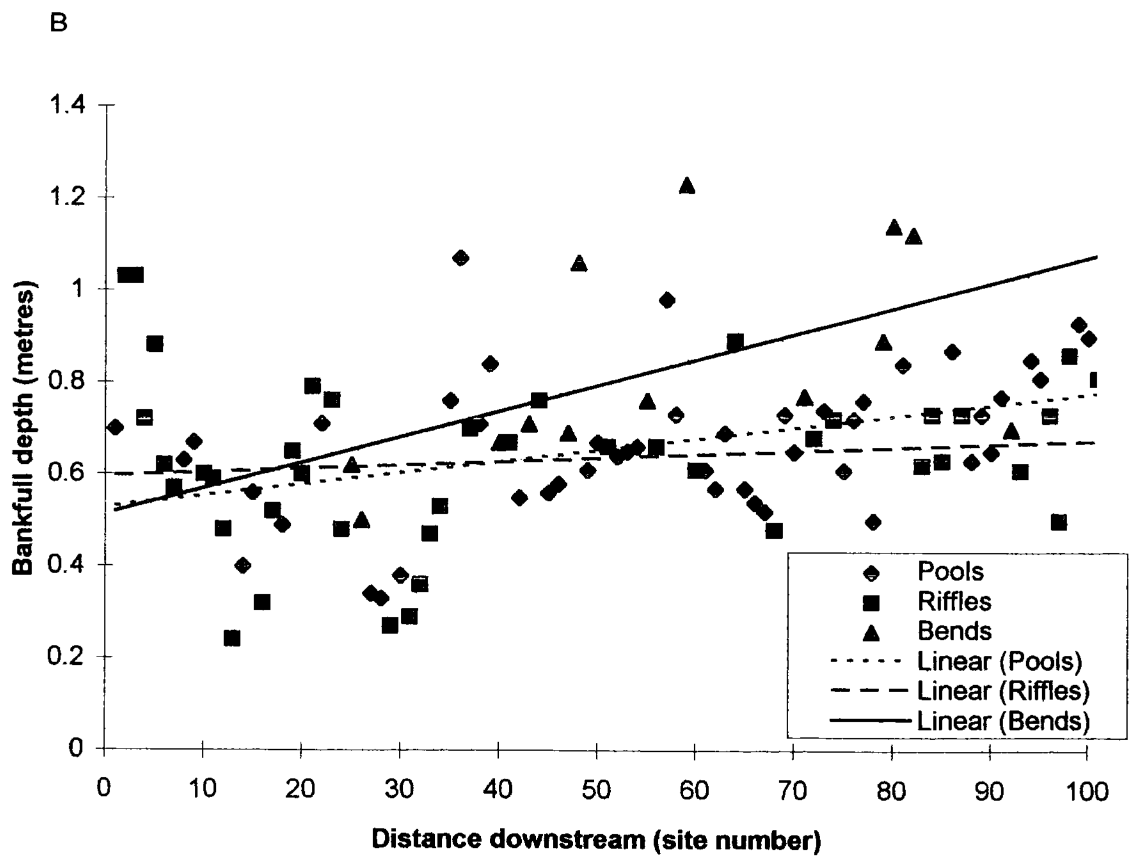


Figure 5.4 Downstream changes in bankfull width (a) and depth (b) of pools, riffles and bends

high angle (Milne, 1982b). The concave bank will tend to maintain a steep angle when retreating.

Evidence of a slight downstream decrease in bankfull width and increase in bankfull depth of all three bedform types may be the result of increasing downstream bank cohesion. When banks are more cohesive they are more resistant to erosion during flood events. Adjustments in the cross-sectional form of the channel occur through vertical erosion of the bed rather than lateral erosion of the banks.

5.5.1 Downstream changes in width:depth ratio in relation to the pool, riffle and meander bend sequence

A downstream decrease in width:depth ratio for pools, riffles and meander bends is shown in Figure 5.5. As mentioned earlier the banks in the upper reaches of Swinhope Burn tend to be less cohesive, often consisting of cobbles in a fine matrix of sandy silt, which generally correlates with a wider, shallower channel. This is in contrast with the lower reaches where the banks are much more cohesive, and the resultant channel is narrower and deeper. A similar pattern is suggested by Milne (1982b) for Kingledoors Burn, a small gravel-bed stream in the Scottish Uplands. Here a wide and shallow cross-sectional form developed as a result of gravel and cobble lateral deposits within the floodplain and in response to a dominantly bedload transport regime.

Examples of surveyed cross-sections from the upper and lower reaches of Swinhope Burn (Figure 5.6) illustrates the cross-sectional form of the channel in relation to the pool, riffle and meander bend sequence. The influence of bank material is a critical factor in determining the development of cross-sectional form as will be identified in the next section. Site 26 and site 82 (Figure 5.6) are both located on meander bends. At site 26 the cross-sectional profile is wider and shallower than site 82. The cohesive nature of the bank material in the lower reaches prevents lateral erosion and encourages deepening of the pool at the apex of the meander bend leading to a narrower and deeper cross-section. Conversely the lack of cohesiveness of the banks in the upper reaches promotes widening of the bend through lateral migration of the channel. A similar pattern can be identified for pools and riffles on straight reaches.

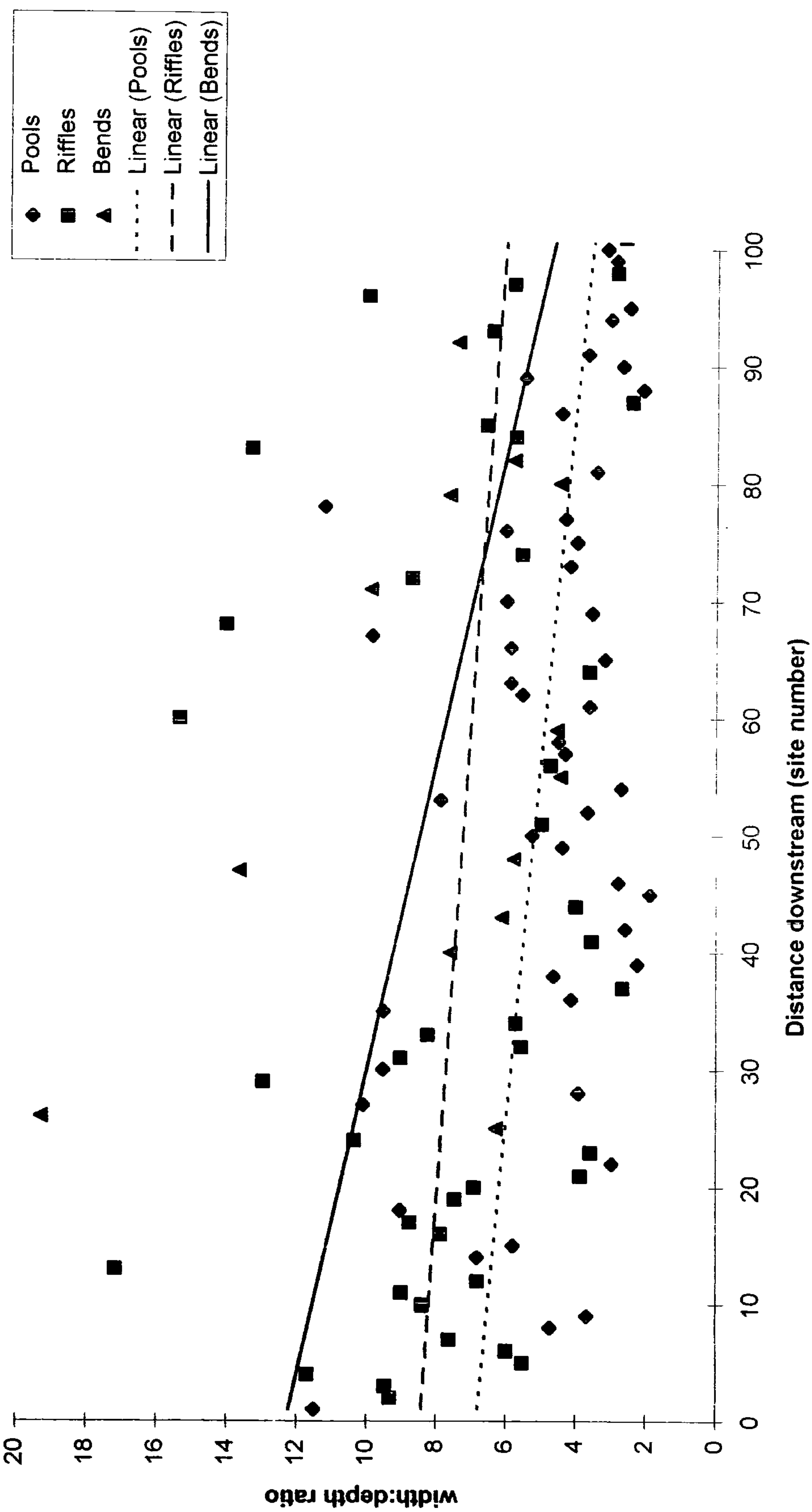
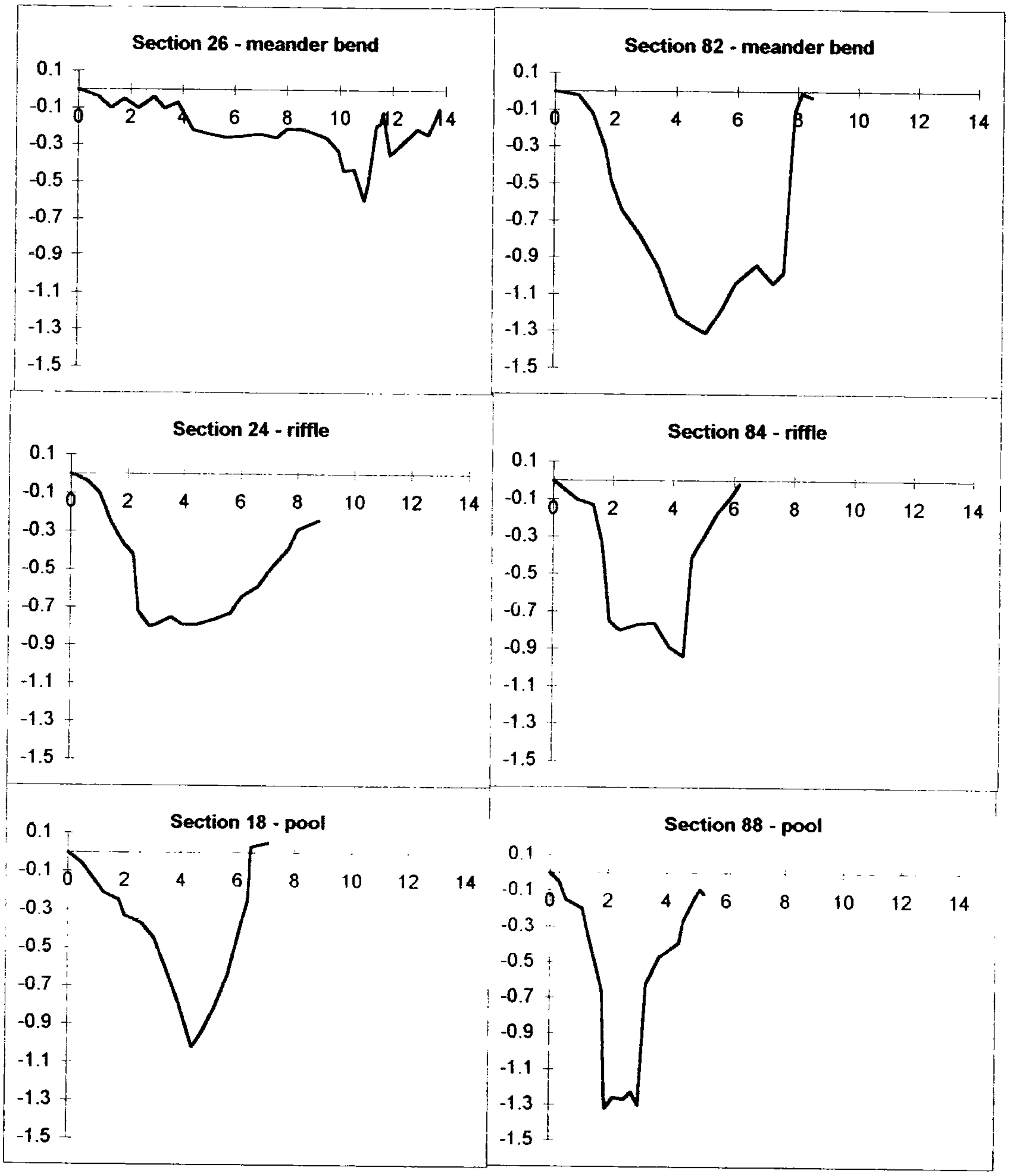


Figure 5.5 Downstream changes in width:depth ratio of pool, riffles and bends



Upper reaches of Swinhope Burn **Lower reaches of Swinhope Burn**

Figure 5.6 Examples of surveyed cross-sections from the upper and lower reaches of Swinhope Burn (vertical exaggeration approximately 7x).

However, a low channel gradient is another factor which influences downstream changes in width:depth ratio. The combination of low channel slope and cohesive bank sediments in the lower reaches is largely responsible for the tight bend curvature and low width:depth ratios.

Meander bends tend to occupy positions in the channel with the highest width:depth ratio (Figure 5.5). The trendline shows the greatest downstream decrease in width:depth ratio, which can be explained by the development of the deepest pools on bends in the lower reaches due to the cohesive nature of the bank material. Pools at meander bends tend to become fixed at bend apices which may lead to increased bend curvature and increased width. This explains the high width:depth ratio exhibited by meander bends in the study reach.

Figure 5.5 shows that riffles tend to be wider and shallower than pools, and in this example riffles exhibit a higher width:depth ratio than meander bends in the lower reaches of Swinhope Burn. The pools in straight reaches are narrower and deeper than riffles or meander bends. Figure 5.7 shows that riffles on meander bends can produce a wide and shallow channel as the flow is deflected from mid-channel accumulations of coarse material causing erosion of both banks. These observations are supported by studies of coarse bedload channels in upland Britain (Milne, 1982b, 1982c) which suggest that pools in straight reaches with a low width:depth ratio relate to cohesive floodplain sediments on both sides of the channel. In comparison it was found that riffles in straight reaches had a greater width:depth ratio resulting from flow diverging from symmetrical central bars and attacking both banks (Milne, 1982c).

5.5.2 Downstream changes in mean grain-size of bed material in relation to the pool, riffle and meander bend sequence.

The difference in mean-grain size between pools and riffles is fairly constant downstream (Figure 5.8). Riffles are coarser than both meander bends and pools. The linear regression line for meander bends converges with pools upstream and riffles downstream. The mean grain-size of bed material sampled from pools in straight reaches are finer than those sampled either from riffles or meander bends. In the deepest pools this may have resulted from infilling by fine material during low flows

Figure 5.7 Photograph of riffle on a meander bend showing high width:depth ratio of the channel (section 59)

(Milne, 1982b). These observations are supported by Keller (1971) who suggested that the velocity distribution over pools and riffles actively sorts the bed material so that the coarsest bed material is deposited on riffles at high flow and the finest bed material is deposited in pools below the level of velocity reversal.

However, Figure 5.8 indicates that there is very little local variation in mean-grain size of pools, riffles and meander bends . These observations are supported by Milne (1982b) for a small, coarse bedload, upland stream: he suggested that riffles and pools are 'downstream oscillations in elevation of a single coarser grained sedimentary unit' and possessed a relatively similar size distribution. However, Knighton (1998) suggests that riffles have coarser bed material than adjacent pools and that through a combination of 'high flow transport through pools and low flow storage on riffles' the pool-riffle sequence is maintained through coarser material being concentrated on the riffles. Conversely, Carling (1991) suggested that pool sediments sampled from the River Severn tended to be coarser than riffle sediments. Carling (1991) argues that as flow discharge increased, the bed roughness of pools decreased rapidly and this led to

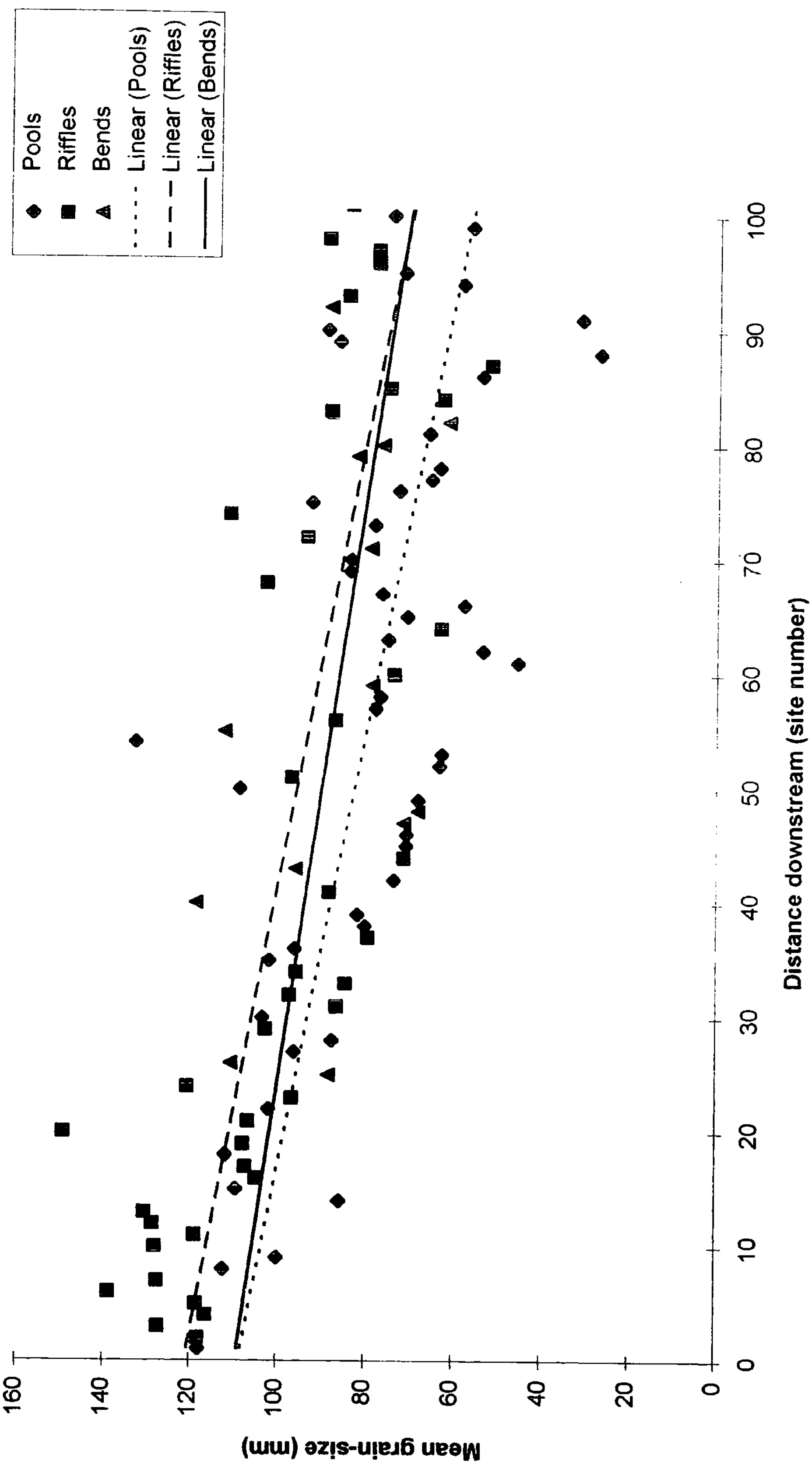


Figure 5.8 Downstream changes in mean grain-size of bed material of pools, riffles and bends

pools being partially infilled, which may account for pool sediments being coarser than riffle sediments in the River Severn. Clifford (1993) and Sear (1996) argue that the distinction between the mean grain-size of pools and riffles can only realistically be made in terms of bed structure. A loose bed structure in pools during high flows leads to the rapid mobilisation of sediment, with pool sediments travelling further during flood events and sediment being transported from pool to pool with storage on riffles.

Summary

Meander bends occupy the widest channel locations. Pools associated with the meander bend apex focus erosion at concave banks increasing the amplitude of the bend. However, the nature of bank sediments is critical in influencing the development of adjacent bed forms. For example, very deep pools may form where the upstream current encounters fine-grained cohesive bank sediment at a high angle (Milne, 1983a) Riffles, however, tend to form where bank sediments are coarse or where there are composite banks. The downstream decrease in width:depth ratio for pools, riffles and meander bends, resulting from a downstream increase in the cohesiveness of bank material and a slight decrease in slope is clearly illustrated using examples of surveyed cross-sections from both the upper and lower reaches of Swinhope Burn (Figure 5.6).

Riffles are bedforms with the coarsest mean-grain size distribution, although the difference in mean grain-size found between pools and riffles was small, supporting Milne's (1982b) observations for upland gravel-bed streams. Riffles are the most stable bedform because they tend to have the coarsest bed material which has a closely packed structure and which resists erosion during high flow events (Clifford, 1993; Sear, 1996). Once riffles have formed, the accumulations of coarse bed material, often mid-channel, or the presence of central bars deflect flow towards one or both banks which leads to bank undercutting and increased channel widening (Milne, 1982c; Knighton, 1998).

Pools tend to have the finest mean grain-size distribution and occupy narrower, deeper locations within the channel. Pools form where bank material is more cohesive which tends to reduce lateral widening of the channel.

5.6 The effects of downstream variations in bank material grain-size and local topography on the development of cross-sectional form

The purpose of this section is firstly to examine the effects of downstream variations in the grain-size of bank material on the width:depth ratio of the channel and the mean grain-size of the adjacent and downstream bed material. Secondly, the influence of channel bars, cut banks and toe deposits on the development of cross-sectional form and the mean grain-size of adjacent and downstream bed material will be examined.

The classification for bank material closely follows the Wentworth grain-size classification whereby the finest grain-size group is clay and the largest grain-size group is very large boulders. There are 15 intermediate grain-size groups ranging from silt and sand to gravel and cobbles. The smallest grain-size is awarded the lowest value (1) with the largest grain-size being allocated the highest value (17) (Table 5.1). For example, coarse gravel, which is mid-classification is awarded a value of 10. Although this method of assigning values to grain-sizes sampled from bank material is only semi-quantitative it gives an indication of downstream variations in grain-sizes sampled from each of the 101 cross-sections within the study reach. The grain-size of bank material at each site is shown for both the left (+ve values) and right (-ve values) banks. If two grain-sizes are recorded at a particular site, as in the case of composite banks, the total becomes cumulative with the second bar being plotted either above (for +ve values) or below (for -ve values) the initial one. In the key, the finest grain-size sampled is plotted as a black bar and is marked with the number (1) whereas the second coarser grain-size sampled is plotted as a clear bar and is marked with the number (2).

5.6.1 Downstream variations is grain-size of bank material in relation to width:depth ratio and mean grain-size of bed material.

Figure 5.9 and Figure 5.10 show downstream variations in the grain-size of right and left bank material in relation to changes in width:depth ratio and mean-grain size of the bed respectively. Upstream of section 23 there is generally more variation in the grain-size of bank material and a large proportion of these sections are composite banks and have both fine and coarse grain-sizes present.

Class	Bank material type
1	clay
2	silt clay
3	sandy silt clay
4	sandy silt
5	fine sand
6	medium sand
7	coarse sand
8	fine gravel
9	medium gravel
10	coarse gravel
11	small cobble
12	mixed cobble
13	large cobble
14	small boulders
15	medium boulders
16	large boulders
17	very large boulders

Table 5.1 Grain-size classification for bank material based on Wentworth grain-size classification

The composition of the banks generally becomes more uniform downstream of section 20-25 although there is still considerable local variation with some very coarse grain-sizes present, particularly at cut-bank locations.

The width:depth ratio decreases downstream and the general pattern is for the grain-size of bank material to decrease also (Figure 5.9). Both the width:depth ratio and the grain-size of bank material show a similar extent of local variation. Upstream of section 20-25 the banks are predominantly composed of gravel and cobbles in a matrix of silty clay

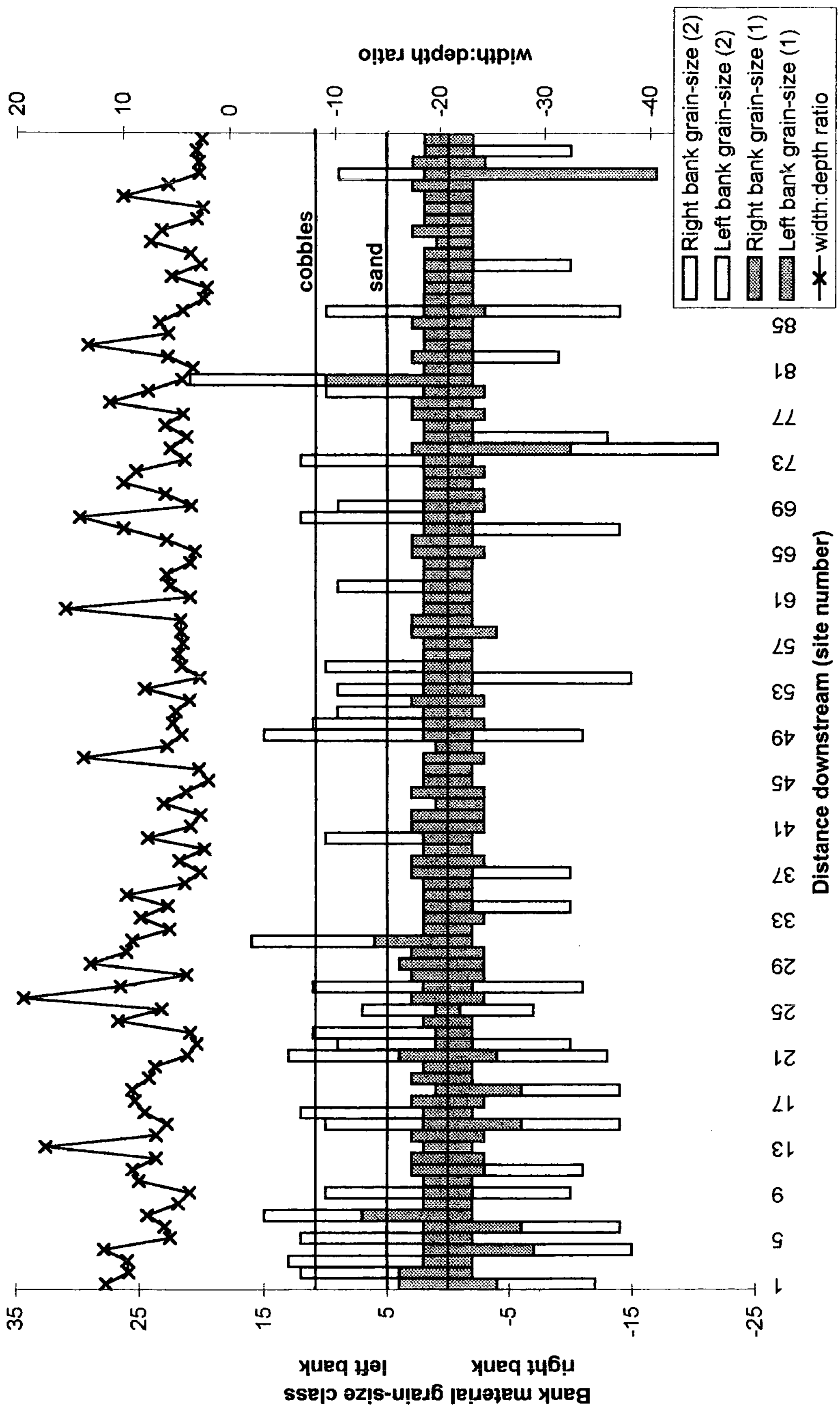


Figure 5.9 Downstream variations in the grain-size of right and left bank material in relation to changes in width:depth ratio

and are less cohesive than in the lower reaches. Since the banks are less cohesive, erosion tends to be lateral leading to a higher width:depth ratio than further downstream and the channel is relatively straight, shallow and wide, particularly upstream of section 14.

There is increasing uniformity of bank material downstream of section 30 with many bank sites consisting of mainly silt clay (Figure 5.9). Since the banks are more cohesive, erosion tends to be vertical rather than lateral. The resulting cross-sectional form in the lower reaches is deep and narrow with a meandering planform and a lower width:depth ratio, for example sections 36-46 and 84-101. Milne (1982b) found that a similar pattern of bank material variation controlled planform development at Kingledoors Burn in the Scottish Uplands (Milne 1982c). He found that the presence of gravel and cobble lateral deposits just below the floodplain surface aided the development of a wide, shallow channel, whereas pockets of fine-grained cohesive bank material controlled the development of deep pools on tight meander bends. Milne (1982c) found that on other gravel-bed streams in upland Britain, where the channel cut through fine-grained cohesive sediment on both sides of the channel, a very narrow and deep cross-sectional form resulted.

Downstream variation in bank material in relation to mean-grain size of bed material is illustrated in Figure 5.10. In the upper reaches both bank material and bed material are coarser than in the lower reaches. Equally, in downstream reaches, where there is greater local variation in the grain-size of bank material, there is also greater local variation in the mean grain-size of bed material.

However peaks in bed material grain-size do not necessarily correspond with peaks in bank material grain-size at each section. Downstream of section 74 where the majority of sections have fine, cohesive banks the mean grain-size of the bed material is also finer. This results from the lack of coarse grain-sizes in the bank material and the inability of the stream to transport coarse bedload along a channel. In the upper reaches coarse grain-sizes on both the right and left banks generally correspond with coarse bed material at each section. Coupling between bank and bed material in the upper reaches is better developed than in the lower channel.

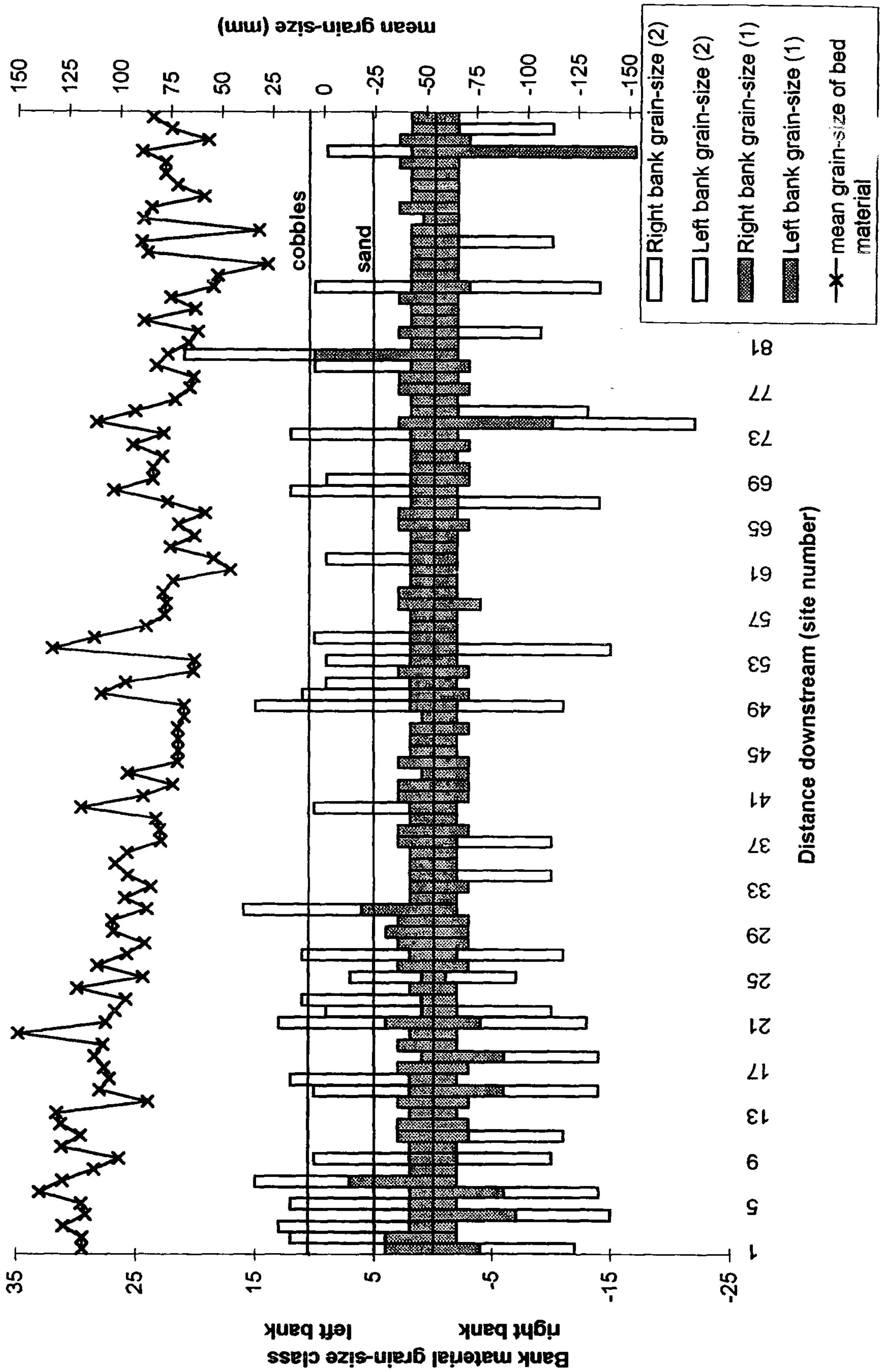


Figure 5.10 Downstream variations in the grain-size of right and left bank material in relation to changes in mean grain-size

5.6.2 The relationship between the presence of bars and cut banks on downstream variations in the width:depth ratio and mean grain-size of bed material

The presence of bars and cut banks at the base of banks influence the development of cross-sectional form and the mean grain-size of the adjacent and downstream bed material. Bars and cut banks are considered jointly since each is a concentration of coarse material within or immediately adjacent to the channel. Concentrations of coarse material in the channel can both reduce bank erosion or encourage erosion of the opposite bank by altering local flow patterns.

Within the study reach bar and slumped bank locations correspond with cross sections where a high width:depth ratio and relatively coarse mean grain-size of bed material is recorded. Figure 5.11 shows an example in which the development of a bar on the right bank corresponds with a high width:depth ratio. In sections where one bank has a coarse lateral bar and the other is a cut-bank (section 13), a supply of coarse sediment results in cross-sections with a high width:depth ratio since the banks are easily eroded and bed erosion is inhibited (Figure 5.12a). The coarse nature of the bed material can be explained through the input of coarse material into the channel from the cut bank. Milne (1983a) has suggested that peaks in width:depth ratio are likely to occur at cut bank locations where the stream impinges on the valley-side slopes since the increased supply of coarse sediment prevents depth adjustment within the channel in relation to its width. However, conversely where coarse material is supplied to the channel from the cut bank, it protects the base of the bank from deep scouring, particularly on meander bends and this can decrease the width:depth ratio in these locations (Milne 1982a).

Other examples of where peaks in width:depth ratio were recorded were on point bars opposite pools on meander bend apices (sections 26 and 47) and on riffles (sections 60 and 83). Likewise, Milne (1983a) identifies deep pools on well-developed bends and riffles as locations at which there is likely to be a peak in width:depth ratio. The coarseness and tight structure of the bed material forming riffles tends to inhibit bed erosion, but encourages lateral widening of the channel as flow is directed away from mid-channel accumulations of coarse material causing bank undercutting.

Figure 5.11 Photograph showing an example in which a bar corresponds with a high width:depth ratio

However, bar locations are not always associated with a high width:depth ratio (section 64). In these instances it may be that the banks are protected from lateral erosion and this encourages vertical erosion of the channel bed decreasing width:depth ratio.

Locations with coarse toe deposits at the base of banks show a decrease in width:depth ratio particularly where they are present at the base of both banks or where there is a bar present on the opposite bank. It appears that the base of the bank is protected by these coarse toe deposits preventing bank erosion unless there is a flood capable of removing them.

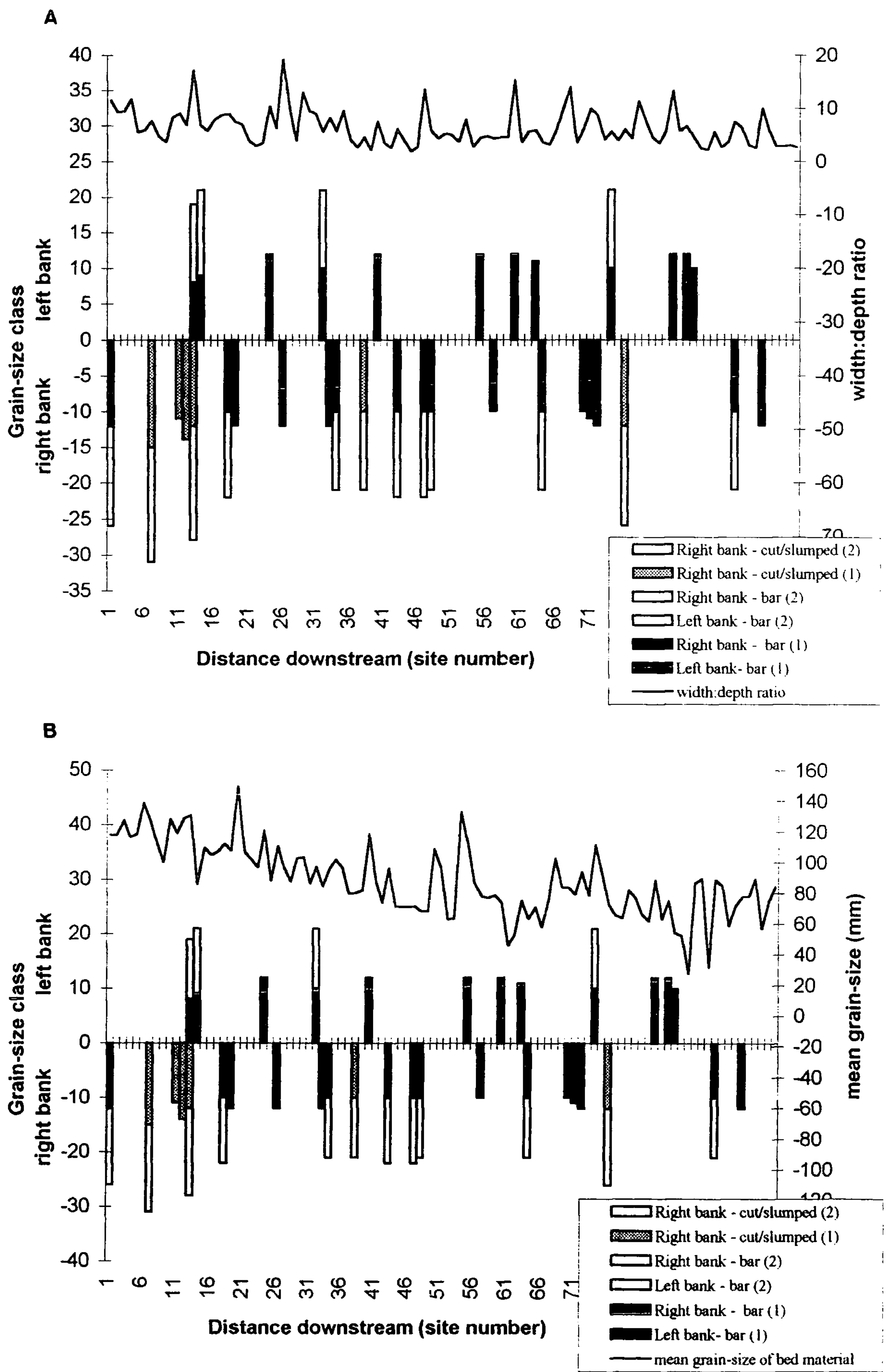


Figure 5.12 Downstream changes in the relationship between the presence of bars and cut banks and variation in the width:depth ratio and mean grain-size of bed material

The relationship between the occurrence of bars and cut banks with downstream changes in the mean-grain size of bed material is investigated (Figure 5.12b). Peaks in mean grain-size of the bed material tend to correspond with bars. Milne (1982) identified bars as the most distinctive bedform, in terms of mean grain-size, on Kingledoors Burn in the Scottish uplands. A similar pattern was found at cut-bank locations. A series of cut-banks from section 12 to section 14 show a corresponding increase in mean grain-size of bed material adjacent to these sites.

Summary

A downstream decrease in the grain-size of bank material along the study reach has resulted in a downstream decrease in width:depth ratio. A comparison of grain-size of bank material with the mean grain-size of bed material shows a loose coupling between bed and banks, particularly at cut banks and bluffs where coarse angular sediment is introduced into the channel. Bars both interact with and influence the flow pattern. They influence channel morphology because they are major stores for bedload and because they encourage bank erosion of the opposite bank through flow being directed away from the bar surface. However, erosion of one bank is approximately compensated by deposition against the other.

5.7 Channel sinuosity

The study reach at Swinhope Burn flows in a predominantly shallow channel along an irregular meandering course and becomes increasingly sinuous downstream (Figure 5.13, Table 5.2). The final bend of the study reach (bend 8, Figure 5.13) is particularly tortuous and during high flow events may become partly cut off with the flow being directed across the floodplain before it re-enters the channel some distance downstream. Although there are some highly sinuous bends (bends 5a, 6, 6c and 7) these are interspersed with relatively straight reaches, for example, between 6b and 6c and between 7b and 7c (Figure 5.14). It may be that these bends coincide with pockets of coarser floodplain material, reducing bank resistance to erosion and producing a straighter channel (Milne, 1983a) or they could be sites of previous cutoffs.

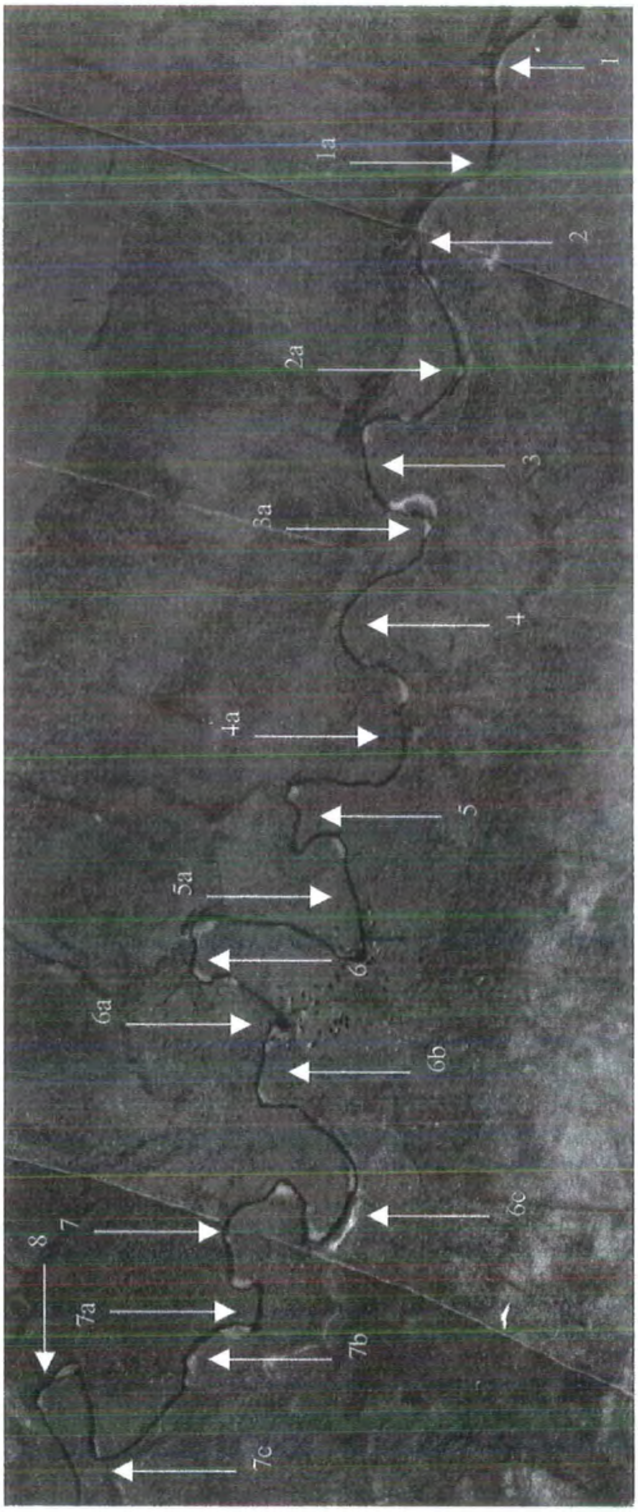


Figure 5.13 Air photograph of Swinhope Burn showing channel sinuosity with bends labelled (courtesy of Aerofilms and Durham County Council).

Bend	Section Number	Bend Sinuosity
1	7	0.96
1a	10	1.15
2	13	1.29
2a	18	1.48
3	21	1.76
3a	25	1.65
4	31	1.7
4a	35	2.04
5	40	2.53
5a	47	3.82
6	53	3.33
6a	59	1.58
6b	63	1.67
6c	71	3.28
7	76	3.18
7a	80	3.14
7b	85	1.06
7c	92	1.58
8	96	6.17

Table 5.2 Downstream variations in channel sinuosity

Figure 5.14 Photograph of high and low sinuosity bends in the lower reaches of Swinhope Burn.

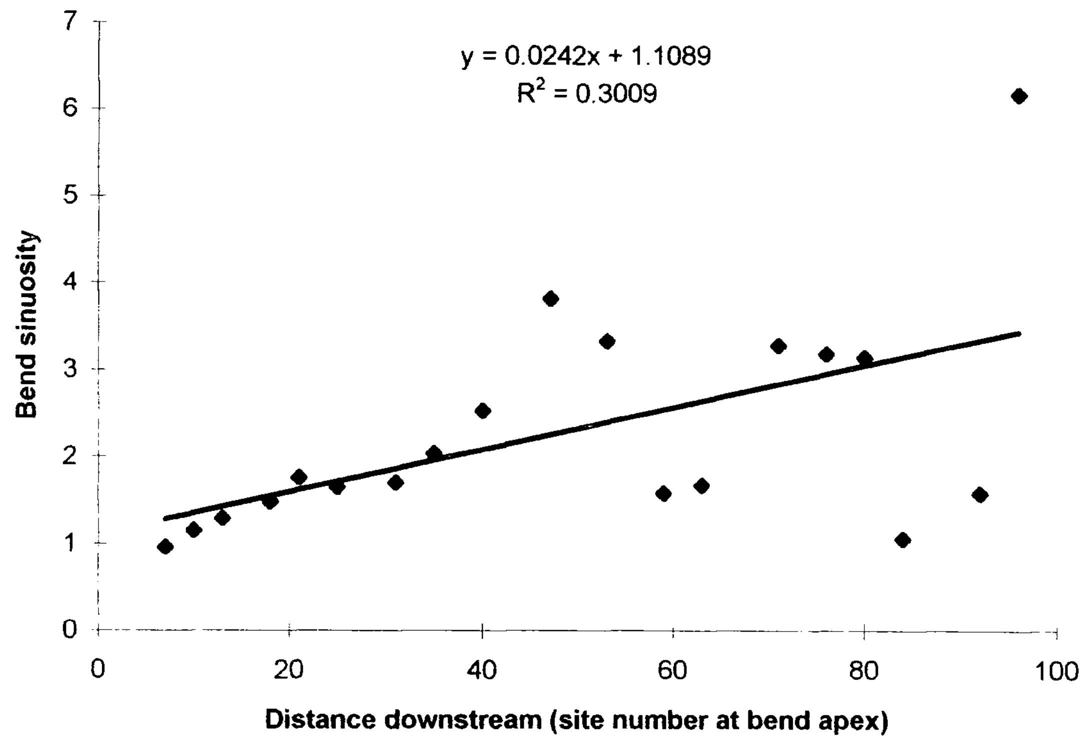


Figure 5.15 Downstream variations in bend sinuosity, Swinhope Burn

Figure 5.15 shows a clear positive relationship between bend sinuosity and distance downstream. This relationship results from a combination of two factors. Firstly, channel gradient decreases downstream, for example in the upper reaches, channel gradient is 0.0172 compared with 0.0091 in the lower reaches, and the banks become increasingly cohesive downstream, as shown in Figure 5.9. Knighton (1998) has argued that the resistance of the banks controls the ability of a stream to shift laterally thus exerting a major control over channel pattern. Milne (1983a) supports this theory, suggesting that where the channel cuts through cohesive floodplain sediments the channel will become increasingly sinuous with some very tight meander bends developing where the current from upstream encounters cohesive bank material.

5.8 The role of channel gradient in determining cross-sectional form and downstream fining of bed material

When compared to other upper Weardale tributary streams, Swinhope Burn has a very distinct long profile. Although the long-profile of Swinhope Burn is concave upwards, after an initially rapid decrease in slope, the basin flattens markedly producing a step in the profile, and this is where the study reach is located. The study reach begins at an altitude of approximately 406 metres and ends at an altitude of approximately 389 metres at the point at which the gradient begins to fall steeply as the stream incises through the Greenly Hills moraine (NY 912 356). The 101 cross-sections cover almost the entire length of the profile, therefore and the study reach encompasses this distinctive segment of Swinhope Burn. The channel gradient along the 1.4 km study reach is very gentle (0.012) but is marginally steeper in the upper reaches compared with the middle and lower reaches.

5.8.1 Detailed long-profile of channel gradient of Swinhope Burn.

Figure 5.16 shows the detailed long-profile of channel gradient of Swinhope Burn. The overall channel gradient of the study reach is 0.012 which is relatively low when compared with the upper reaches of other adjacent streams in Weardale. Such a low channel gradient has implications for both the way in which the cross-sectional form of the channel responds to flood events and the extent of downstream fining of the mean grain-size of bed material.

A sequence of high and low elevations, associated with the pool and riffle sequence can be identified on Figure 5.16. In the upper reaches differences in bed elevation between pools and riffles is very slight owing to the lack of cohesivity of bank material which encourages lateral erosion of the channel thus inhibiting vertical erosion. An increasing number of deep pools are evident in the lower reaches of Swinhope Burn where the bed becomes increasingly irregular in depth as the channel becomes more sinuous. Increasing bank cohesivity and a reduction in channel slope in the lower reaches impedes lateral widening of the channel, thus increasing vertical deepening which accentuates the differences in elevation between pools and riffles observed on Figure 5.16. Variations in channel gradient show an undulating pattern which corresponds

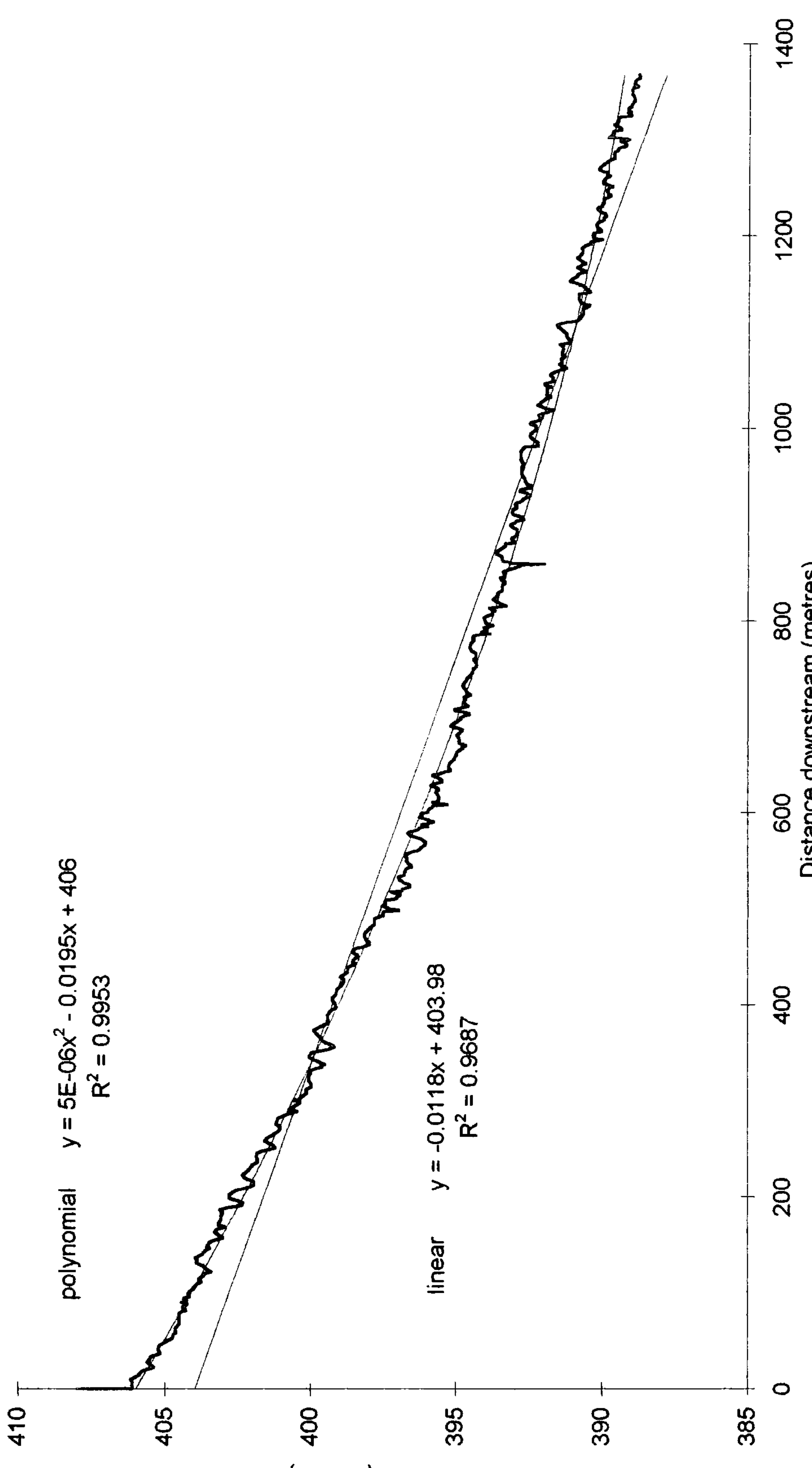


Figure 5.16 Detailed long-profile of channel gradient at Swinhope Burn

with a cyclical meandering pattern characteristic of the study reach. However this is less obvious in the upper reaches, where some channel segments are relatively straight.

Table 5.3 shows the differences in gradient when the study reach is divided into three sections, the upper, middle and lower sections.

Section	Elevation (metres)	Distance downstream (metres)	Channel gradient	Stream power (Wm^{-1})
Upper	406 - 398	0 - 460	0.0172	0.42
Middle	398 - 393	460 - 920	0.0109	0.27
Lower	393 - 388	920 - 1370	0.0091	0.22

Table 5.3: Downstream variation in channel gradient for Swinhope Burn

Table 5.3 indicates firstly that there is very little downstream change in channel gradient from the upper to the lower reaches, with a change in slope of 0.0081, and secondly that the overall gradient is very low. The gradient falls 17 m over a distance of 1400 m. However, even slight variations in channel gradient from one section of the channel to another can influence the extent of downstream fining of bed material and changes in cross-sectional form and channel pattern. Stream power falls from 0.42 Wm^{-1} in the upper reaches to 0.22 Wm^{-1} in the lower reaches as a result of decreasing channel gradient. Stream power calculations are based on the following equation:

$$\Omega = \rho\, g\, Q\, S \text{ /unit channel length}$$

where $\rho = 1$, $g = 9.81 \text{ m s}^{-1}$, $Q = 2.5 \text{ m}^3\text{s}^{-1}$ and S varies downstream (Table 5.3). 2.5 m^3s^{-1} is the estimated bankfull discharge for Swinhope Burn (Warburton and Danks, 1998). When compared with stream power at bankfull discharge in other upland streams values shown in Table 5.3 for Swinhope Burn are very low (Ferguson, 1981).

5.9 Discussion and Conclusions

The irregular cross-sectional geometry and channel planform of Swinhope Burn results largely from downstream variations in bank cohesion, bed topography and the presence of bars and streamside scars. The composition of the banks influences where bank erosion is most likely to occur and consequently conditions patterns of channel sedimentation (Milne, 1982b, Ferguson, 1987). In the upper reaches, a wide, shallow and relatively straight channel pattern has developed in response to the composite banks which consist of gravel and cobbles overlain by finer floodplain sediments. In the lower reaches where the banks mainly consist of cohesive silt clay, the channel is narrow, deep and meandering.

Although riffles and particularly meander bends occupy the widest channel locations, the nature of the bank sediment is critical in influencing the development of the adjacent bed forms. In the lower reaches, deep pools form on meander bend apices where the upstream current meets fine-grained cohesive bank sediment at a high angle. These bends become persistent features through erosion of the concave bank and deposition on adjacent point bars (Milne, 1982c, 1983a).

However, along the study reach channel gradient is very low and there is a slight decrease in slope from the upper to the lower reaches. This decrease in channel gradient has implications for changes in channel pattern. In the lower reaches of Swinhope Burn, a combination of highly cohesive bank sediment and low channel gradient (0.0091) has led to the development of a highly sinuous channel (Figure 5.14) of low mean grain-size, which is in contrast with the upper reaches of Swinhope Burn.

Although riffles on straight reaches have a slightly coarser mean grain-size than pools on either straight reaches or bends, there is very little variation in mean-grain size from one bedform to another (Milne, 1982b). However, there is a very clear pattern of downstream fining in the mean-grain size of bed material which is related to the channel gradient. The low channel gradient at Swinhope Burn has resulted in a low stream power, particularly in the lower reaches (0.22 Wm^{-1}), which reduces the capacity for transport of the coarsest fraction of the bedload through the reach during flood events. In effect, Swinhope Bottoms has many characteristics of a sedimentation zone.

CHAPTER 6

CONTEMPORARY RIVER CHANNEL CHANGE: THE RESPONSE OF AN UPLAND STREAM TO FLOODS

6.1 Introduction

A full appreciation of the way in which upland channels respond to flood events requires a knowledge of both historic and contemporary river channel change. As illustrated in a previous chapter, although a series of detailed maps, air photographs and flow records from adjacent catchments may be available for the historical period, it is still difficult to identify the role of individual flood events in causing channel change. In ungauged catchments it is very difficult to precisely date the occurrence of individual flood events since the use of indirect source materials provides a sampling interval too coarse to identify the impact of each flood event. Even where a flood event can be precisely dated and correlated with channel changes identified on maps or air photographs, it is still only possible to assess the role of flood events in relation to changes in channel planform.

The study of contemporary river channel change can provide detailed information on three dimensional change in fluvial systems, which is critical in order to identify vertical channel stability or instability. Published maps and air photographs provide evidence of general changes in channel planform at timescales covering centuries down to a few years, whereas contemporary survey of channel cross-sections concentrates on the detailed erosional and depositional effects of individual flood events.

Ferguson (1981) has suggested that the majority of British rivers are naturally inactive and have not changed their courses measurably since maps produced in the late 19th century. He attributes this inherent stability in many rivers to stream power being insufficient to overcome bank resistance. Milne (1979) suggested that cross-sectional shape was relatively stable at flows less than the dominant discharge for eleven confined upland streams flowing through coarse-bed material in Northern England and the Scottish Borders. Analysis of historical river channel change in Swinhope Burn over the past 180 years has established that channel planform has been remarkably stable, with only one period of relatively short duration where the channel pattern changes from a meandering pattern to one of low sinuosity with a braid bar in the upper reaches. Likewise, the recent flood history of Swinhope Burn showed that although high magnitude flood events had passed through the basin, the channel pattern remained largely unaltered. In the light of this knowledge it might be expected that contemporary flood events would have had very little lasting effect on the study reach at Swinhope Burn.

However, upland stream channels are generally viewed as dynamic landforms which often change planform in response to floods (Acreman, 1983; Harvey, 1986; Coxon et al, 1989) . Other streams within the Northern Pennines have shown major channel changes in response to high magnitude flood events (Carling, 1986; Macklin, 1986). Initial observations of the study reach at Swinhope Burn prior to field monitoring suggested that it too was a highly active stream with evidence of steep cut-banks and bluffs along the channel. Extensive areas of fresh gravel within the channel provided additional evidence of the importance of frequent flooding. A flood in the Swinhope Burn basin in January, 1995, the largest recorded within the upper Wear catchment since records began, initially prompted investigations into channel change at the site. Nine months later the evidence of the flood was still widespread in the form of extensive bank undercutting, failure of slopes bounding the channel and the deposition of large cobbles on the outside of meander bends.

In order to test the hypothesis that the study reach is an active gravel-bed channel in which major channel change occurs in response to flooding, downstream changes in the channel form were measured following the passage of a large flood event through the Swinhope Burn basin on 19th and 20th February, 1997.

6.2 Scope of chapter

The objective of this chapter is firstly to provide a background for recent research into contemporary river channel change in the British uplands. A major flood event, which occurred between the 19th and 20th February, 1997 is described and its immediate effects on the channel and adjacent floodplain are discussed. Downstream variations in cross-sectional form and mean grain-size of bed material resulting from the flood are then discussed. The following section examines the role of bed topography, in terms of the pool, riffle, meander bend sequence, local bank cohesivity, reach sinuosity and channel gradient in influencing the nature and extent of channel response to a major flood event. The chapter concludes with a discussion of how each of these factors interrelate to bring about the observed localised changes in cross-sectional form and mean-grain size, and secondly provides a possible explanation for the observed stability of the channel within the study reach at Swinhope Burn.

6.3 Background

The response of upland stream channels to large infrequent flood events has been well documented over the past two decades (Anderson and Calver, 1980; Acreman, 1983; Harvey, 1986; Carling, 1986; Coxon et al, 1989). Often large flood events have occurred during field monitoring of a stream (Hitchcock, 1977; Werritty and Ferguson, 1980; Werritty, 1982; Ferguson and Werritty, 1983) which has enabled pre and post flood channels to be compared. This has provided a valuable insight into the processes of post-flood recovery. For example, Ferguson and Werritty (1983) monitored the episodic advance of alternate bars in the River Feshie in the Cairngorms during floods. Werritty and Ferguson (1980) measured slight changes in elevation of the floodplain during flood events on the River Feshie in order to explain the sudden switching of the stream from one channel to another. Field monitoring of the response of stream channels to flood events is critical in the understanding of contemporary river channel change. Research on the role of flood events on river channel change may concentrate on one particular dimension of adjustment, such as cross-sectional form, channel planform or the long-profile (Figure 6.1). However, all three are interrelated and adjust according to variations in imposed discharge and sediment load. Figure 6.1 shows the interrelations of these components for a braided channel system.

Ferguson (1981) has suggested that the dominant control on downstream changes in channel dimensions is the volume of streamflow carried by the channel during flood events which causes both erosion and deposition. However, he identifies bank material, bedform configuration and channel pattern as secondary controls which have very localised, more variable effects. In the absence of these complications it is assumed that width, depth and width:depth ratio will increase downstream.

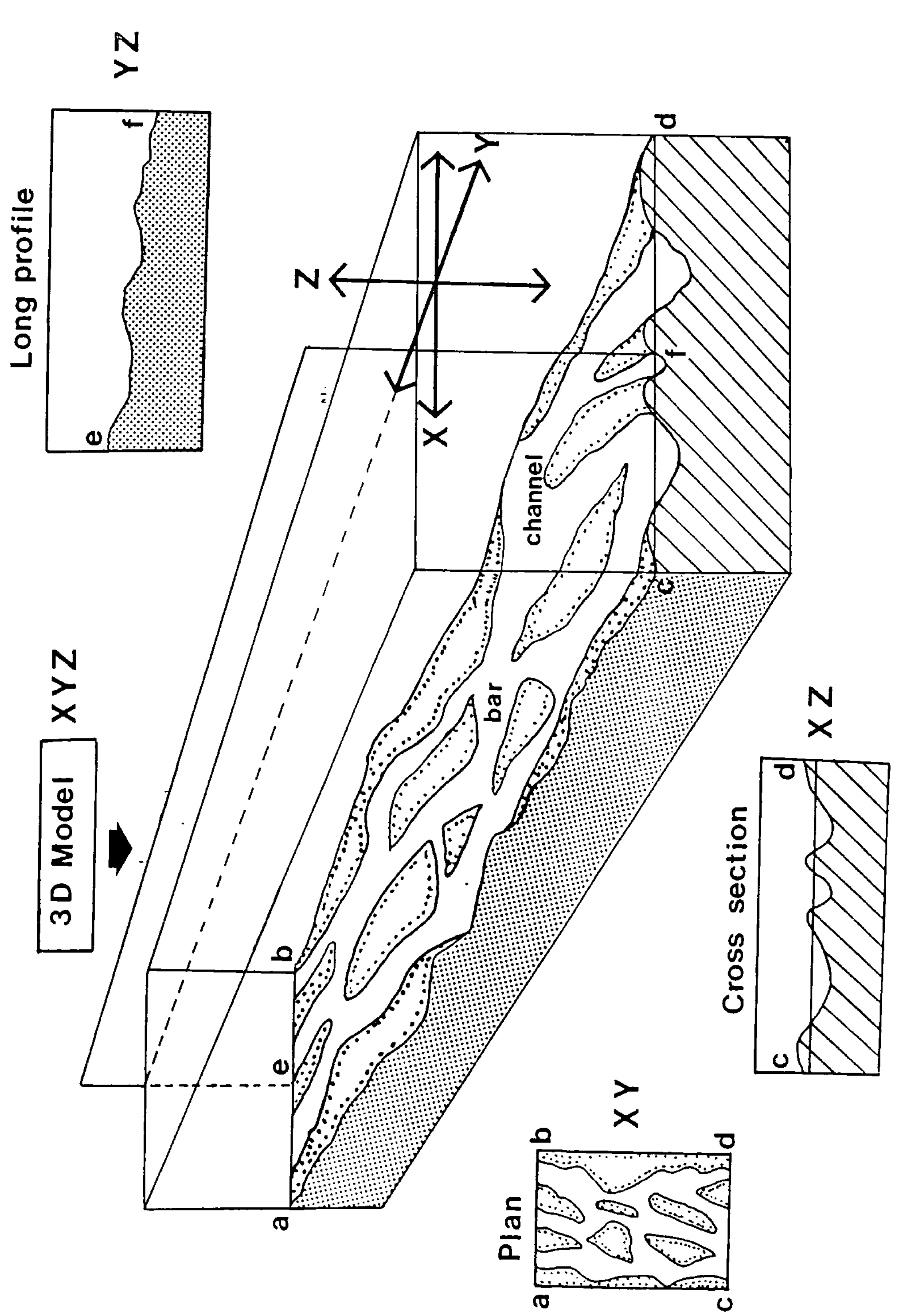


Figure 6.1 The major components of channel form (planform, long-profile and cross-section) represented as different planes of adjustment in a three-dimensional river system. Studies of channel change often focus on a particular dimension (e.g. YZ, XY or XZ) to describe channel adjustments.

Lane et al (1996) stress the importance of the relative timing and magnitude of discharge and sediment fluctuations during flood events on downstream changes in cross-sectional form. Local channel morphology and the occurrence of sediment waves passing through a reach are detailed as the two major controls over morphological change within the channel. Although this research centres on a glacierised catchment it is suggested that discharge and sediment waves associated with a single diurnal discharge fluctuation may be symptomatic of dominant channel forming processes associated with individual flood events in other fluvial environments. The theory that sediment production from localised sources during flood events is routed downstream as a series of waves leading to changes in channel morphology is supported by Richards (1993) and Lane et al (1996).

Similarly, over a distance of 1.2 km with discharge and slope almost constant, the mean diameter of the ten coarsest pebbles on barheads was shown to fall from 270 mm to 155 mm on the River Feshie, Cairngorms as a result of size-selective transport (Ashworth and Ferguson, 1989). A decrease in the percentage movement and mean distance moved by progressively coarser classes of pebbles was demonstrated to be the cause of downstream fining in this upland gravel-bed stream. However, in the case of the Allt Dubhaig in the Scottish Highlands, pronounced downstream fining over a distance of 2.5 km was demonstrated to have resulted from a sharp reduction in slope due to the presence of a local base level, in the form of an alluvial fan. It is suggested that in upland environments, a wave of bedload finer than the existing bed material migrates along the channel and this initially promotes bed material fining with rapid aggradation of the bed. Scatter in downstream variation in grain-size was shown to be due to inputs of sediment from banks and streamside scars. Scatter in downstream variations in grain-size has also been attributed to the effect of tributaries and bank erosion in introducing coarser sediment into the channel (Knighton, 1980). Sorting and abrasion also act on sediment from headwater reaches to cause downstream fining.

Downstream variations in the response of cross-sectional form to the passage of a large flood can be explained with reference to local variations in bed configuration and bank composition and at the reach scale, with variations in channel planform and channel gradient. Milne (1983a) argues that the link between the scale of cross-sectional adjustment and channel planform is related to the mean annual flood. A range of flows will determine channel capacity and through channel width will control channel planform. Milne's (1983a) survey of eleven coarse bedload channels in the British uplands suggested that, when summarised over a reach, mean cross-sectional width is correlated with relatively high but frequently occurring flows. Likewise, Ferguson (1981) indicates that cross-sectional dimensions are adjusted by erosion and deposition so that the channel can contain all but the highest flows. However, Milne (1983a) suggests that an irregular cross-sectional geometry is a common feature of many streams in the British uplands. Milne (1983a) argues that irregular cross-sectional geometries in upland streams result largely from downstream variations in the cohesivity of bank material, variations in bed topography (associated with the pool-riffle sequence), inputs of coarse sediment from bluffs and bend sinuosity.

6.4 The Flood of February 1997 - Field Observations

Between the 19th and 20th February, 1997 a large flood event occurred within the Swinhope Burn basin. Downstream on the River Wear at Stanhope, at 18.30 hours on the 19th February, a peak over threshold value of 160.39 m^3s^{-1} was recorded. This was the fourth highest value recorded on the River Wear since gauging records began in 1958. On the following day, 20th February at 00.00 hours, a peak value of 119.5 m^3s^{-1} was recorded. Mean daily flows were recorded as 49.1 m^3s^{-1} on the 19th February and 51.4 m^3s^{-1} on the 20th February.

In order to determine the spatial extent of the flood, mapping of ephemeral flood limits was undertaken along the study reach at Swinhope Burn on the 21st February, 1997 (Figure 6.2). Flood limits were mapped using Differential GPS and identified by using small-scale geomorphic indicators including vegetation trash lines (Figure 6.3), the distribution of overbank fines, flattened vegetation and standing water (Figure 6.4). Trashlines, four metres from the edge of the channel were most visible at the top of the reach (Figure 6.3) as they incorporated woody debris from a tree plantation immediately upstream. As illustrated in Figure 6.2, most overbank flooding occurred on the left bank of Swinhope Burn since there are a number of steep streamside scars on the right bank and in places the channel is confined by steep valley side slopes. The width of the flooded area increases downstream since in the lower reaches the channel is less

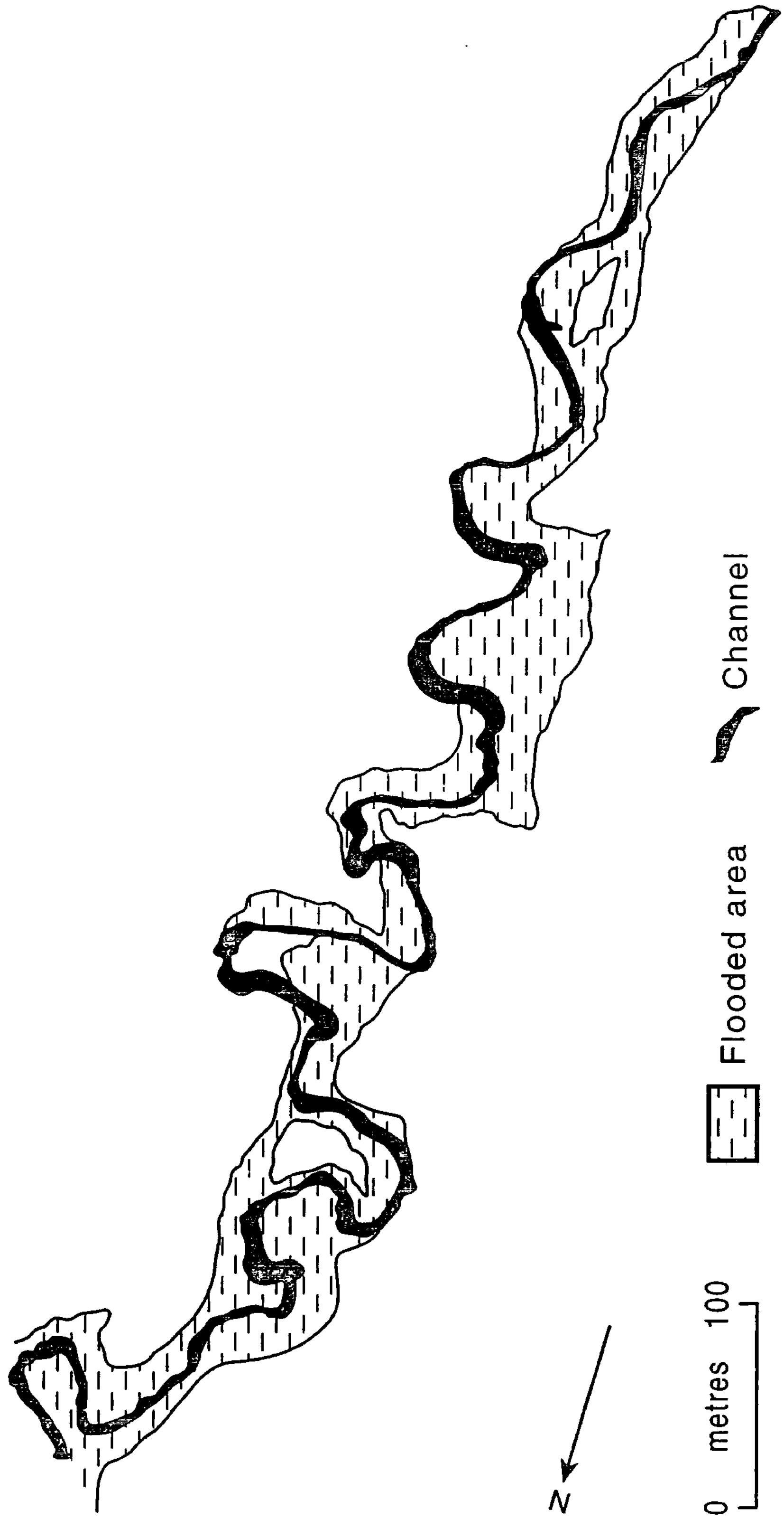


Figure 6.2 Map of extent of a flood of 19th and 20th February, 1997 at Swinhope Burn

confined by valley side slopes. For example, in the uppermost sections of the study reach a steep slope bounding the channel confined the overbank flow to within four metres of the channel. However, in the upper reaches, floodplain morphology is more open which allows overbank flow to cover a greater area. Most of the flooding occurred immediately downstream of narrow channel segments although some areas close to the channel remained unaffected by the flood.

Two days after the flood the channel remained at bankfull level (Figure 6.5). Palaeochannels traversing the floodplain which had been re-occupied during overbank flooding were still partly in use (Figure 6.6). Along the length of the study reach, where the pre-existing channel had become blocked with sediment, small chute cut-offs had occurred with the flow cutting across the floodplain and re-entering the channel a short distance downstream, for example between section 26 and section 35 as shown by the arrows on Figure 6.6. A further example of a temporary change in channel planform occurred between sections 92 and 100, which resulted in some of the flow being diverted overbank along the line of a potential meander bend cutoff (Figure 6.7). At section 35 a major left bank tributary was still contributing approximately half of the overbank flow to the channel two days after the flood (Figure 6.8).

The relatively fine nature of the bed material moved during the flood was evident in the extent of overbank splays, particularly in the upper reaches (Figure 6.4). On the day following the flood event, suspended sediment was observed to be still in transport. Erosion of fine bank material is the probable source of the fines observed during the immediate post-flood period. At some points along the channel, the banks were still saturated and had been severely undercut.

In some reaches there was evidence that both erosion and aggradation of the bed had occurred, with aggradation being especially pronounced in the upper reaches. There was evidence to suggest that flow had been overbank at all of the 101 cross-sections. Debris associated with the flood was found up to fifty metres from the channel (Figure 6.2). In the upper reaches, flow remained overbank for a considerable period of time after the flood event (Figure 6.9). Structural damage to walls and fences adjacent to the channel provides an indication of the floods power during the event.

Figure 6.3 Trashlines in the uppermost reaches of Swinhope Burn showing the extent of the flood of February 19th and 20th February, 1997

Figure 6.4 Overbank fines, flattened vegetation and standing water, providing evidence of flood limits in Swinhope Burn.

Figure 6.5 Flow remaining at bankfull level in upper reaches, 21st February, 1997

Figure 6.6 Palaeochannels in middle reaches of Swinhope Burn utilised by flood flows, 22nd February, 1997

Figure 6.7 Site of potential meander bend cut-off (sections 92-98)

Figure 6.8 A major left bank tributary (around section 35) still contributing half of the overbank flow to the channel, 22nd February, 1997

Figure 6.9 Overbank deposition of fines in the upper reach of Swinhope Burn, 22nd February, 1997.

6.5 Downstream variations in cross-sectional form and mean grain-size of bed material

A network of 101 cross-sections along the 1.4 km reach of Swinhope Burn has allowed downstream variations in cross-sectional form and mean grain-size to be related to the February, 1997 flood event (Figure 6.10). Cross-sections were initially surveyed on the 15th and 16th April, 1996 and re-surveyed following the flood (on the 19th and 20th February, 1997) during the period 7th March, 1997 to 18th March, 1997. In the intervening period weekly checks on the channel and some limited re-survey demonstrated between survey change to be negligible. The initial grain-size survey to identify the existing composition of the bed material and to examine the downstream distribution of mean grain-sizes was carried out between the 28th August, 1996 and 22nd September, 1996. Re-sampling after the flood took place during the period 25th March, 1997 to 6th April, 1997. Again weekly checks established very little change in the intervening periods.

Figure 6.10 Location map showing the distribution of selected cross-sections along the study reach of Swinhope Burn (Air photograph courtesy of Aerofilms and Durham County Council)

This section identifies downstream changes in bankfull width, mean bankfull depth and width:depth ratio following the flood. Downstream variations in erosion and deposition for the bed and banks at each cross-section are calculated using measurements in m^2. Finally, downstream variations in mean-grain size resulting from the flood event are described.

6.5.1 Downstream variations in width:depth ratio in Swinhope Burn

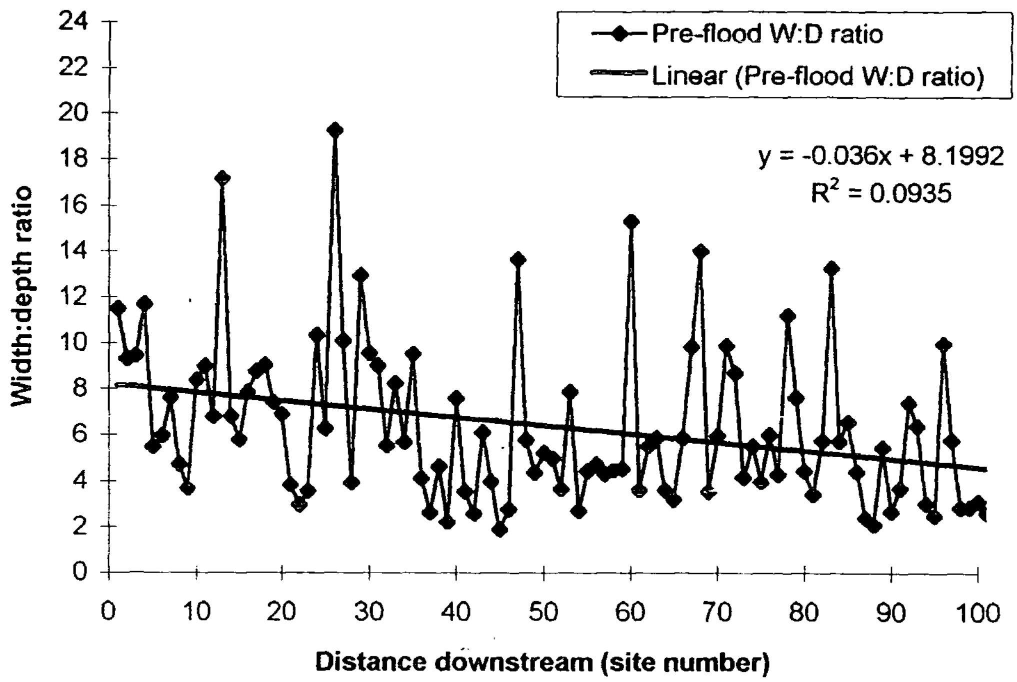
Figure 6.11 shows downstream changes in the pre and post flood width:depth ratio at the 101 stream cross-sections. Width:depth ratio decreases downstream for both the pre and post flood channel. This illustrates the tendency for the study reach to be wider and shallower in the upper reaches and narrower and deeper in the lower reaches, as identified in Chapter 5.

In the upper reaches, bank material is coarser, consisting of gravel and cobbles in a fine silt matrix which encourages lateral adjustment of the channel through bank erosion rather than vertical adjustment through bed erosion (Milne, 1983a). During high flows, once the gravel and cobbles become dislodged, for example through bank undercutting (Thorne and Tovey, 1981), the remaining fine alluvial deposits above the coarser material become unstable and collapse. In the lower reaches the banks are more cohesive consisting primarily of silt clay which discourages lateral bank retreat but encourages vertical erosion of the bed resulting in a much narrower and deeper channel.

The linear trendlines fitted to the two data series (Figure 6.11) have low negative slopes which indicates a small decrease in width:depth ratio downstream. The similarity in the regression equations and pattern in the data suggests very little change as a consequence of the flood.

Figure 6.12 shows downstream net differences between the pre-flood and post-flood width:depth ratio. A positive value shows an increase in width:depth ratio from the pre-flood to the post-flood survey whereas a negative value shows the opposite. The linear regression line has a very low slope and shows no real downstream trend.

A



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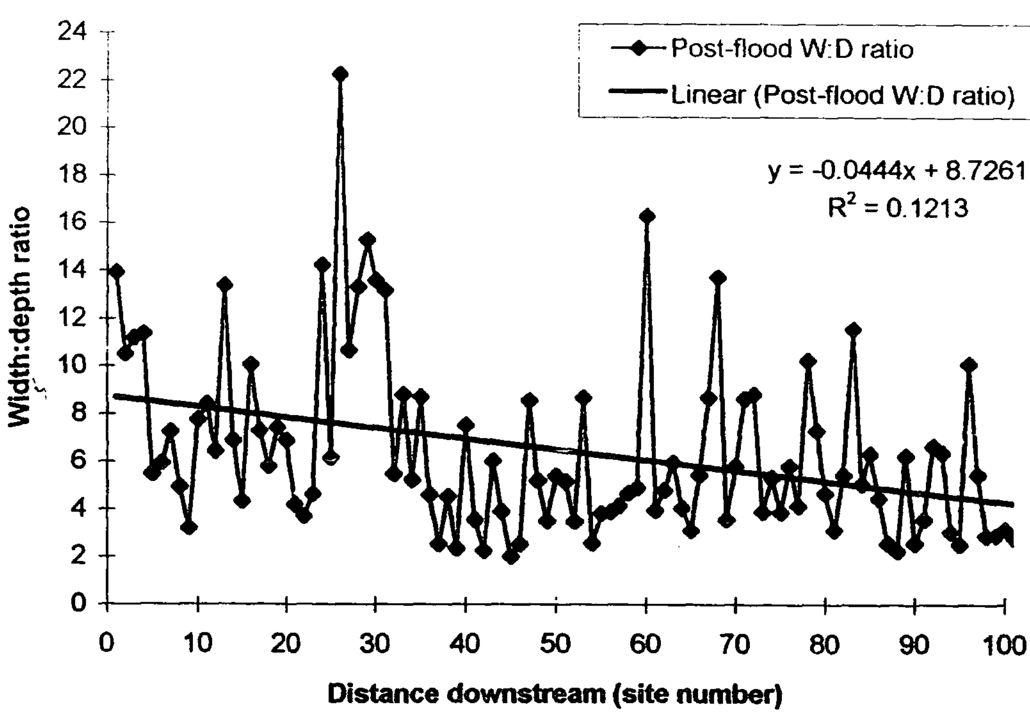


Figure 6.11 Downstream changes in pre (a) and post flood (b) width:depth ratio

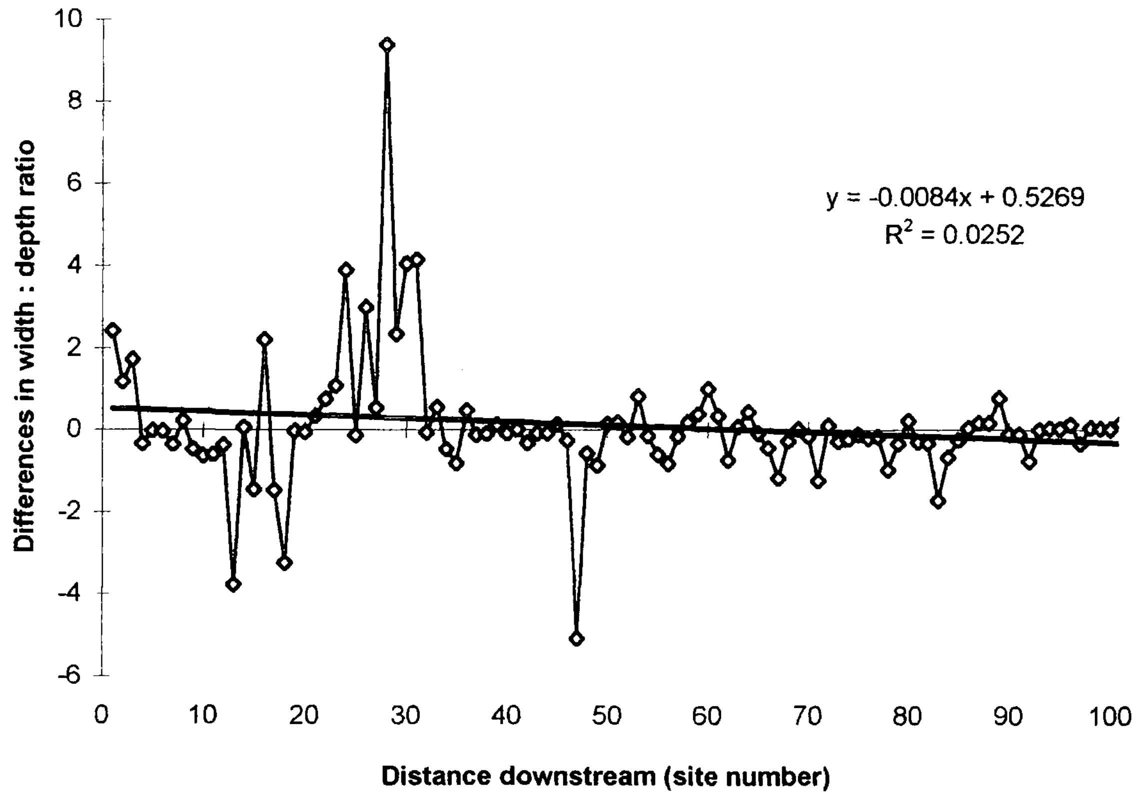


Figure 6.12 Differences between pre-flood and post flood width:depth ratio

At the majority of cross-sections and particularly downstream of section 32 the change in the width:depth ratio is very small. However, upstream of section 32 there is considerable local variation and an increase in width:depth ratio is evident. Some cross-sections in the upper reaches have undergone a substantial amount of change with the channel initially becoming narrower and deeper around sections 13 to 18 and then progressively wider and shallower from sections 20 to 32 (Figure 6.13).

It is likely that this increase in width:depth ratio in the upper reaches (sections 24 -31) has resulted from a combination of bank erosion and bed aggradation. Plots showing the differences between pre-flood and post-flood bankfull width (Figure 6.14a) and depth (Figure 6.14b) confirm that between sections 20 and 31, bed aggradation occurred and between sections 25 and 31 bank erosion had occurred which corresponds with the increase in width:depth ratio shown between sections 24 and 31 on Figure 6.12.

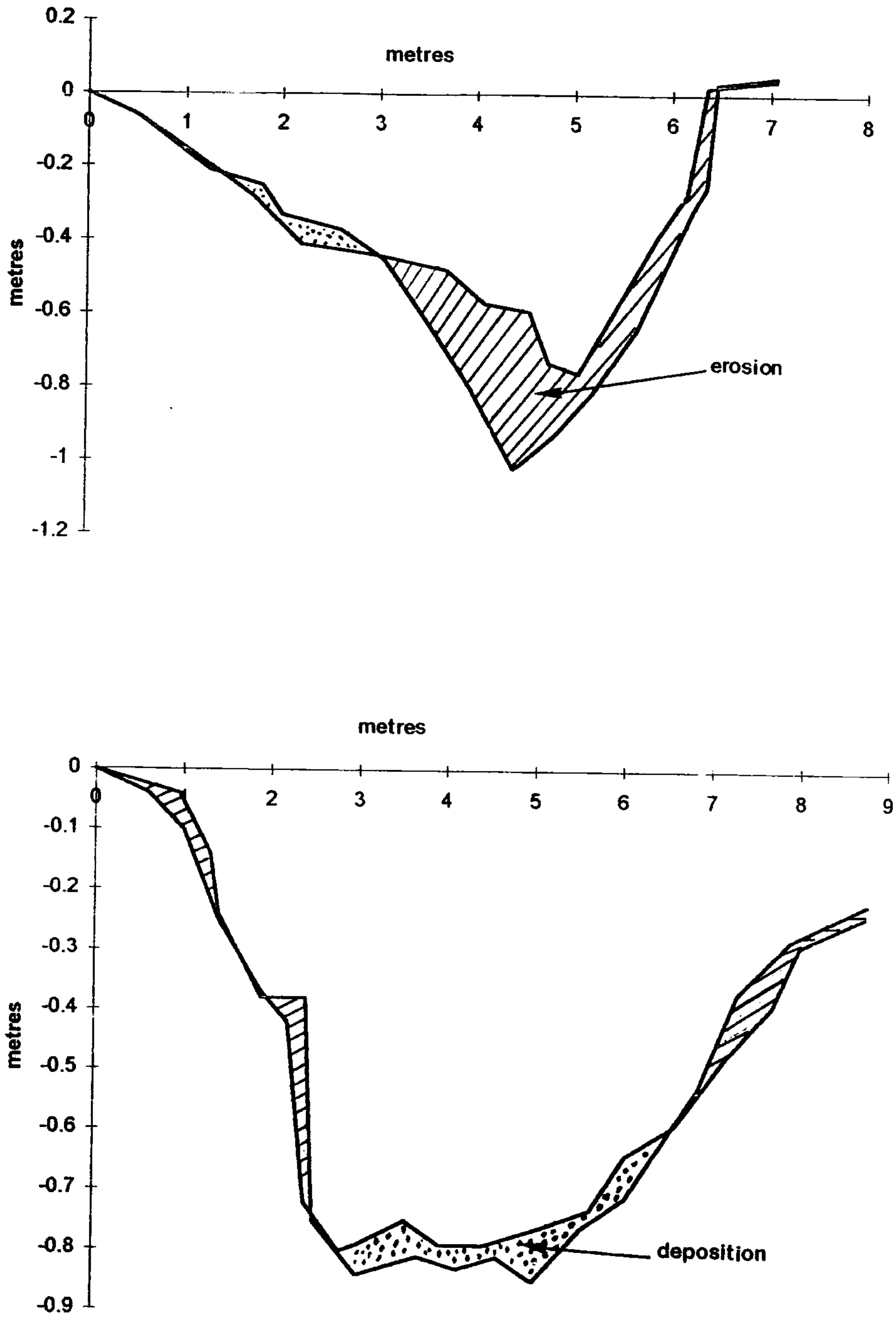


Figure 6.13 Cross-sectional changes at Swinhope Burn, showing channel erosion and deposition (top, section 18, bottom, section 24)

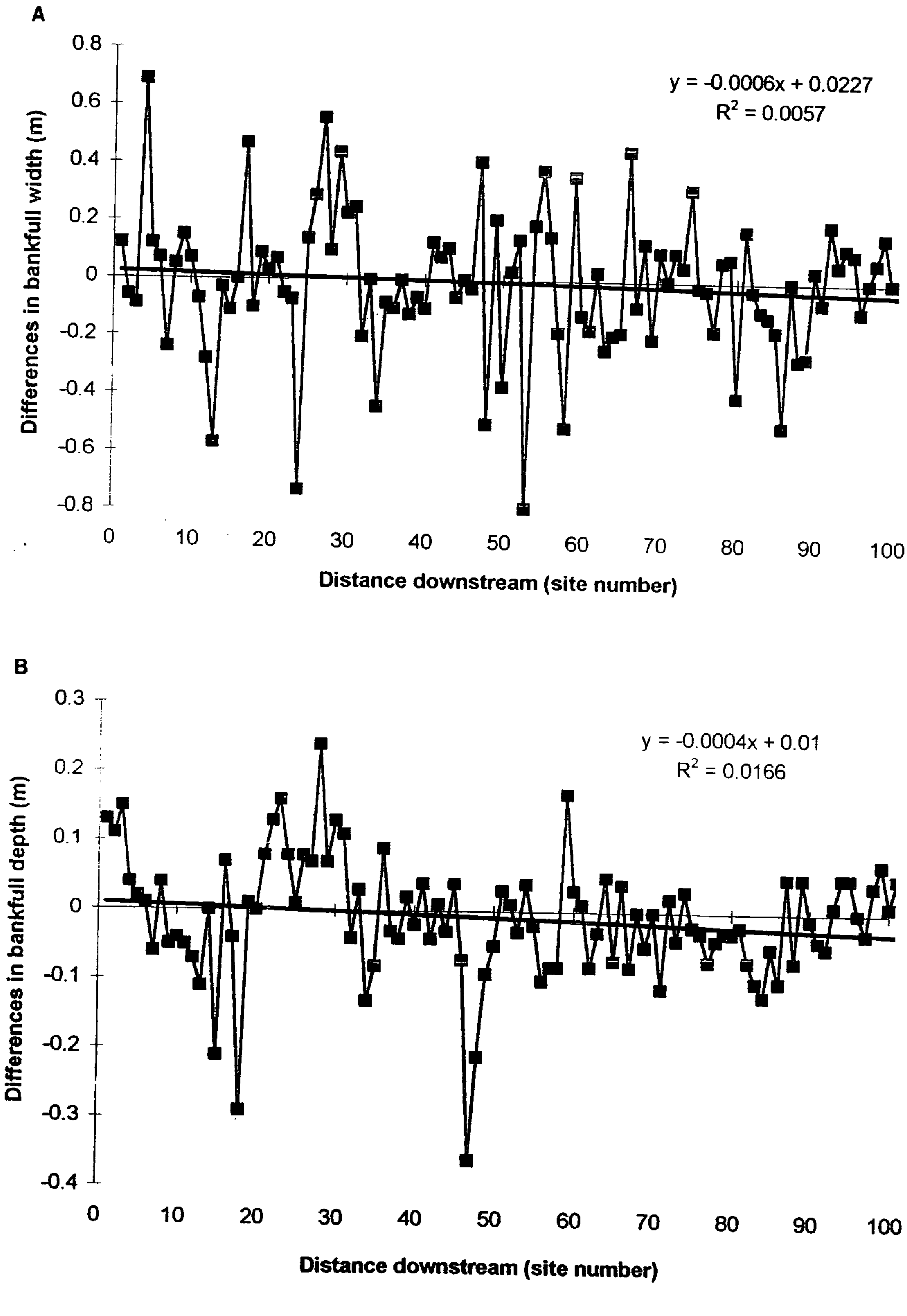


Figure 6.14 Differences in bankfull width (a) and depth (b) between pre-flood and post-flood channel.

The meander bend in question is migrating and since the channel is quite shallow around this bend overbank flow occurs relatively frequently and bank erosion is active. A further explanation for the increase in bed aggradation on Figure 6.14 from sections 20 to 31 is the recorded bed erosion immediately upstream, from sections 13 to 18. Within these sections the channel had become narrower and deeper following the flood. It is possible that erosion of the bed provided the source of sediment which was subsequently deposited on the next meander bend downstream leading to bed aggradation.

The greatest increase in bankfull depth from the pre-flood to the post flood survey occurred either in pools on straight reaches or in deep pools on meander bends (Figure 6.14b) which supports the velocity-reversal hypothesis that during high flows pools are actively scoured with deposition occurring on downstream riffles. This indicates that the deepest parts of the channel are becoming deeper and the shallower parts are becoming shallower, thus increasing the tendency for a downstream decrease in width:depth ratio. The trendline on Figure 6.12 indicates a slight increase following the flood for the width:depth ratio to decrease downstream. This may be the result of increased bed aggradation in the upper reaches and bed erosion in the lower reaches due to the possible passage of a sediment wave through the study reach during the flood.

6.5.2 Downstream variations in channel erosion and deposition

Figure 6.15 shows bank erosion and bank accretion recorded at each of the cross-sections. In general there is little overall change in the stream channel and bank erosion is of the same order of magnitude as bank accretion. However, there is a slight downstream decrease in the total area of both bank erosion and bank accretion.

Peaks in bank accretion occur at three sections at the head of the study reach, all of which are riffle locations. Further peaks in bank accretion occur intermittently downstream, but particularly on meander bends, for example, sites 26, 27, 53, 71 and 80. In the upper reaches, for example, sites 26 and 27, bank accretion has probably resulted from the deposition of fine sediment on the inside of these meander bends since overbank flooding was significant in these reaches. In the lower reaches, for

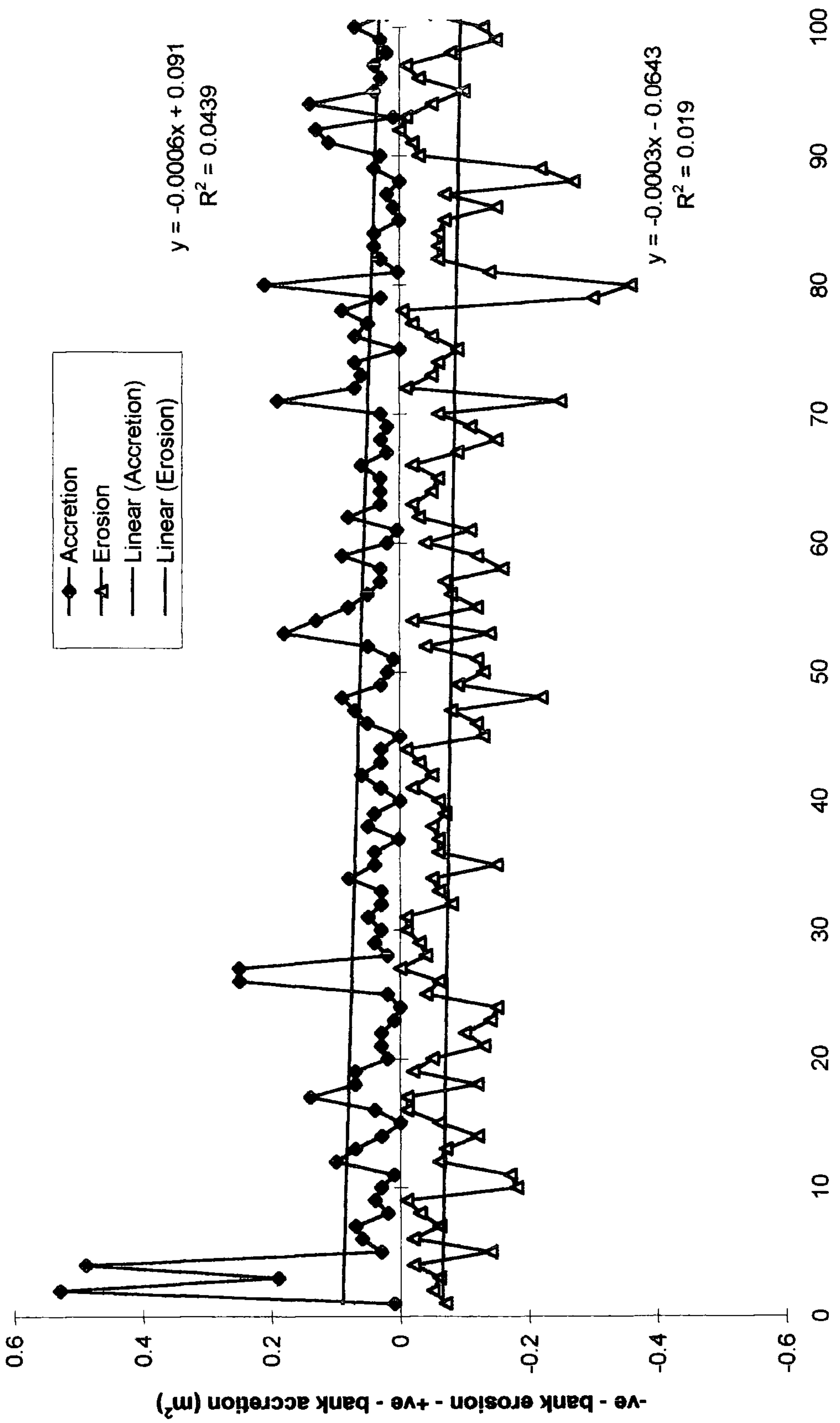


Figure 6.15 Downstream changes in total area bank erosion and bank accretion

example, sites 53, 71 and 80 recorded bank accretion probably represents deposition of fine sediment on point bars opposite meander bend apices, where the bends are actively migrating. For example, at sites 53, 71 and 80, there are corresponding peaks in both bank accretion and bank erosion, suggesting that at bend locations the erosion of meander bend apices supplies fine sediment which is transported to the opposite point bar through secondary circulations during flood events (Milne, 1983a). Other intermittent peaks in bank erosion may have resulted from localised bank instability and collapse.

Figure 6.16 shows the area of bed erosion and bed aggradation recorded at each of the cross-sections. As in Figure 6.15, in general there is very little change in the area of the bed along the study reach. A downstream trend in bed erosion is difficult to identify as the trendline is strongly influenced by peakiness of the plot. However, there is a clear downstream decrease in bed aggradation.

Figure 6.16 is dominated by a number of large peaks in both bed erosion and bed aggradation. Peaks in bed aggradation between sites 25 to 35 are due to the high volume of sediment from upstream which was also deposited on the channel banks as identified in Figure 6.15. A peak of bed aggradation at site 27 occurs immediately downstream of a wide meander bend. Field observations in the immediate post-flood period revealed that this straight section of channel had aggraded almost up to the level of the floodplain. A peak at site 59 is also related to deposition on a meander bend. However, it should be noted that peaks in bed erosion relate to relatively small adjustments in cross-sectional form. For example, a peak in bed erosion at site 47 is due to small bank collapse. Another peak in bed erosion is recorded on a meander bend at site 80 but is offset by a peak in bed aggradation at the same site.

Figure 6.17 shows the total area of channel erosion and deposition for each of the cross-sections. The graph shows a downstream decrease in the area of channel deposition and a slight downstream increase in channel erosion. As the two trendlines are overlain, a pattern of channel deposition in the upper reaches and channel erosion in the lower reaches can be identified. The linear regression lines cross around section 49. Similarly, plots of downstream variations in the net change in bank and bed area show

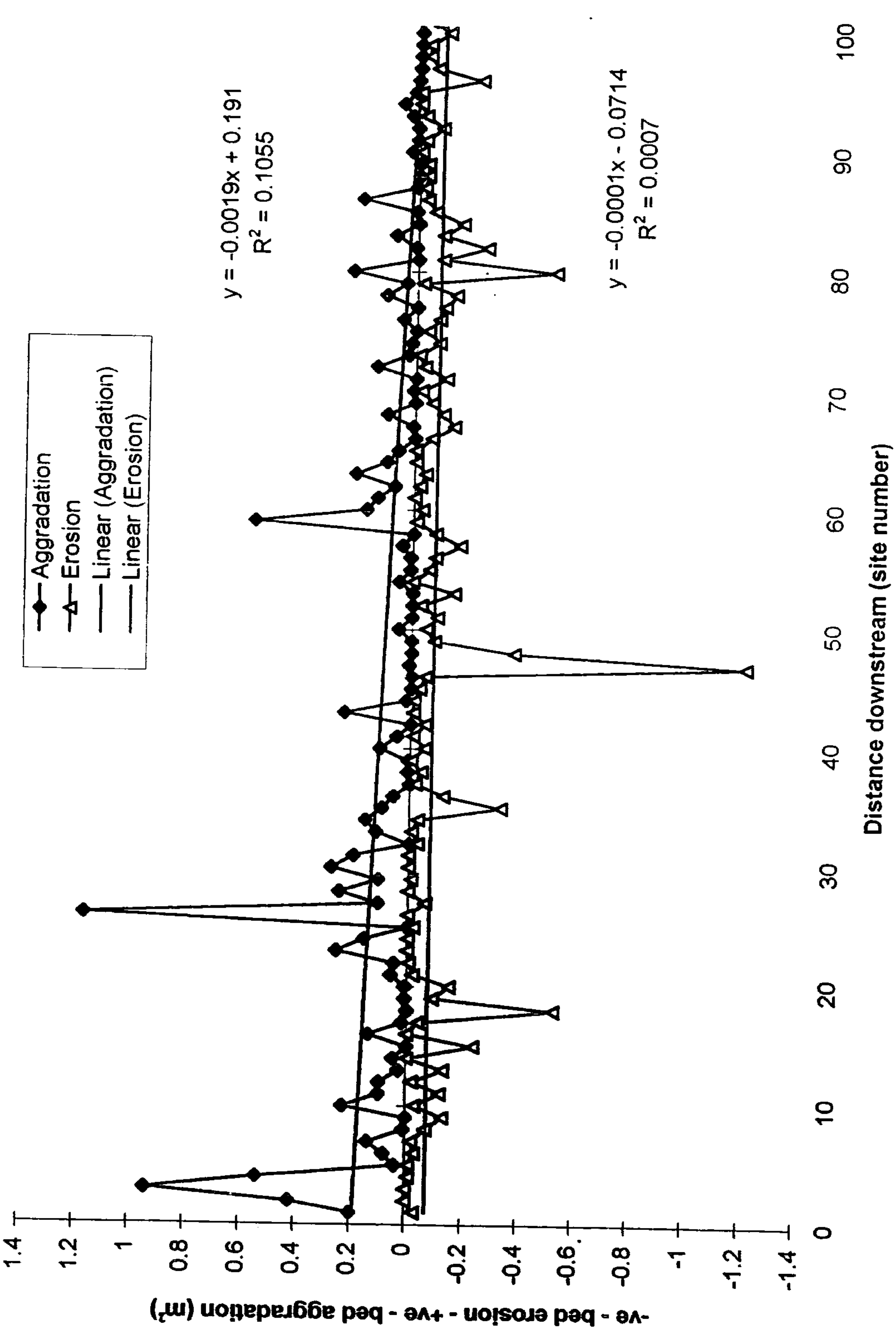


Figure 6.16 Downstream changes in total area bed erosion and bed aggradation

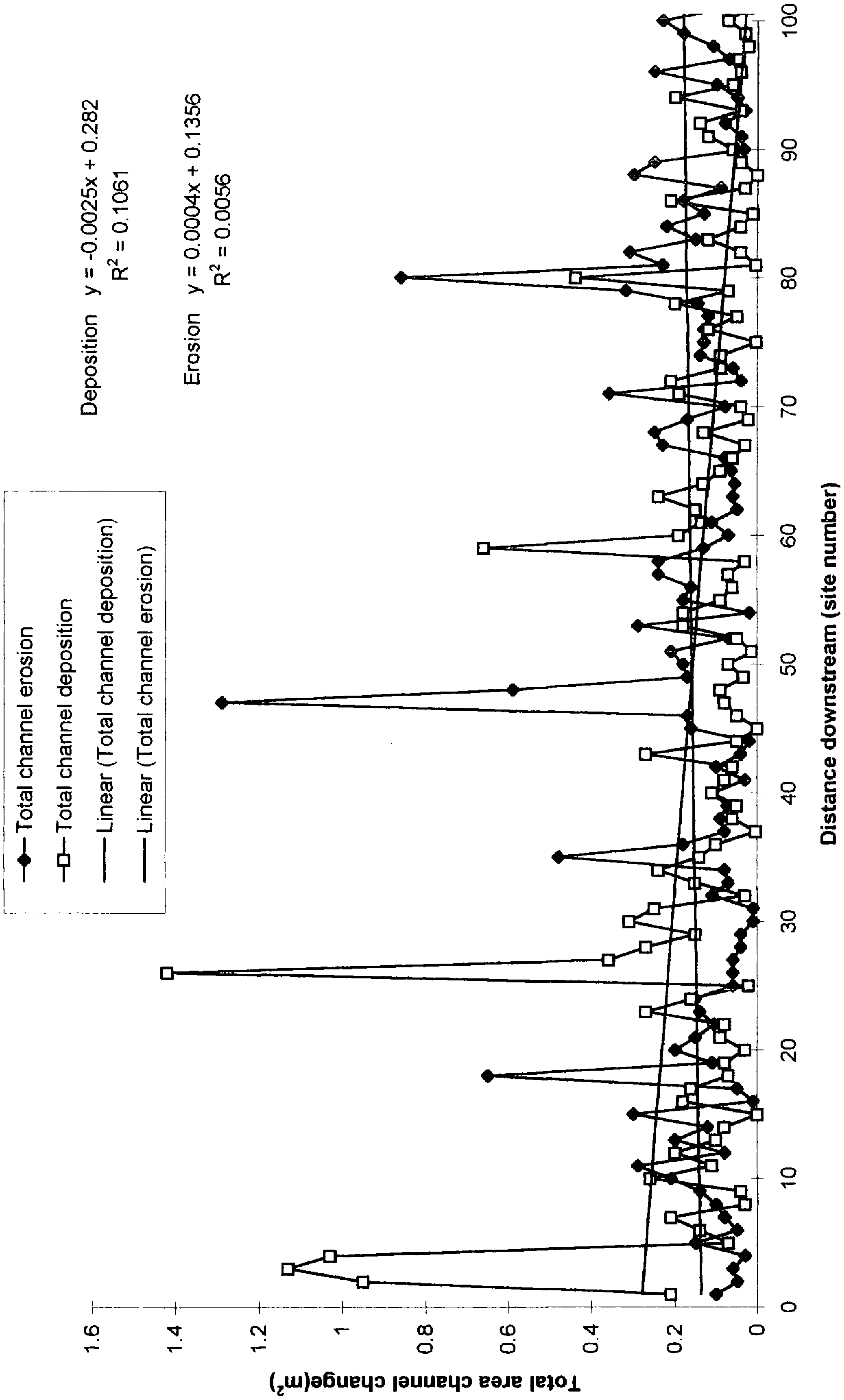


Figure 6.17 Downstream variations in total area channel erosion and deposition

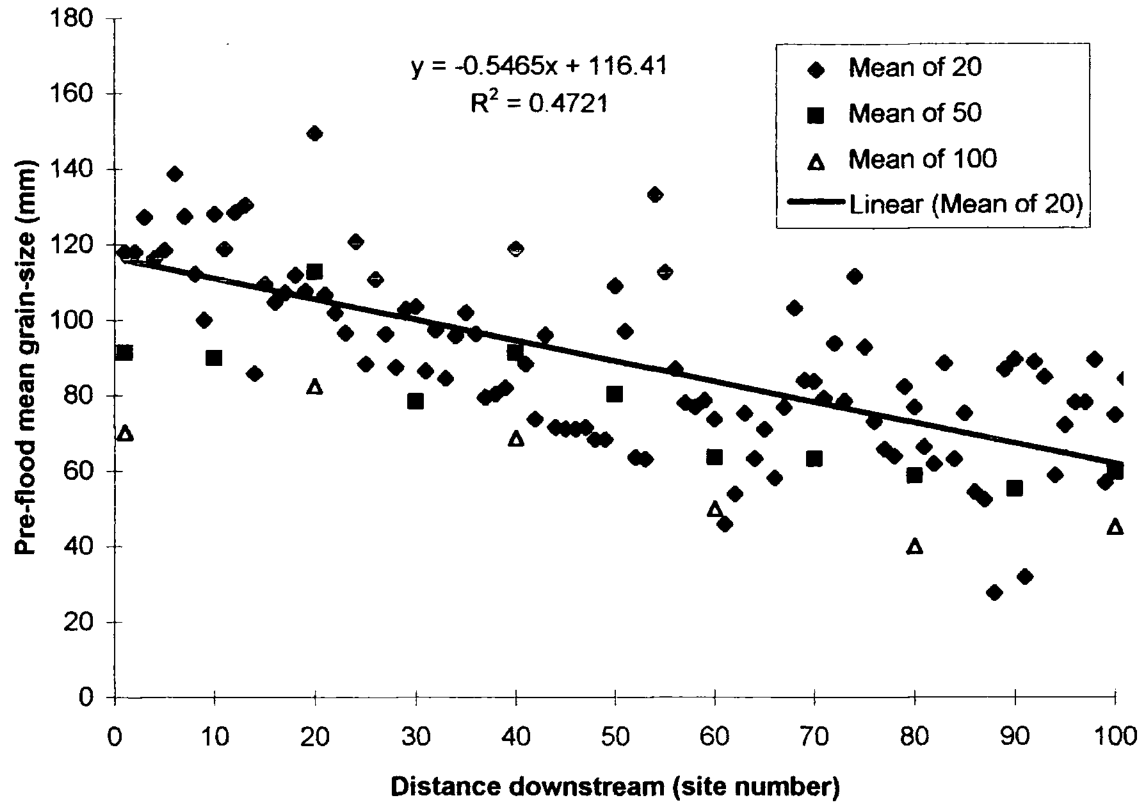
the importance of channel aggradation in the upper reaches and channel erosion in the lower reaches. However, the net effect of these changes on cross-sectional form is relatively small.

Nevertheless, this downstream pattern of erosion and deposition suggests the passage of a wave of sediment into the study reach, which probably stopped abruptly around section 49. Aggradation of the bed would have occurred in the upper reaches, but once the sediment wave had halted, the flood is likely to have eroded bed and banks downstream. Field evidence of aggradation in the upper reaches is provided by change in pre and post flood cross-sections and by quantities of overbank fines covering the floodplain adjacent to the channel margins.

6.5.3 Downstream variations in mean grain-size distribution

Figure 6.18 shows downstream changes in the mean grain-size distribution from the pre-flood to the post-flood survey. Both plots show a downstream fining of mean grain-size. When the pre and post flood mean grain-sizes are overlain, the inflection point of the two linear regression lines at section 40 indicates that upstream there has been a decrease in mean-grain size and downstream there has been an increase in mean-grain size. Figure 6.19, showing post-flood minus pre-flood mean grain-size confirms that grain-sizes sampled in the repeat survey were finer in the upper reaches and coarser in the lower reaches. The moving average shows an almost cyclical pattern corresponding to the meandering pattern of the stream where coarsening of the bedload occurs immediately downstream of bluffs (site 13, 22 and 54). The inflection point of trendlines around section 40 (Figure 6.18) approximately corresponds with the inflection point at section 49 shown on Figure 6.17 which shows a clear pattern of channel aggradation in the upper reaches and channel erosion in the lower reaches.

One possible explanation for the link between these inflection points is that aggradation within the channel in the upper reaches was caused by large inputs of fines from upstream bank erosion being introduced to the channel which effectively 'fined' the pre-existing coarser bed material. This would explain the decrease in mean-grain size of bed material in the upper reaches following the flood. An increase in mean-grain size in the lower reaches may have resulted from the removal of fines stored within the



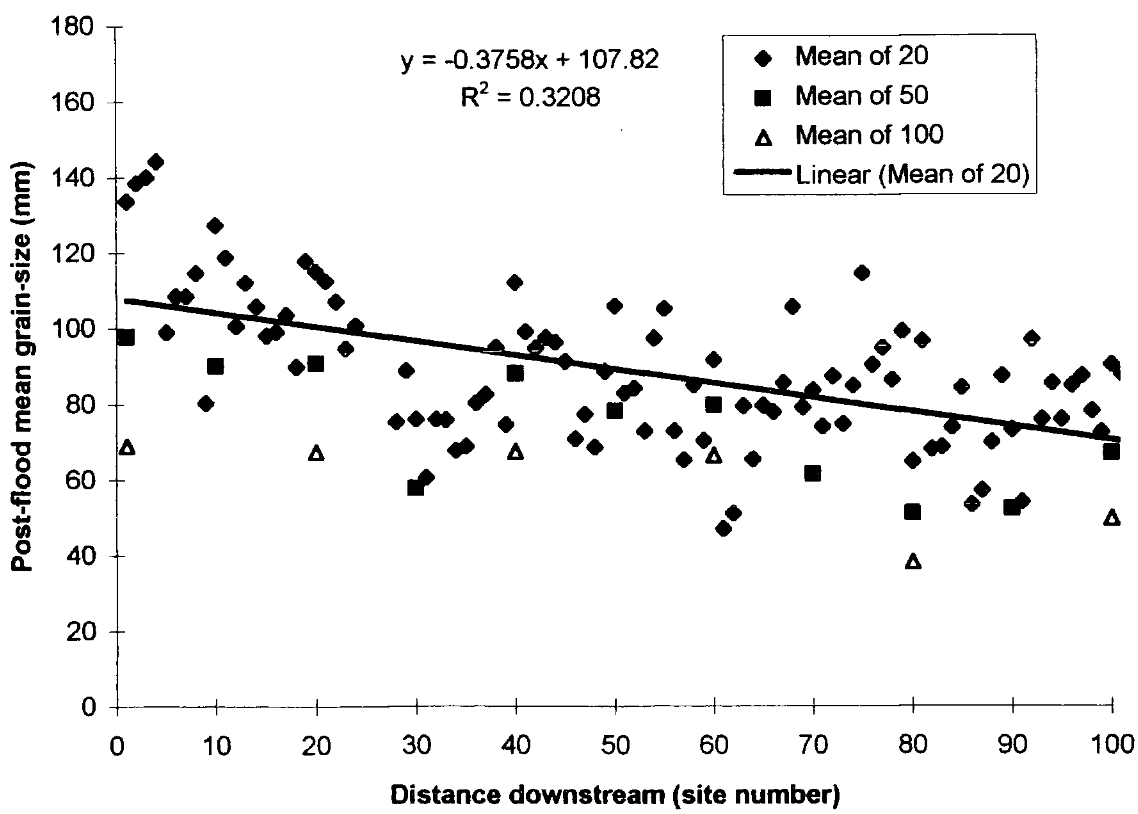


Figure 6.18 Mean grain-size of the 20 largest surface particles, measured at the 101 cross-sections, before and after the February, 1997 flood

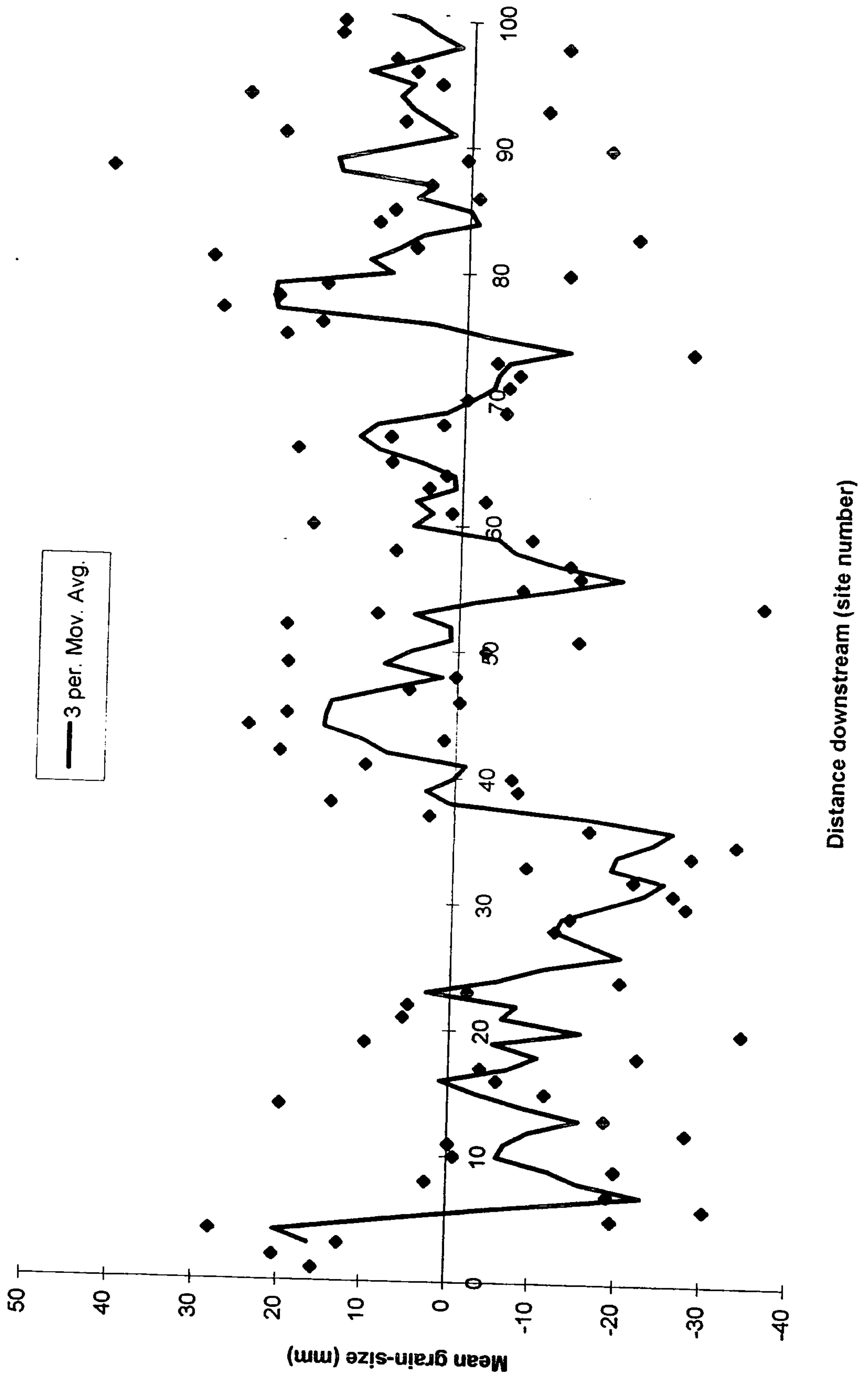


Figure 6.19 Post-flood mean grain-size minus pre-flood mean grain-size

bed, thus leaving the bed surface marginally coarser than it had been before the flood. Alternatively a proportion of the new load would travel through the channel.

However, it must be stressed that the recorded changes are very small and represent minor adjustments to the channel as a whole. Although both before and after the flood downstream fining is apparent, changes in the mean-grain size of bed material are small. Statistical comparison of the regression line slopes shown in Figure 6.18 shows that there is no significant difference between mean grain-sizes sampled before and after the flood of February 1997 at the 95% confidence interval.

Figure 6.20a shows cumulative erosion and cumulative deposition along the length of the study reach. Although channel deposition is largely responsible for changes in cross-sectional morphology in the upper reaches, channel erosion tends to dominate in the lower reaches. However, plots of cumulative channel erosion and deposition converge towards the end of the study reach, which indicates that the changes along the channel as a whole tend to balance out.

Figure 6.20b and 6.20c show the area of cumulative erosion and deposition for bed and banks respectively. Downstream changes in the banks tend to occur gradually, whereas downstream changes in the bed tend to occur more abruptly producing small steps within the cumulative profile. Steps in the plot of cumulative bed aggradation is mirrored by a steps in the plot of cumulative bed erosion further downstream. The net result is a greater amount of bed aggradation than bed erosion along the study reach. Figure 6.20c clearly shows the importance of bank accretion in the lower reaches and bank erosion in the upper reaches with the point of interception being around site 50.

In summary, very little downstream change in cross-sectional morphology or mean grain-size of the bed material has been recorded. Small adjustments in channel form usually associated with migrating meander bends are observable but on the whole the channel is highly stable even during a major overbank flow event.

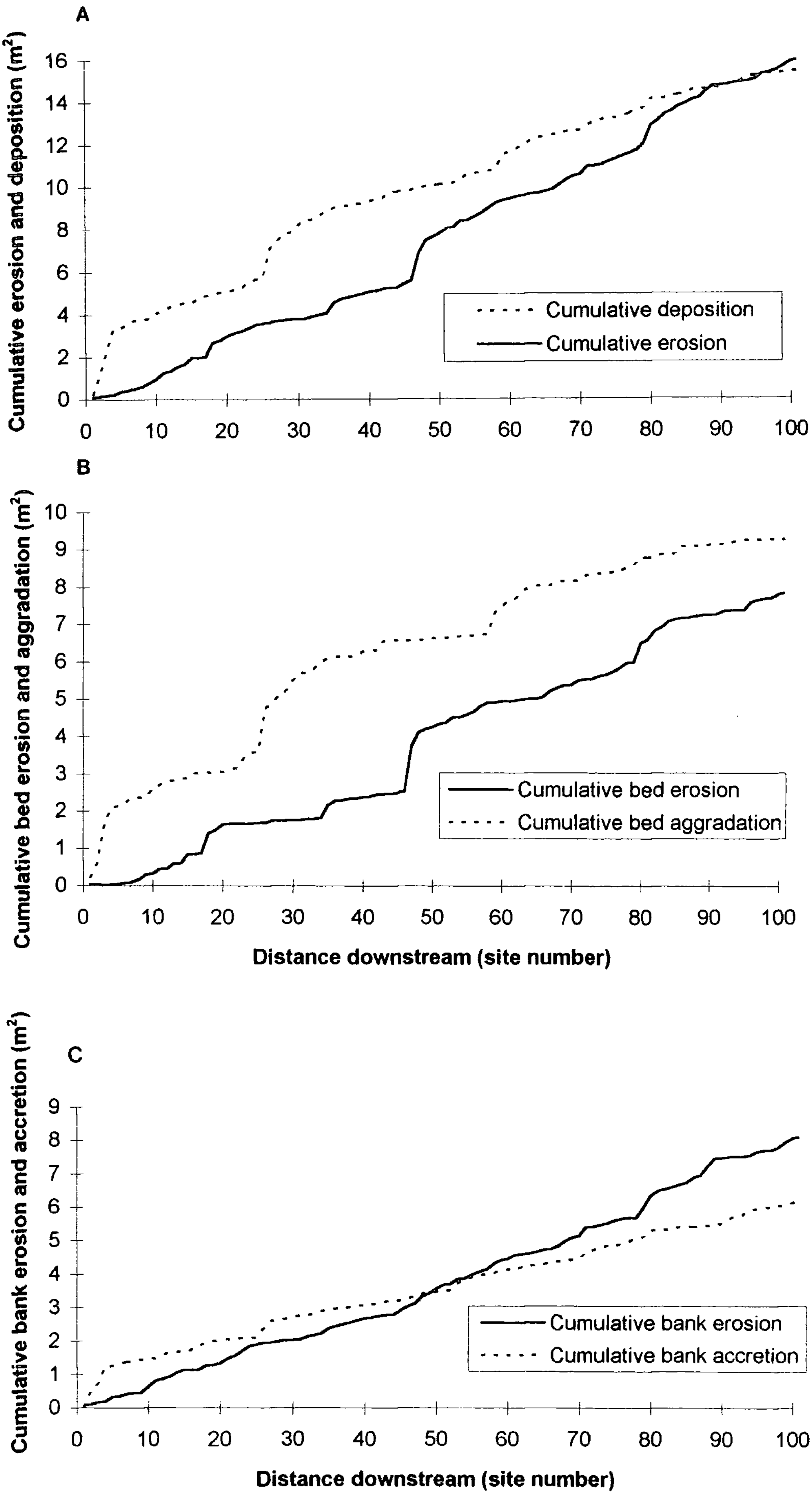


Figure 6.20 Cumulative channel erosion and deposition along the study reach

6.6 Correlation of channel variables

In order to examine the relationship between variables used to describe and analyse changes in cross-sectional form resulting from the passage of the flood, a correlation matrix helped to identify the relative importance of the processes of erosion and deposition on changes in the bed and banks within the study reach. Examination of the relationships between variables suggested several causal linkages.

Correlation between the total area of channel erosion and both bed erosion and bank erosion (Figure 6.21) revealed that channel erosion had occurred more as a result of erosion of the channel bed rather than the banks. Similarly, correlation between the total area of channel deposition and both bed aggradation and bank accretion (Figure 6.22) revealed that channel deposition had occurred more as a result of deposition on the channel bed rather than the banks. However changes identified on both Figure 6.21 and Figure 6.22 are relatively small, with some very strong outliers.

The lack of lateral change in channel pattern observed over the historical period is substantiated by evidence that the contemporary response of the channel to a major flood event is through vertical adjustments within the channel bed rather than lateral adjustment through the migration of meander bends.

Summary

Observations of downstream variations in cross-sectional form of the channel have identified that the overbank flood of 19th and 20th February, 1997 produced very little channel change. This is shown by both the morphological data and grain-size data. Plots showing cumulative erosion and deposition along the length of the channel have been shown to converge towards the end of the study reach indicating the overall lack of change over the reach. This is consistent with findings from other unconfined alluvial channels, where bank erosion has been shown to complement point bar deposition on the inside of meander bends with erosion matching deposition (Ferguson, 1981). Change at individual cross-sections is often highly erratic and localised.

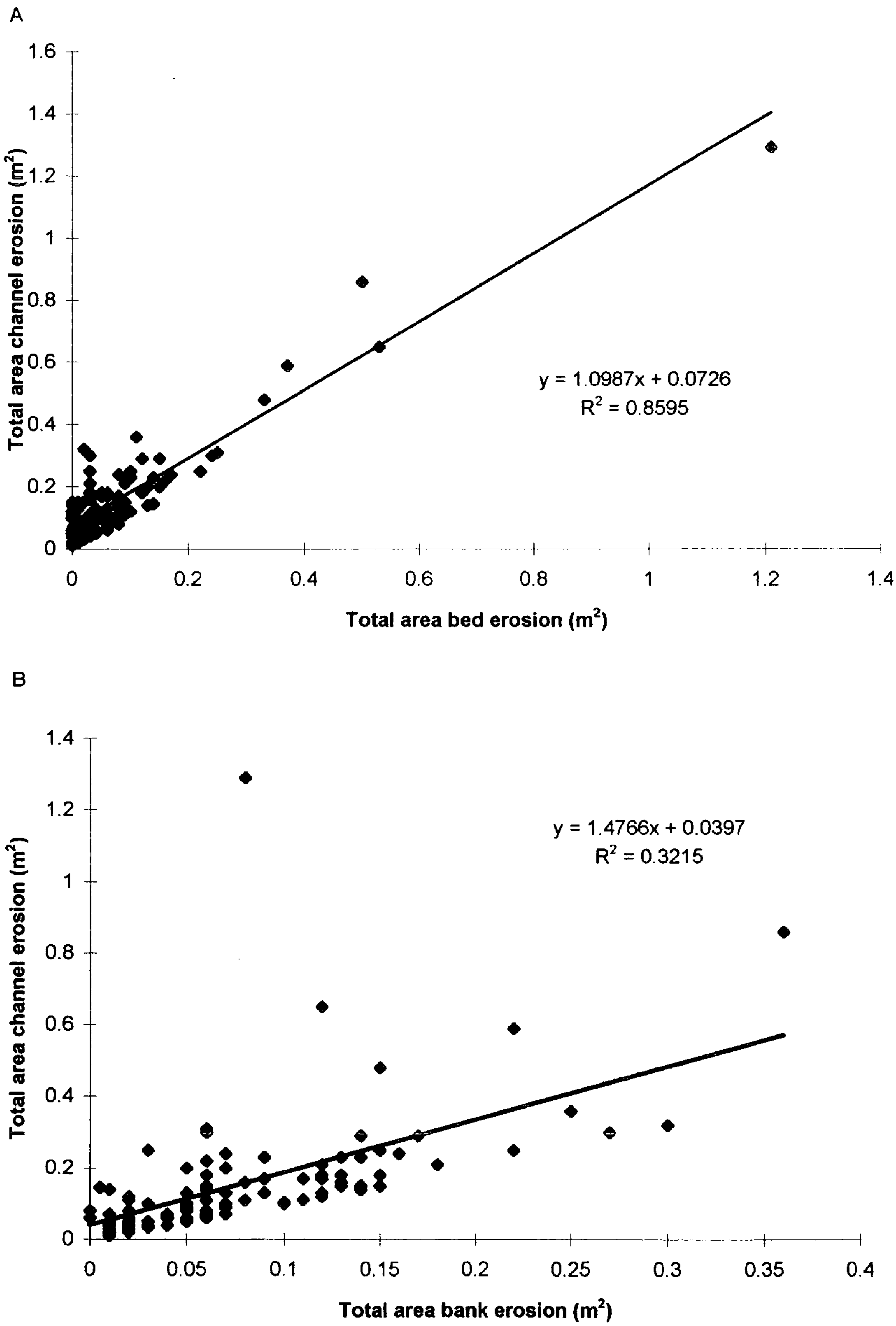
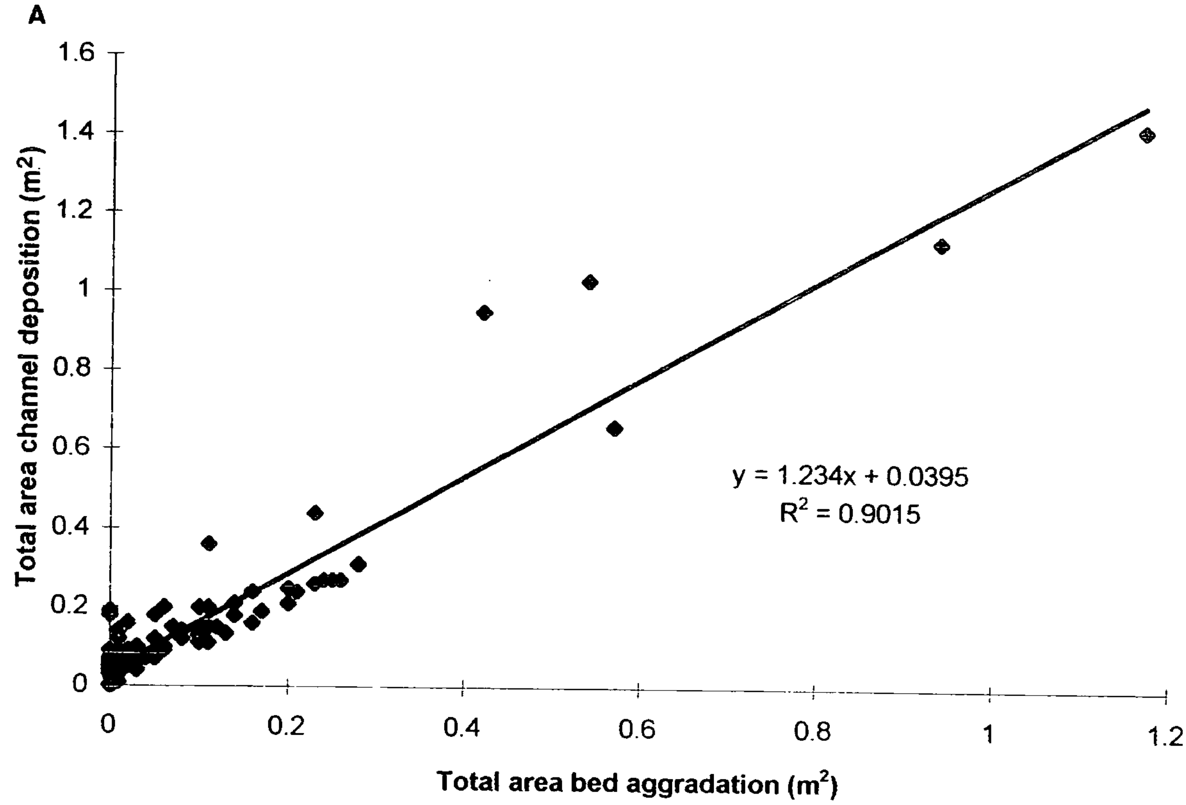


Figure 6.21 Total area of bed (a) and bank (b) erosion in relation to total channel erosion



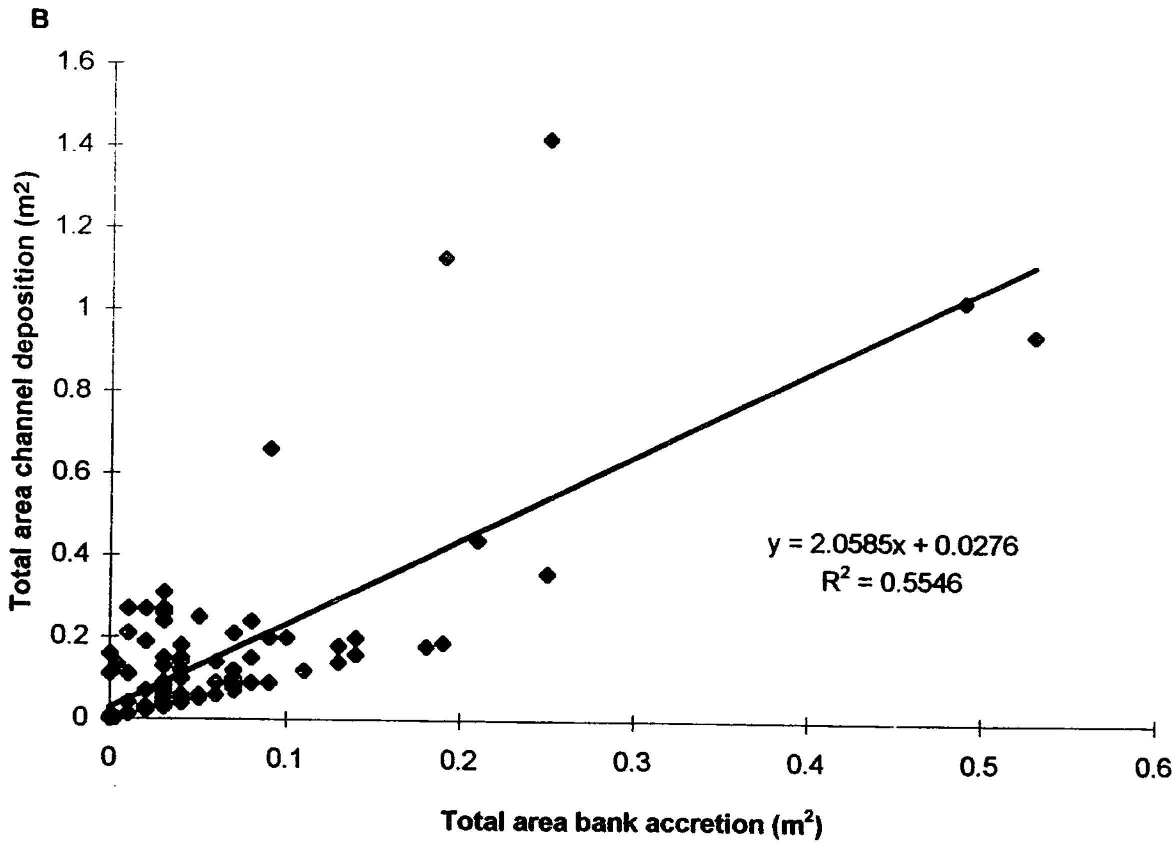


Figure 6.22 Total area of bed (a) and bank (b) deposition in relation to total channel deposition

Table 6.1 indicates that following the flood, mean values of bankfull width and bankfull depth have been practically unaltered by the flood event and channel erosion is of the same order of magnitude as channel deposition. Table 6.1 also shows that pre and post flood mean grain-size are virtually identical. The standard deviations of pre and post-flood bankfull width are identical, and there is only a very slight difference between the standard deviations of pre and post flood bankfull depth and mean grain-sizes.

	Pre-flood Mean	Stand. Dev.	Post-flood mean	Stand. Dev.
Bankfull width (metres)	4.05	2.04	4.06	2.04
Bankfull depth (metres)	0.67	0.19	0.68	0.21
Grain-size (mm)	88.5	23.3	88.4	19.5
Channel erosion (m^2)	-	-	0.16	0.17
Channel deposition (m^2)	-	-	0.15	0.23

Table 6.1 Mean channel characteristics before and after 19th and 20th February flood

It appears that the changes which did occur within the channel resulted predominantly from changes within the bed rather than the banks. However, analysis of downstream variations in channel change have revealed that channel aggradation was more prominent in the upper reaches, whereas channel erosion was greater in the lower reaches. A possible explanation for this has been linked to the passage of a sediment wave downstream which may have halted approximately mid-way along the study reach.

It is very clear that downstream variations in cross-sectional response to flood events occurs in relation to highly localised factors. These include bed topography, in terms of the position of pools, riffles and meander bends; local bank cohesion and stability; channel slope and channel planform. In the following section, the influence of each of these four variables on the response of channel morphology and mean grain-size to a major flood event will be assessed.

6.7 The role of pre-existing bed topography (pools, riffles and meander bends) on changes in cross-sectional form and mean-grain size distribution following a flood event

This section identifies the nature of channel change resulting from the flood of February, 1997, in relation to the bed topography of Swinhope Burn.

Flood induced change in cross-section - the pool, riffle and meander bend sequence

During a high flow event the majority of channel erosion tends to occur on bends and the amount of channel erosion at these locations tends to increase downstream (Figure 6.23). However, it is difficult to generalise because only a small number of bend sites are included in this study (approximately 12% of all sites are located at bends). In downstream reaches, pools on meander bends are deeper than those in the upper reaches. This illustrates the important influence of bank cohesiveness on vertical scouring of pools during high flow events. Cohesive bank sediments ensure that the concave bank maintains a steep angle while retreating, and the majority of erosion is vertical, increasing the depth of the pool rather than the width (Milne, 1982b). On straight reaches, at pool locations there was slightly more erosion than at riffle locations. There is very little downstream variation in the total area of channel erosion for both pools and riffles although all sections show some change.

As mentioned earlier, pools situated at meander bends tend to promote unidirectional bank erosion, which is concentrated at the bend apex, particularly where pools are tightly curved. It has been suggested that during high flows, pools situated at meander bends are scoured increasing the depth of the pool (Milne, 1982b), but during low flows the pool is infilled. Knighton (1998) suggests that since velocity and slope are greater and depth is less over a riffle than in a pool at low flows, the net effect of these differences is to produce a more even distribution of flow over pools and riffles at high flows.

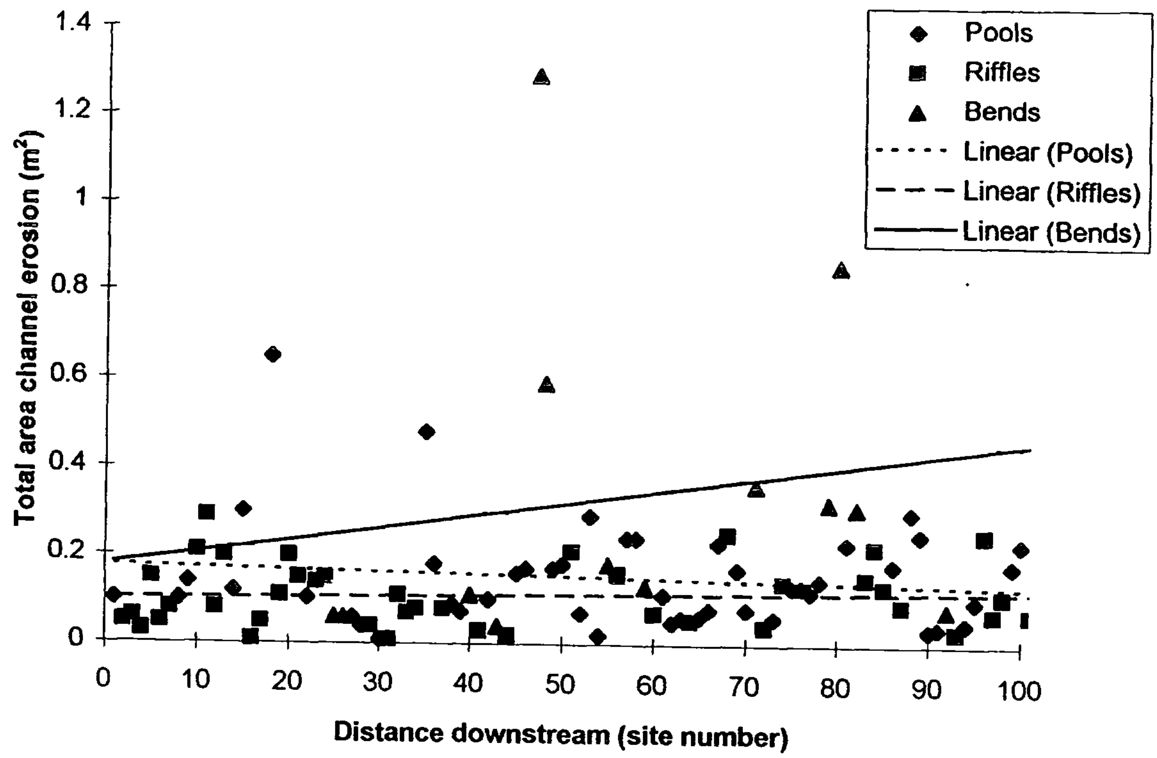


Figure 6.23 Downstream changes in the total area of channel erosion in relation to the pool-riffle, bend sequence

The velocity reversal hypothesis initially suggested by Keller (1971) argues that during high flows, competence is likely to be higher in pools than in riffles. Although the velocity reversal hypothesis is not accepted by some authors as a means by which competence is higher in pools than over riffles during high flow events, it is argued that the more open structure of bed material in pools produces variations in competence in pools and riffles (Clifford, 1993; Sear, 1996) and can lead to an increase in erosion of pool beds (Knighton, 1998). This may explain why higher levels of total channel erosion are plotted for meander bends and pools in straight reaches, whereas the least amount of channel erosion is plotted for riffles. However, it must be noted that there is very little difference in channel erosion between riffles and pools in straight reaches, particularly in the lower reaches.

Carling (1991), in his appraisal of the velocity-reversal hypothesis for pool- riffle sequences on the River Severn, suggests that for this 'reversal' to occur, riffles need to be significantly wider than pools during high flows in order to accommodate lower velocities. Since bank strength on the River Severn prevents the widening of riffles compared with pools there was no evidence for a velocity reversal at within-bank flows. Since the banks are very cohesive in the lower reaches of Swinhope Burn, it may be that the process of 'reversal' of velocity may be impeded. This may explain a slight decrease in the total amount of channel erosion in pools in the lower reaches.

The observation that the least amount of channel erosion occurs at riffles is supported by Lisle (1979) who found that velocity reversal occurred at 50 - 90% bankfull, at which point the pools cease scouring and begin to fill. However, for flows up to twice bankfull, he found that d_{50} particles were not entrained, thus inhibiting bed erosion.

Figure 6.24 shows that during a high flow event, the majority of channel deposition occurred on meander bends and, in contrast to Figure 6.23, decreases downstream. However, for pools on straight reaches, total channel deposition per unit width is fairly constant downstream. On straight reaches, upstream of section 40 there is more channel deposition per unit width on riffles and less in the pools with the converse being the case downstream of section 40. When compared with Figure 6.23, Figure 6.24 shows more deposition on riffles, concentrated upstream of section 41. Knighton (1998) in accordance with the velocity reversal hypothesis, suggests competence is either more evenly distributed or, in some cases, lower on riffles than in pools during high flow events. This could explain the increased deposition on riffles (Figure 6.24) compared with decreased erosion on riffles at high flow (Figure 6.23). There is also a tendency for pools in straight reaches downstream of section 45 to infill with sediment in comparison with riffles in straight reaches.

It should be noted that if the total area of channel deposition is used rather than per unit width, riffles and meander bends show a much steeper downstream decrease in channel deposition than for pools since their width tends to be more variable.

Figure 6.24 shows that channel deposition tends to be concentrated on meander bends, for example, sites 26, 59, and 80. Since bank erosion is concentrated on the concave bank of a meander bend, aggradation within the channel and on adjacent point bars

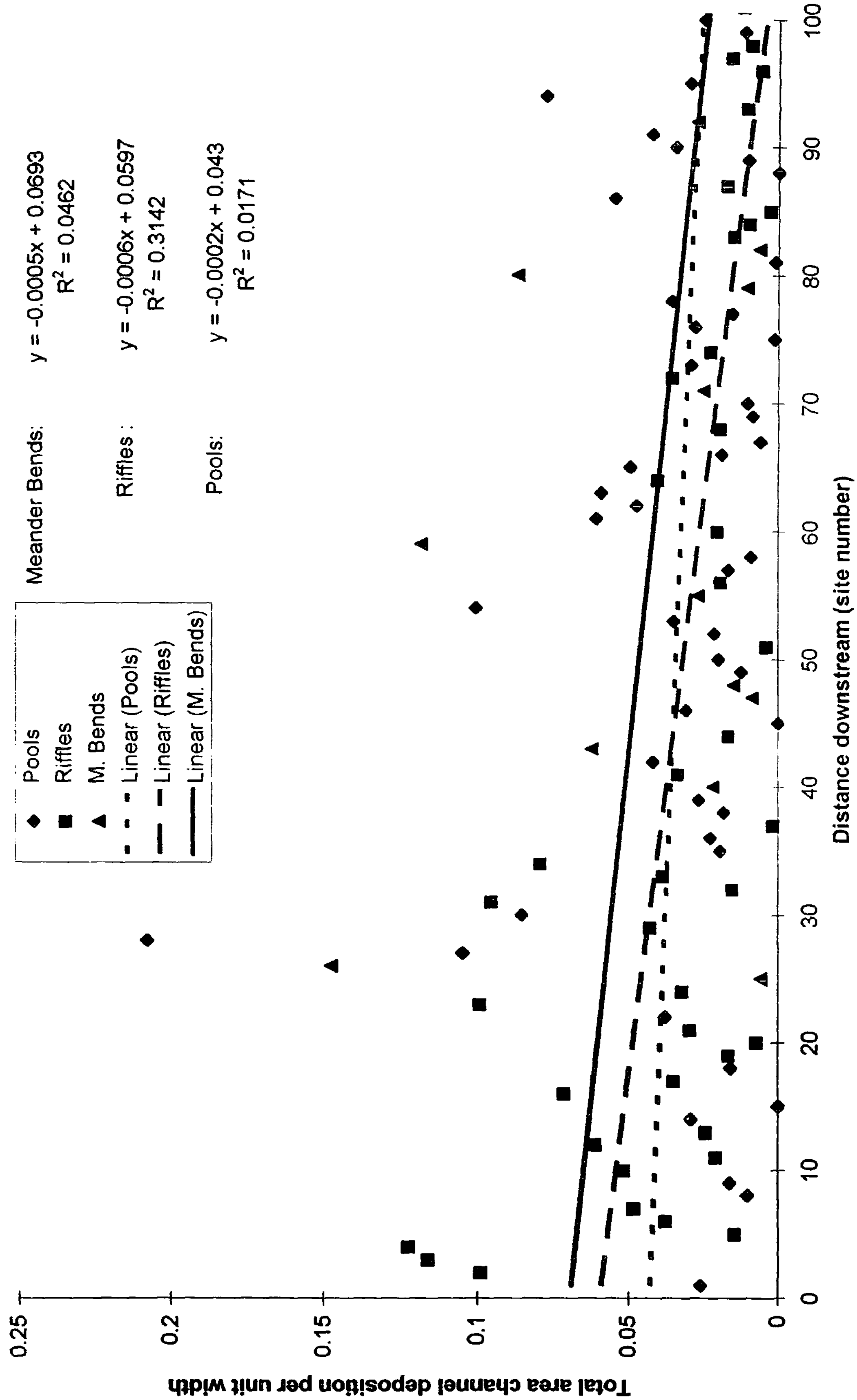


Figure 6.24 Downstream changes in total area of channel deposition/unit width in relation to the pool-riffle, bend sequence

tends to result from transverse bed material transport and secondary circulations (Milne, 1982b). This explains why both erosion and deposition are greatest at meander bends.

Figure 6.25a shows that meander bends have the greatest net change in bank area, the greater change resulting from bank erosion rather than bank accretion. A negative value on Figure 6.25a indicates bank erosion whereas a positive value indicates bank accretion.

In the upper reaches on meander bends there is a greater tendency for the net change in bank area to occur through bank accretion with bank erosion becoming more prominent in the lower reaches. This downstream increase in bank erosion may be explained by the fact that bend sinuosity in the lower reaches increases due to the cohesiveness of bank sediments. The channel geometry on high sinuosity bends means that the secondary circulation within the channel has increased erosional power, leading to increased bank undercutting and bank erosion. Milne (1982c) observed that along sinuous reaches of coarse bedload channels the widest bed forms occurred in association with bank erosion at tightly curved pool sites around actively migrating bends.

The net change in bank area of riffles is slightly less than that for meander bends, although it shows a similar downstream pattern. Bank erosion tends to be prevalent at riffle sites since the flow is directed towards one or both banks from coarse riffle deposits within the channel. The steeper gradient and subsequent increase in velocity of flow over riffles is a contributory factor to bank erosion at these locations.

The trendline for pools in straight reaches shows very little net change in bank area, although the lower reaches show increasing bank erosion. However, overall for pools, riffles and meander bends there is very little change in net bank area, particularly in the upper reaches of the channel.

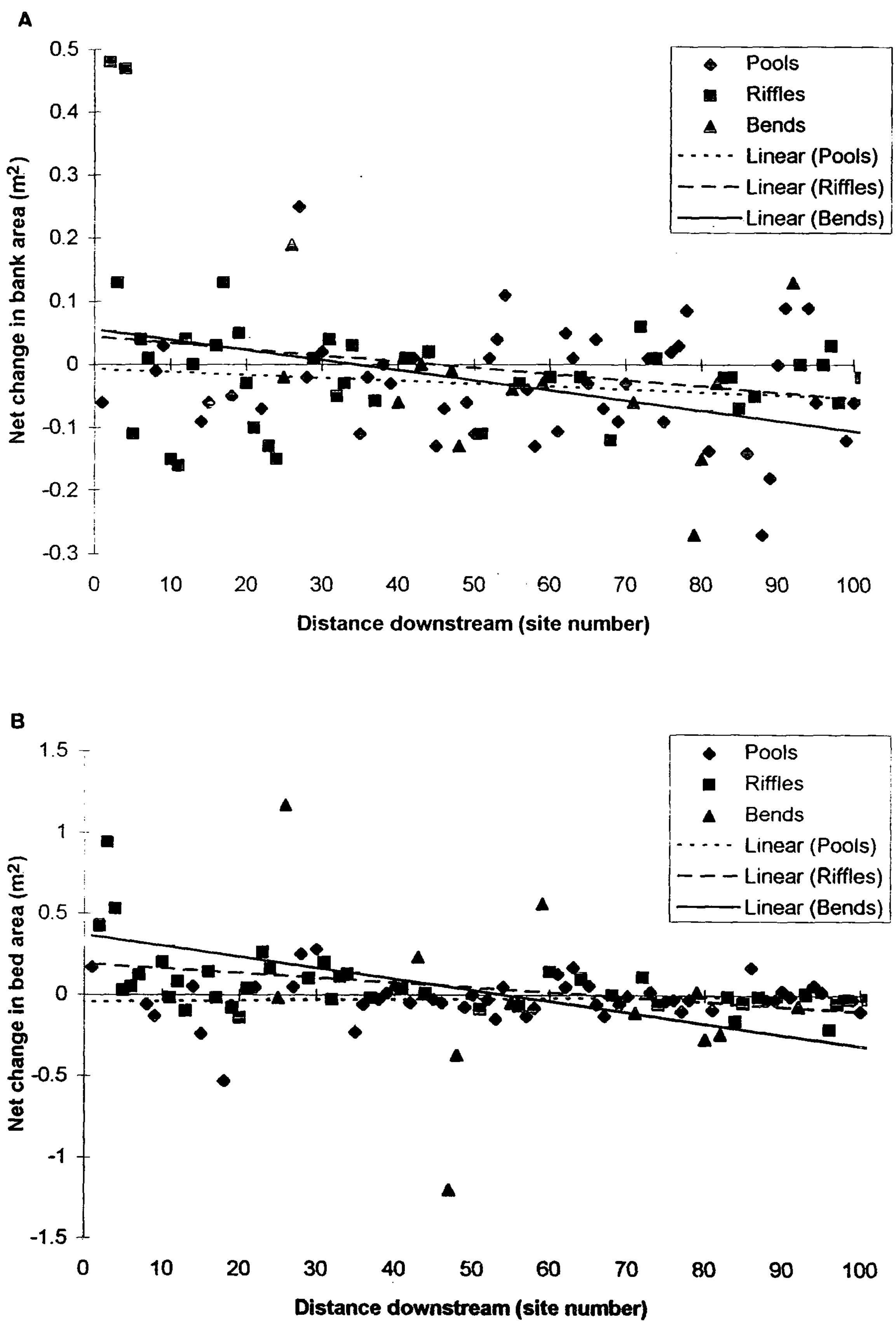


Figure 6.25 Downstream changes in net change in bank (a) and bed (b) area in relation to the pool-riffle sequence

Figure 6.25b shows a very similar pattern to Figure 6.25a in that at the majority of sites there is very little change in the area of bed and banks. Whereas Figure 6.25a shows a large amount of scatter in the plot due to local variations in bank material, Figure 6.25b shows less scatter and the majority of change in bed area occurred at a small number of sites. For example there are two striking outliers, the first at site 26 with over $1m^2$ in bed accretion and site 47 with over $1m^2$ in bed erosion, both of which are meander bend locations. It is likely that these two outliers have influenced the linear regression line producing a much steeper slope for meander bends than for pools and riffles.

The majority of change in bed area has occurred at meander bend locations, with predominantly bed aggradation in the upper reaches and bed erosion in the lower reaches. As mentioned earlier, bend locations are the sites where the majority of channel change takes place. However, there are exceptions where bends impinge on streamside scars. The input of coarse sediment prevents bank erosion and scouring of the bed at the base of cut banks. This may help to explain some of the level of scatter on Figures 6.25a and 6.25b.

The net change in bed area of riffles follows a similar downstream pattern to that of meander bends with bed aggradation on riffles in the upper reaches and bed erosion on riffles in the lower reaches, although the net change is less. This may have resulted from the compact structure of the channel bed over riffles, which may have minimised erosion during high flow events (Clifford, 1993; Sear, 1996). During high flow events riffles tend to lose their identity as water depth increases resulting in equalisation of competence of pools and riffles. Research has shown that it is unlikely for the largest particles on riffles to move except during extreme flows, which means that riffles are stable bedforms (Knighton, 1998). Therefore it is more likely for bed aggradation to occur at riffle locations than bed erosion.

The trendline for net change in pools in straight reaches shows very little change resulting from the passage of a flood. In contrast, Knighton (1998) suggests that due to the velocity reversal hypothesis competence could be higher at high flows in the pools since their bed sediments have a more open structure than that of riffles. This would suggest that a greater change in bed area would be expected in pools than is illustrated in Figure 6.25b.

Summary

The least amount of channel erosion occurs at riffles which supports the theory that at high flows competence is higher in pools than it is in riffles. Nevertheless, bank erosion tends to be important at riffle sites where flow is directed towards the channel banks. However, if the flow is sufficiently high or is overbank, mid-channel bars and coarse deposits may become submerged thus inhibiting further bank erosion.

It must be stressed that at the majority of sites there was very little change in either the bed or banks in response to the flood of February 19th and 20th, which may be partly due to flow being overbank in many sections of the channel, particularly in the upper reaches. Once the flow is overbank the erosive power of the stream is significantly reduced and deposition of material is spread over the adjacent floodplain. However, the overall lack of change observed at the 101 cross-sections is also a function of channel gradient, which will be considered in a further section. This correlates well with observations of historical channel change which suggest very slow rates of channel evolution.

6.8 The relationship between bank material, bank form and the occurrence of channel bars and changes in cross-sectional form during flood events

The composition of the banks along the study reach at Swinhope Burn is an important factor determining both the cross-sectional form of the channel and development of channel planform. Bank composition influences where bank erosion is most likely to occur and consequently conditions patterns of channel sedimentation (Milne, 1982b, Ferguson, 1981).

The presence of channel bars, slumped and cut banks and toe deposits at the base of banks will also influence the response of cross-sectional form to major flood events. Channel bars both protect banks from erosion and often direct the flow towards the opposite banks causing local erosion. The presence of a slumped bank or substantial

Figure 6.26 Example of flow being directed away from a bar surface towards the opposite bank causing bank undercutting and failure

channel deposits at the base of banks also locally reduce or prevent bank erosion and redefine local flow patterns.

The influence of the presence of channel bars and cut banks on channel erosion and deposition during flood events will be assessed.

The influence of channel bars and cut banks on channel erosion and deposition during flood events

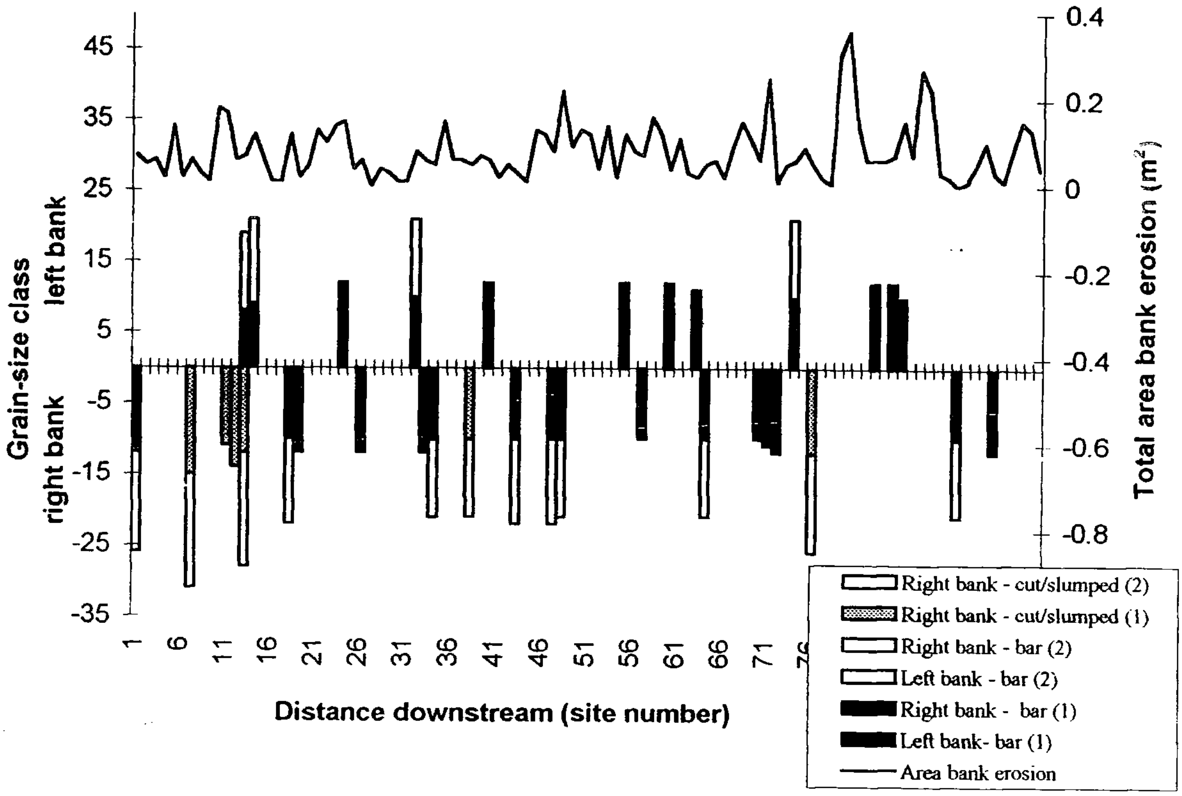
Channel bars are sites of channel deposition and are therefore locations of net bank accretion. However, as the river shifts its course laterally deposition against one bank will be approximately compensated by erosion of the opposite bank. This may result from flow being directed away from the bar surface towards the opposite bank causing bank undercutting and erosion, which is the case along the study reach at Swinhope Burn (Figure 6.26). Therefore at sections with channel bars there is also likely to be an increase in bank erosion at that section or an adjacent one, for example, at section 71 (Figure 6.27a).

Bank accretion tends to occur at sections with coarse channel toe deposits since other particles may become trapped at the base of the bank during transport. These deposits may direct flow towards the opposite bank causing erosion. Cut banks are also sites of bank accretion since as the stream impinges on the valley side slope basal undercutting leads to slope instability, failure and the deposition of coarse material into the channel for example at sections 11-13 (Figure 6.27b).

However, the extent of bank erosion opposite bars will be largely controlled by the cohesiveness of the banks. As the grain-size of bank material decreases downstream bank erosion is likely to decrease and bed erosion may increase opposite bars, as the channel is deepened rather than widened. Conversely, in the upper reaches bank erosion opposite bars is likely to increase due to the presence of composite banks. Bank erosion will occur by fluvial entrainment of material from the lower, cohesionless bank at a much higher rate than the material from the upper more cohesive bank. This leads to bank undermining and subsequent bank failure (Thorne and Tovey, 1981).

In Swinhope Burn, the evidence that channel bars promote erosion of the opposite bank is not as clear as the evidence that channel bars are sites of channel deposition. This is because the influence of bank composition on bank erosion is equally as important as the presence of a bar on the opposite bank. Ferguson and Werritty (1983) identify the process of episodic elongation and advance by deposition of bars during floods as a mechanism by which bank erosion occurs in adjacent channels in the River Feshie, Scotland. This process may occur within particular sections of the study reach of Swinhope Burn where the banks are less cohesive. This theory is supported by Lewin (1976) on the River Ystwyth in Wales where, following straightening of the channel during a few competent flood events, bars became attached to alternate sides of the channel and erosion occurred on the opposite bank, re-establishing a meandering channel. Bank erosion was quite localised and occurred at the point of maximum current attack. Knighton (1998) suggests that erosion of one bank is approximately

A



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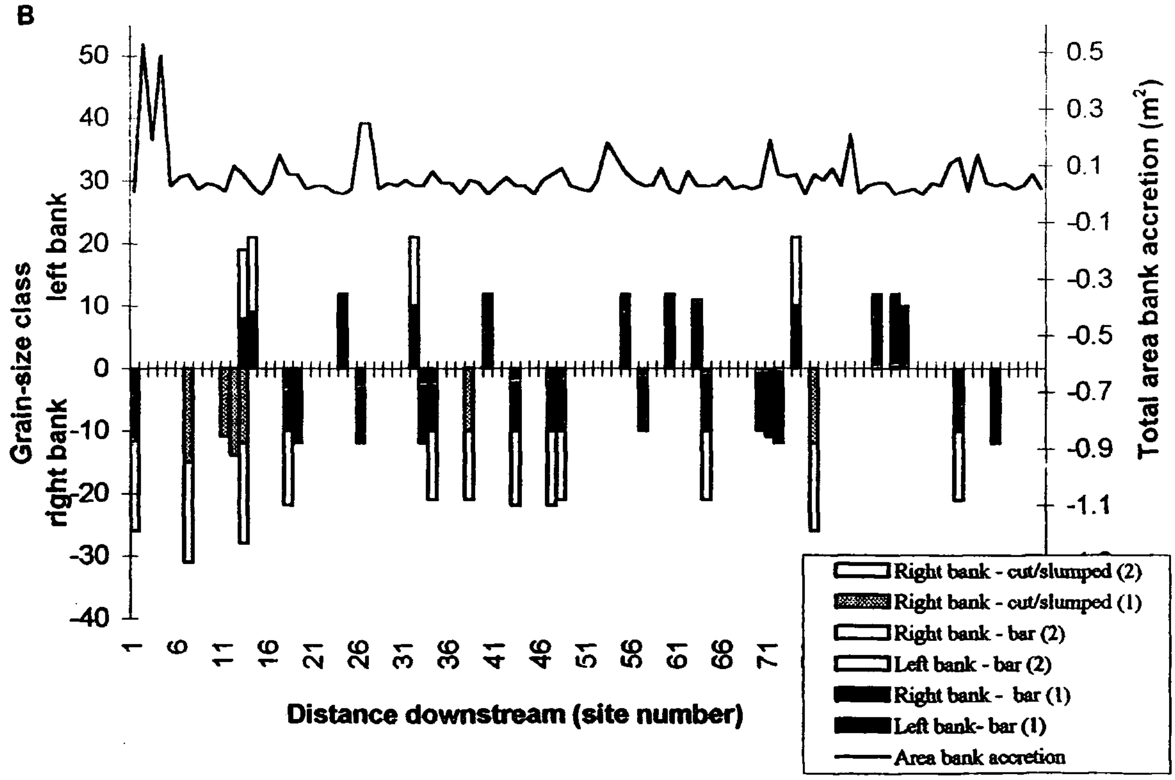


Figure 6.27 Downstream changes in bank erosion and bank accretion in relation to the presence of bars and cut-banks

compensated by deposition against the other as the river alters its course laterally, leading to the development of either a point or lateral bar. This appears to be the case throughout the study reach at Swinhope Burn.

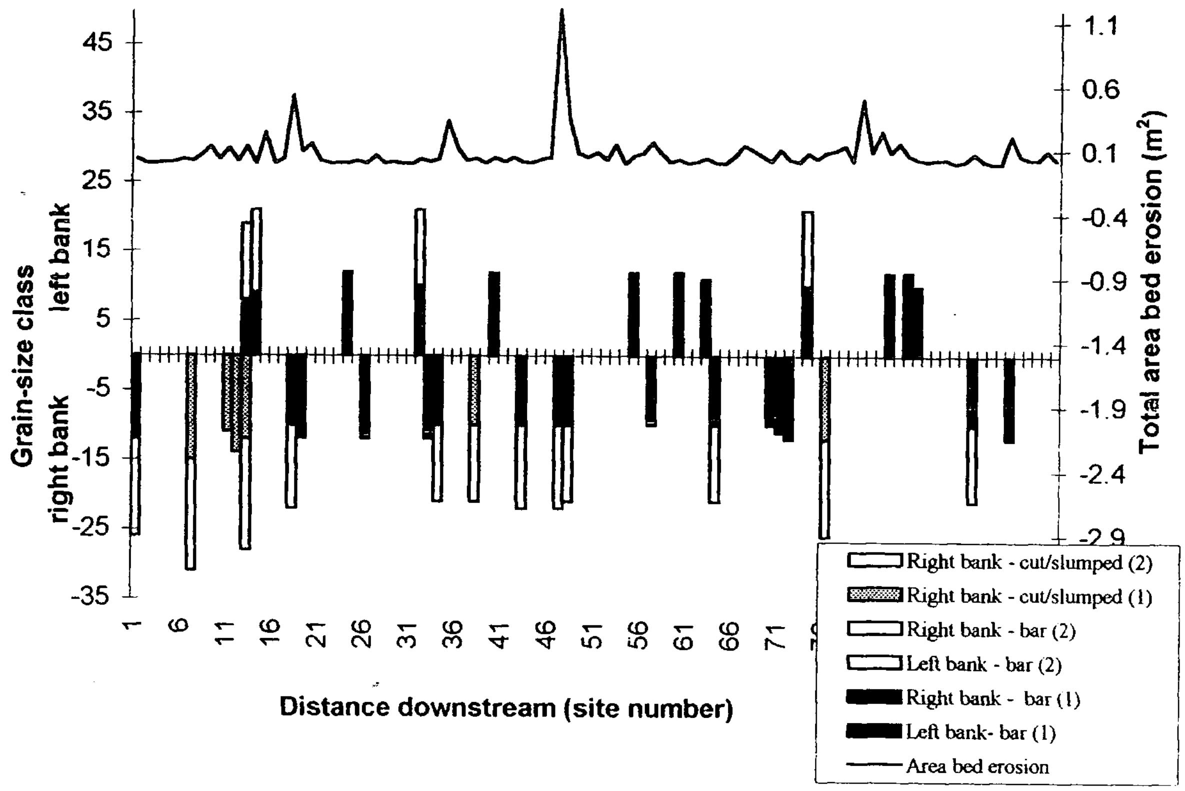
From the foregoing discussion on the role of channel bars in promoting the occurrence of both bank accretion and bank erosion (on the opposite bank) it may be suggested that vertical accretion on bar surfaces may eventually build the bar to a height which can restrict the lateral migration of the channel, although bank erosion may still occur on the bar and on the bank opposite the bar.

In the case of Swinhope Burn, there is also evidence to suggest that the presence of a bar encourages erosion of the adjacent channel bed, for example, at sections 18 , 47 and 96 (Figure 6.28a). This relationship is most clear in the lower reaches where increased bank cohesivity further encourages vertical rather than lateral adjustment of the channel. There is also evidence to suggest that bars encourage bed aggradation, for example, sections 26, 43 and 86 (Figure 6.28b). This may be due to the restriction of overbank flow at these locations since the channel banks can confine flood flows, which can subsequently lead to deposition within the channel. Knighton (1998) has suggested that as bars increase in height as sediment is carried on to its surface by inundating flows, the channel will eventually become confined. Progressively finer sediment will be deposited on the bar as it increases in height. If the channel is confined it is more likely that high flows will be contained within the channel, with deposition occurring on the channel bed rather than overbank.

<u>Summary</u>

Where banks are cohesive, bars can encourage bed erosion. The influence of bank composition on bank erosion is as important as the presence of a bar on the opposite bank. Accumulation of material on lateral bars may also increase bankfull depth and inhibit overbank flow leading to further channel deepening. Conversely, bars help to protect banks from erosion and as vertical accretion on the bar top advances they become attached to the floodplain. The presence of a bar can also encourage bed aggradation as bedload in transport becomes trapped on the bar surface during high flows.

A



B

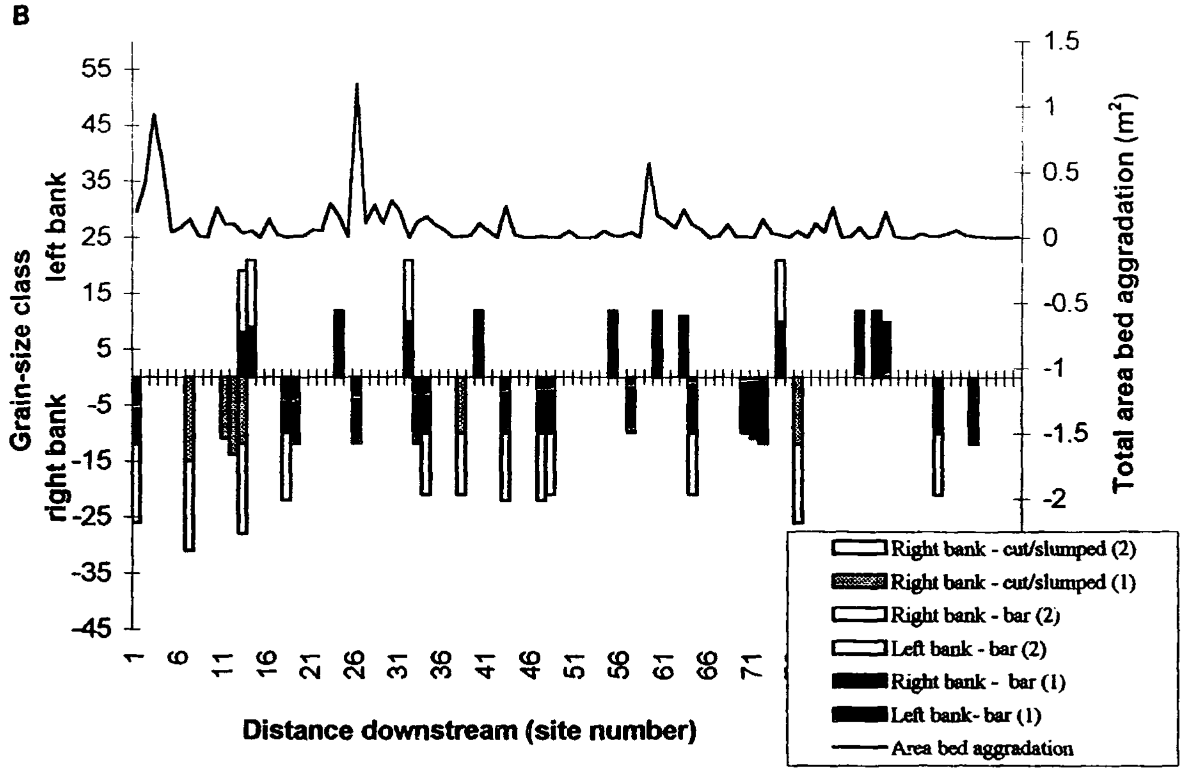


Figure 6.28 Downstream changes in bed erosion and bed aggradation in relation to the presence of bars and cut-banks

Similarly, the presence of a cut bank which introduces coarse material into the channel, protects the base of the bank from erosion but can encourage erosion of the bed particularly if the channel is confined on both sides. Coarse channel toe deposits can similarly influence the location of within-channel deposition as bedload becomes trapped within these deposits. Toe deposits within the channel can divert flow towards the channel banks causing erosion and those at the base of banks can protect the banks from erosion.

6.9 The role of channel sinuosity in determining changes in cross-sectional form

The sinuosity of individual bends within meandering streams can influence resistance to flow during flood events and this can result in downstream variations in the amount and location of channel erosion. Meandering streams will develop if there is sufficient energy to cause bank erosion but sufficient bank resistance to maintain a sinuous course. However, if a meandering channel cuts through cohesive bank sediments the channel will become increasingly sinuous with some very tight turns particularly where the current from upstream meets the cohesive sediment (Milne, 1983a). Stronger secondary circulations then become established leading to better developed point bars and greater cross-sectional asymmetry. This will reinforce secondary currents, increase bend curvature and further concentrate erosion. This promotes excessive deepening of the channel and restricts the lateral migration of the channel as the bank retreats slowly at a steep angle. Therefore a strong correlation between high sinuosity and bed erosion might be expected.

In order to investigate the effect of variations in channel sinuosity on channel erosion at the bend apex along a 1.4 km reach of Swinhope Burn, the sinuosity of each bend was calculated. The position of each of the 19 bends shown on Table 6.2 is identified on Figure 6.29.

Table 6.2 shows total channel erosion, total bank erosion, total bed erosion and bend sinuosity for the 19 bends identified on Figure 6.29. Bend sinuosity varies from 0.96 from sections 3-10 at the head of the study reach, to 6.17 from sections 92 to 101 at the end of the study reach. In general, the channel becomes more sinuous downstream.

Bend	Section	Bend	Total area	Total area	Total area
	Number	Sinuosity	bank erosion (m^2)	bed erosion (m^2)	channel erosion (m^2)
1	7	0.96	0.06	0.02	0.08
1a	10	1.15	0.18	0.03	0.21
2	13	1.29	0.07	0.13	0.2
2a	18	1.48	0.12	0.53	0.65
3	21	1.76	0.13	0.02	0.15
3a	25	1.65	0.04	0.02	0.06
4	31	1.7	0.01	0	0.01
4a	35	2.04	0.15	0.33	0.48
5	40	2.53	0.06	0.05	0.11
5a	47	3.82	0.08	1.21	1.29
6	53	3.33	0.14	0.15	0.29
6a	59	1.58	0.12	0.01	0.13
6b	63	1.67	0.02	0.04	0.06
6c	71	3.28	0.25	0.11	0.36
7	76	3.18	0.05	0.08	0.13
7a	80	3.14	0.36	0.5	0.86
7b	85	1.06	0.06	0.16	0.22
7c	92	1.58	0	0.08	0.08
8	96	6.17	0.03	0.22	0.25

Table 6.2 Relationship between bend sinuosity and channel erosion at bend apex

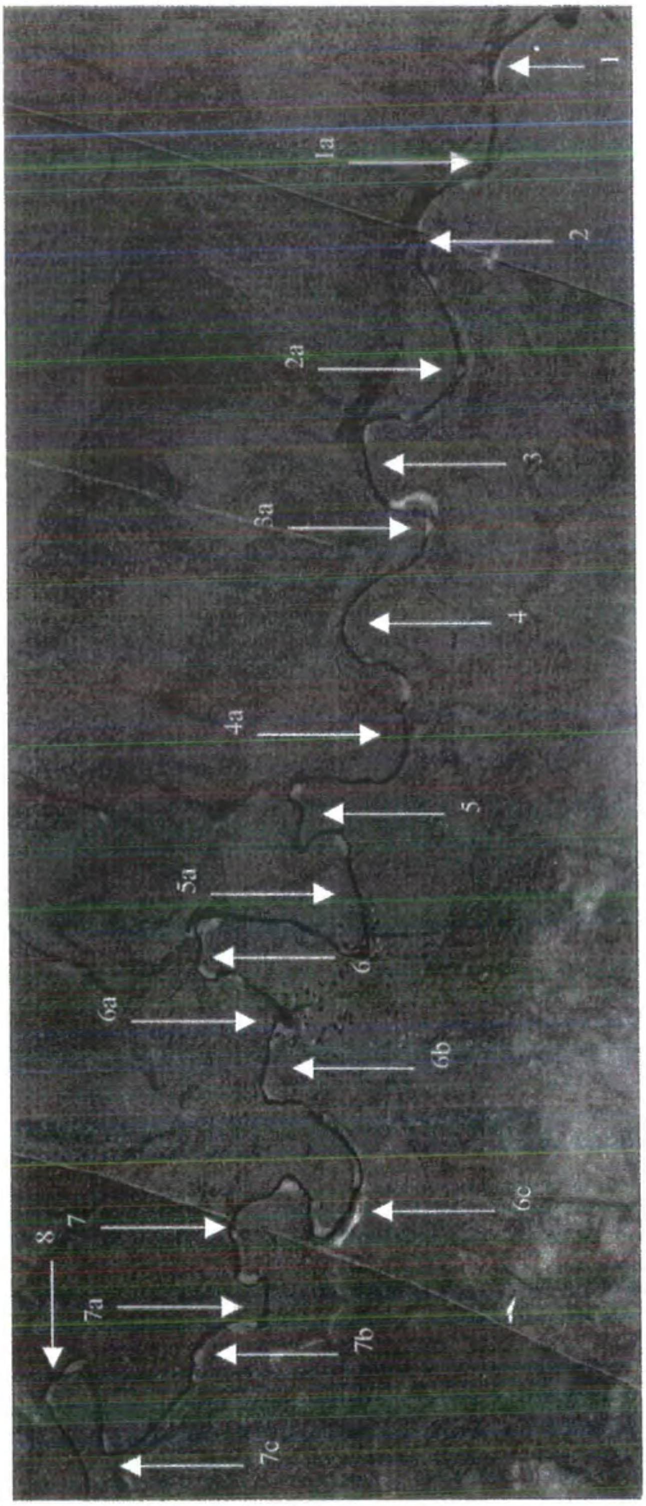


Figure 6.29 Location map of bends along the study reach of Swinhope Burn (Air photograph courtesy of Aerofilms and Durham County Council).

However, there are anomalies, for example, two gently curving bends (Bends 7b and 7c) lie immediately upstream of the most sinuous meander bend from sections 92 to 101 (Bend 8).

Figure 6.30 shows the relationship between bend sinuosity and channel erosion at the bend apex for the 19 sections of channel identified in Figure 6.29 (Table 6.2). Although there is some scatter in Figure 6.30, as illustrated by the low R^2 values, it appears that there are positive relationships between channel erosion at the bend apex and bend sinuosity. The relationship between bend sinuosity and bed erosion may be explained with reference to Milne's (1983a) hypothesis that highly sinuous channels which cut through cohesive bank material often have tight bends where the current from upstream meets cohesive sediment which restricts channel widening and encourages vertical deepening of the channel through bed erosion.

However the relationship between increasing bend sinuosity and increased channel erosion is not clear. It appears that once a meander bend has developed a very high sinuosity index, for example the final bend of the study reach (sections 92 - 101, bend 8) channel erosion at the bend apex during flood events may not be as high as expected. Table 6.2 shows a bend sinuosity of 6.17 at bend 8, compared with a bend sinuosity of 3.14 at bend 7a, positioned slightly upstream. Although the sinuosity of the final bend of the study reach is twice that of the bend between sections 77 and 84 (7a), the total area of channel erosion recorded at the apex of bend 8 was $0.25m^2$ compared with $0.86m^2$ at the apex of bend 7a.

A possible explanation for this is the observation that during the flood some of the flow was re-directed from the bend apex at section 92, crossed the floodplain and entered the channel around section 100, effectively cutting off the meander bend (Figure 6.7). Since the flow was diverted across the floodplain, discharge flowing through the bend would have been reduced, leading to a reduction in stream power. This may have resulted in the observed decrease in channel erosion at this bend.

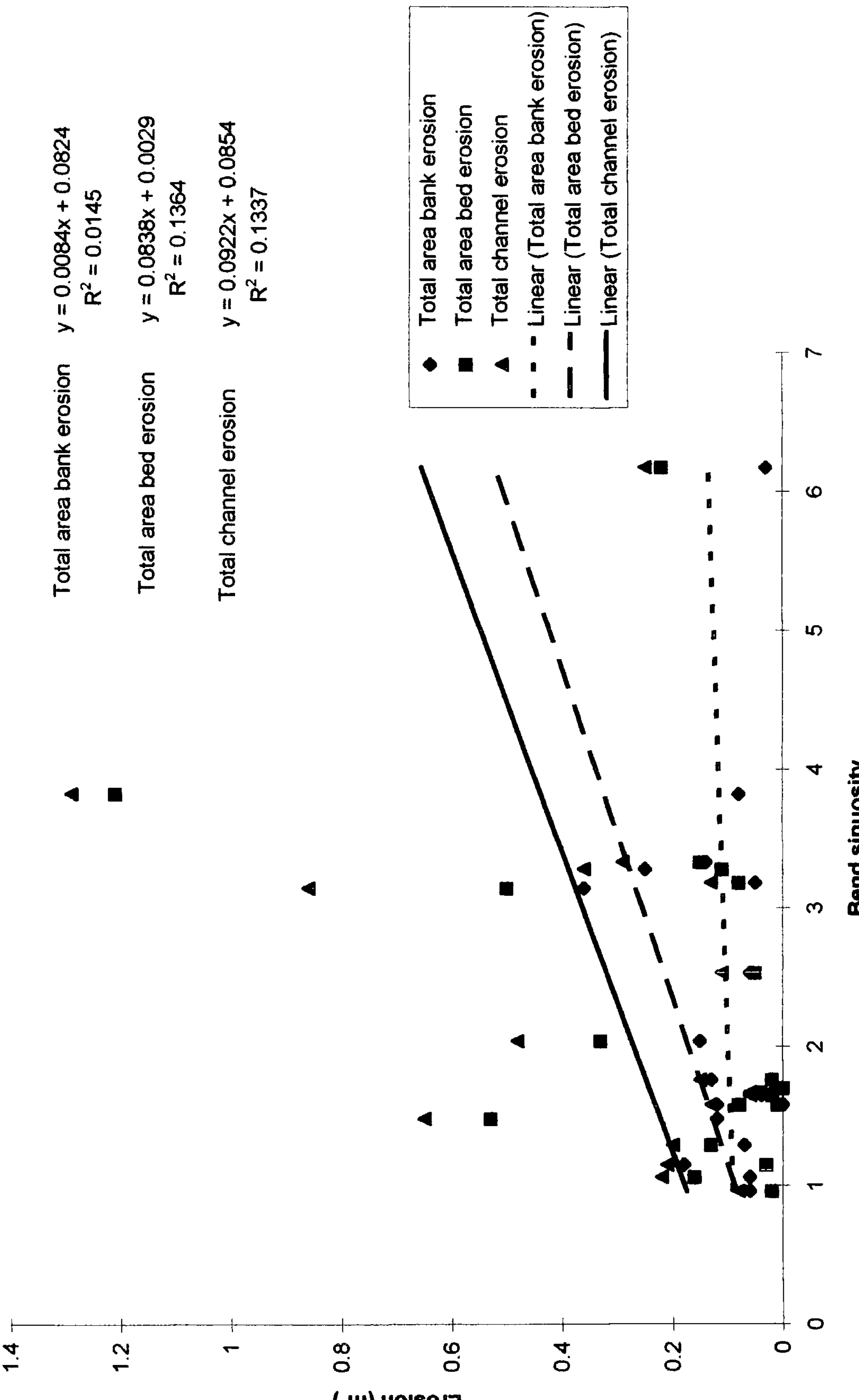


Figure 6.30 Relationship between bend sinuosity and channel erosion at the bend apex

On two particular bends within the middle reaches of Swinhope Burn (bends 4 and 5) sinuosity is relatively high (1.7 and 2.53 respectively). However, total channel erosion at the bend apex on these bends is extremely low. Although there is a relationship between increasing bend sinuosity and channel erosion, in particular bed erosion, localised factors such as the pool, riffle, meander bend sequence, local channel gradient, inputs of coarse sediment from channel bluffs and variations in bank cohesion can complicate the relationship (Ferguson, 1981). In Swinhope Burn, a good correlation between bank erosion at the bend apex and sinuosity is not expected given variations in bank material and the slow rate of change.

6.10 The role of channel gradient on change in cross-sectional form and mean grain-size distribution

The step in the long profile of Swinhope Burn, produced by a local base-level, is highly significant when considering stream response to contemporary flood events. The extent and nature of change in cross-sectional form and mean grain-size observed at each of the 101 cross-sections following a major flood event is largely controlled by the very low gradient of the basin. Ferguson et al (1996) explain the process of downstream fining of bed material in response to a local base level as occurring in the form of a wave of fine bedload migrating along the concave profile of the Allt Dubhaig, initially causing rapid aggradation and fining of bed material in the upper part of the study reach. However as channel gradient is reduced, degradation and coarsening occurs in the lower parts of the study reach. The influence of a low channel gradient on the adjustment of upland stream channels to floods has been identified by Milne (1983a), who suggested that high amplitude 'goose-neck' meander bends form particularly where low gradient channels flow through thick fine-grained cohesive sediments. Milne (1979) suggests that a locally low gradient can lead to a lack of stream competence and a weak relationship between the width and depth of the channel.

Influence of channel gradient on downstream fining of bed material and cross-sectional channel stability

Analysis of downstream changes in the mean grain-size distribution of Swinhope Burn following a major flood event identified a clear pattern of downstream fining throughout the reach. The significant reduction in downstream grain-size can be attributed to the decreasing gradient of the channel. In the absence of any major tributaries supplying coarse sediment to the study reach it is more likely that particle size will decrease systematically downstream with channel gradient (Knighton, 1975, 1980).

Statistical comparison of downstream variations in mean grain-size before and after the flood event indicated that there was no significant difference between the two series, although a slight fining of bed material in the upper reaches and coarsening of bed material in the lower reaches was distinguishable (Warburton and Danks, 1998). The channel bed throughout the reach is largely static although during high flow events limited sediment movement does occur. The lack of movement in the channel for long periods of time is confirmed by the presence of a thick algal slime which covers the gravel-bed surface, particularly during the summer months. The mean grain-size is relatively small within the study reach when compared with the headwater reaches of Swinhope Burn where the grain-size of bed material is considerably coarser. In the headwater reaches, the gradient is much steeper and coarse material is supplied by numerous small tributaries and from composite channel banks.

The relatively low slopes bounding the study reach and the extremely low channel slope have reduced the ability for sediment transport. Also, lack of coupling between the channel and valley sides slopes reduces stream access to possible sources of coarse sediment. Any coarse sediment entering the reach tends to become trapped since the gradient is not steep enough to transport it. Finer material is likely to be transported downstream at very low rates, which suggests that the study reach is essentially a sedimentation zone.

The lack of change in cross-sectional form of the channel following a major flood event is partly due to the very low channel gradient throughout the study reach. During flood events, a reduction in stream power along the reach occurs due to the low channel gradient, as identified in Chapter 5 (Table 5.3). Since upstream of the study reach, channel gradient is much steeper it is likely that once the gradient begins to flatten out, in the upper sections of the study reach, channel deposition is likely to occur rather than channel erosion.

Local increases in channel gradient in relation to the pool-riffle sequence cause localised increases in channel change, but the effect of floods on the reach as a whole is relatively insignificant, with bankfull width and depth remaining virtually unchanged and channel erosion being of the same order of magnitude as channel deposition.

6.11 Discussion and Conclusions

Survey of the 101 cross-sections, measurements of bed material grain-size and mapping of the channel pattern and flood limits using Differential GPS suggested little change in response to the flood. Bankfull width and bankfull depth were virtually unaltered and channel erosion is of the same order of magnitude as channel deposition (Table 6.1). Although it was found that channel deposition was largely responsible for changes in cross-sectional morphology in the upper reaches and channel erosion was largely responsible for observed changes in the lower reaches, plots of cumulative channel erosion and deposition (Figure 6.20) converge towards the end of the study reach indicating that the changes along the channel as a whole are insignificant. Although both before and after the flood downstream fining in mean grain-size was apparent, comparison of the two data series indicates there is no significant difference in mean grain-size sampled before and after the flood.

Although very little downstream change in cross-sectional morphology or the grain-size of bed material was observed indicating that the channel is highly stable even during a major overbank flood, small adjustments in channel form and grain-size of bed material occurred at individual cross-sections. Local variations in bed configuration, bank cohesion and bar formation, and channel planform in addition to the more reach-wide influence of channel gradient influence stream response to the passage of the flood. During high flow the majority of channel erosion and deposition occurs at meander bend locations. Pools associated with the meander bend apex focus erosion at the concave bank. However, the nature of bank sediments is critical in influencing the development of adjacent bed forms (Milne, 1982b). Riffles are the most stable bedform because they tend to have the coarsest bed material which has a closely packed structure and which resists erosion during high flow events (Clifford, 1993; Sear, 1996). Once formed, riffles, mid-channel accumulations of coarse material or the presence of central bars can deflect flow towards one or both banks which encourages bank undercutting and increased channel widening (Milne, 1982c; Knighton, 1998). Local variations in the response of pools and riffles to high flow events has been explained with reference to the velocity reversal hypothesis. Higher velocities in pools than on riffles during high flow events and their loose bed structure compared to riffles combines to encourage transport of the largest particles through the pool rather than their deposition within it (Clifford, 1993). This explains why during the February, 1997 flood event, higher levels of total channel erosion are observed on meander bends and pools in straight reaches than for riffles. The infilling of pools with coarse material, during high flow events, which is then quickly transported from one pool to another over intervening riffles is well documented in the recent literature (Carling, 1991; Clifford 1993; Sear, 1996). In the case of Swinhope Burn, very little net change in bed area of pools in straight reaches (Figure 6.25b) suggests that coarse sediment is transported from one pool to another rather than being deposited in pools during flood events.

The composition of the banks along the study reach at Swinhope Burn is an important factor determining both the cross-sectional form of the channel and development of the channel planform. The composition of the banks influences where bank erosion is most likely to occur and consequently conditions patterns of channel sedimentation (Milne, 1982b; Ferguson, 1987).

It has been suggested that bars both interact with and influence the pattern of flow throughout the reach. They influence channel morphology because they are major stores for bedload and because they encourage bank erosion of the opposite bank through flow being directed away from the bar surface. However, erosion of one bank is approximately compensated by deposition against the other (Knighton, 1998). The role of cut banks and coarse channel toe deposits in both protecting the base of the channel banks from erosion and encouraging erosion of the opposite banks through flow diversion is demonstrated.

Substantial local variations in channel erosion and deposition at the cross-sectional scale have been linked with variations in bed topography, in relation to the pool, riffle and meander bend sequence, and with a downstream decrease in grain-size of bank material and the presence of bars and cut banks. On a reach scale, bend sinuosity and channel gradient are critical factors which influence the movement of bedload through the channel during floods and control the amount of channel erosion and deposition along the channel.

In order to investigate the effect of variations in reach sinuosity on the development of cross-section form the relationship between increasing bend sinuosity and channel erosion was determined, based on variations in sinuosity of eight bends along the study reach. It was found that channel erosion at the bend apex increased with increasing bend sinuosity, the strongest relationship being with bed erosion at the bend apex. However the relationship between increasing bend sinuosity and increased channel erosion is not clear and is likely to have been complicated by more localised factors including bed configuration, bank cohesivity and localised lateral inputs of coarse sediment.

The step in the long profile of Swinhope Burn, produced by a local base-level in the form of the Greenly Hills moraine, is highly significant when considering stream response to contemporary flood events. The step in the profile has led to the development of a very low channel gradient (0.0118) when compared with the upper reaches of other adjacent streams in Weardale. This has led to a significant reduction in downstream grain-size, which is demonstrated by the grain-size survey both prior to and following the February, 1997 flood event.

The low channel gradient and slopes bounding the study reach have reduced the ability of the stream to transport sediment and coarse inputs of sediment from the valley-side slopes are limited. The study reach is essentially a zone of sedimentation, since although fine material can be transported downstream, coarse bedload from the uppermost reaches of Swinhope Burn becomes trapped and is only transported downstream at low rates even during major flood events. The observed lack of change in cross-sectional form following the flood is related to the reduction in stream power caused by the very low channel gradient. Where stream power is momentarily increased due to local increases in channel gradient associated with the pool, riffle, meander bend sequence, small adjustments in channel form are observed, for example, the scouring of deep pools on meander bends.

It seems highly likely that the low gradient of the study reach at Swinhope Burn is the key factor in determining contemporary channel response to flood events. When major flood events occur, the low channel gradient limits coarse bedload transport through the reach and deposition tends to occur within the upper reaches. Changes in cross-sectional form occur largely in response to increased discharge and sediment transport during high flows events. Therefore, if low channel gradient reduces the ability for sediment transport, morphological change is likely to be restricted and the channel will be inherently stable. This explains why no major changes in cross-sectional form or in the mean grain-size of bed material occurred in response to the flood of February, 1997.

However, bank cohesion is also important in channel response to flood events. In the upper reaches, composite banks have resulted in a wide and shallow cross-sectional form. High flows generated during the February 1997 flood could not be contained within the shallow channel and the flood was overbank in many places. This reduced the erosive power of the stream and limited the amount of channel change. However, in the lower reaches, highly cohesive bank material combined with a very low gradient has produced a highly sinuous, deep and narrow channel. The highly sinuous nature of the channel led to increased amounts of channel erosion during the flood event.

The analysis of downstream variations in the patterns of erosion and deposition has revealed that channel aggradation was more prominent in the upper reaches, whereas channel erosion was greater in the lower reaches. Field evidence of aggradation in the upper reaches is provided by the presence of large quantities of overbank fines covering the floodplain adjacent to the channel margins. Likewise sampling of bed-material prior to and following the flood indicated that upstream of section 40 there had been a decrease in mean grain-size with an increase in mean-grain-size downstream of this point. It is possible that the input of fines from upstream bank erosion diluted the pre-existing coarser bed material.

A possible explanation for these observations is that a sediment wave moved through the upper reaches of Swinhope Burn during the flood of February 1997. This caused aggradation and a fining of the existing bed material. As the gradient began to level out, around the middle reaches of Swinhope Burn, the sediment transport capacity of the stream was reduced and the wave came to a standstill. Once deposition had occurred an increase in sediment transport capacity may have caused erosion of the finer fractions of the bed material, resulting in the observed coarsening of mean grain-size in the lower reaches following the flood. This hypothesis of the role of channel gradient and migration of sediment waves in upland gravel-bed streams is supported by Ferguson et al (1996).

CHAPTER 7

CONTEMPORARY RIVER CHANNEL CHANGE: WITHIN-REACH SEDIMENT DYNAMICS

7.1 Introduction

A secondary objective of this project is to examine both within reach and through reach sediment dynamics during contemporary flood events at Swinhope Burn. In order to achieve this objective a sediment tracing experiment was designed. The experiment aims to establish the relative importance of within-reach sediment dynamics, (sediment supply from cut banks), in comparison with main channel sediment transport through the reach.

In order to identify spatial and temporal variations in sediment dynamics and investigate the mechanisms by which floods transport coarse bedload through the channel system, painted magnetic tracer pebbles were placed at the top of the study reach and painted tracer pebbles were placed at four cut-bank locations along the channel. The tracer pebbles were placed in the stream in March, 1997. Monitoring of the tracer pebbles at Swinhope Burn is ongoing, however, for the purposes of this project results are reported for the first 18 months of the experiment.

Pebble tracing offers a practical method of assessing bedload transport by flood events within upland catchments which may be relatively remote and inaccessible during the winter months (Gomez, 1983; Carling and Hurley, 1987; Carling, 1988; Ferguson and Ashworth, 1991).

Whereas the previous chapter identified changes in downstream grain-size distributions both before and after the flood of February 19th and 20th, 1997, the pebble tracing experiment provides a more precise method of measuring sediment dynamics. Although the tracer pebbles were placed along the study reach in March, 1997, a flood of sufficient magnitude to entrain and transport the tracer pebbles did not occur until January 8th, 1998. The results of the sediment tracing experiment in relation to the 1998 flood event are presented and discussed in this chapter. Results are likely to reflect similar processes to those observed during the February, 1997 flood but exact patterns of sedimentation cannot be assumed.

The objective of this brief chapter is to summarise recent research into within-reach sediment dynamics in gravel-bed streams paying particular attention to the use of painted and magnetic tracer techniques. The design of the sediment tracing experiment used for the study reach at Swinhope Burn is described and the field and laboratory techniques associated with pebble collection, laboratory treatment and re-introduction of the pebbles to the channel are discussed. A flood event which occurred in January, 1998, during which the majority of tracer pebbles were entrained, is described.

7.2 Background

The nature and incidence of bedload transport during flood events has been well documented in the Northern Pennines in recent years (Carling and Reader, 1982; Carling and Hurley, 1987; Carling, 1989). In response to a need for investigations into bedload transport processes in upland environments, pit-type bedload traps have been used to measure bedload movement (Carling 1988). These investigations clearly identify the importance of infrequent high magnitude flood events in transporting the majority of bedload in upland gravel-bed streams.

However, direct sampling of bedload in upland gravel-bed streams is problematic. For instance, large bedload samplers tend to disturb the flow and modify transport rates near the device, and where the stream contains large cobbles and small boulders the sampler becomes too heavy and hard to handle (Ergenzinger and Custer, 1982). Where smaller hand -held devices are used during peak flow the equipment can be badly damaged. Tracers provide logistical and safety advantages because they may be installed at low flow thereby avoiding direct sampling of bedload during floods. Although direct bedload sampling provides information on rates of bedload transport and the size of bedload being transported, it does not always provide information on sediment dynamics within or through a reach.

The problems associated with using direct bedload sampling techniques to measure bedload transport during flood events in gravel-bed rivers have necessitated the

Table 7.1 **BEDLOAD TRACING EXPERIMENTS IN UPLAND GRAVEL-BED STREAMS - EXAMPLES FROM THE U.K.**

Tracer Method	Type of Pebble Locator/Sensor	Number/ Weight of Pebbles	Size of Pebbles	Tracer Recovery Rate	Location	References
Magnetic and painted tracer pebbles	Magnetic locator	1460	Above and below local bed D_{50}	66% magnetic 33% painted	Allt Dubhaig, Scottish Highlands	Ferguson et al (1996)
Painted tracer pebbles	Sight	>3,700 in 3 streams	Range 24 mm to 238 mm b-axis.	Alt Dubhaig (av. 72%) Feshie (av. 62%) Lyngsdalselva (av. 58%)	Allt Dubhaig and River Feshie, Scottish Highlands and R. Lyngsdalselva, Norway.	Ashworth and Ferguson (1989)
Painted tracer pebbles	Sight	647 (Great Eggleshope Beck) 279 (Carl Beck)	80% cobbles, 20% pebbles & some boulders Wentworth-Lane nomenclature (15mm - 138 mm b-axis - Carl Beck).	78% - 98% for one particular event the remainder being buried.	Carl Beck and Great Eggleshope Beck, Northern Pennines	Carling (1987)
Magnetically enhanced natural tracers using laboratory heat treatment.	Field search loop with digital read-out.	20-40 kg	Unknown	Data not available	River Trannon, mid-Wales	Newson and Leeks (1987)
Magnetically enhanced natural tracers (optimal laboratory treatment, Oldfield et al, 1981)	Sight (pink pigmentation resulting from heat treatment), Submersible search coil, susceptibility sensing using ferrite probe.	4550 grams	Range 1.4 mm to 44.5 mm.	63%	Ditch system within the Tanllwyth system, Plynlimon Experimental Catchment and River Severn, River Llwyd and River Wye, Wales.	Arkell et al (1983), Moore and Newson (1986)

development of sediment tracing techniques. Since direct bedload sampling is less reliable for the coarsest fractions of the bedload pebble tracer techniques are particularly useful in upland gravel-bed streams (Ashworth and Ferguson, 1989). The use of pebble tracing techniques for fluvial environments were first developed in the mid 1960's (Leopold et al, 1964) and have been used in the British uplands (Carling, 1987; Newson and Leeks, 1987; Ashworth and Ferguson, 1989; Ferguson et al, 1996; Wathen et al, 1997) and in a whole range of locations worldwide (Laronne and Duncan, 1992; Wilcock, 1997; Gintz et al, 1996) (Table 7.1). Initially painted tracer pebbles were used and yielded very poor recovery rates of 5% (Laronne and Carson, 1976). However, improvements of the technique increased recovery rates to 72% on the Allt Dubhaig and River Feshie in the Scottish Highlands (Ashworth and Ferguson, 1989) and 98% on Carl Beck and Great Eggleshope Beck in the Northern Pennines (Carling, 1987). The number of painted tracer pebbles placed on the surface of the bed varies immensely, ranging from 150 in the small, meandering gravel-bed Sagehen Creek in California (Andrews and Erman, 1986) to 4823 in Seale's Brook, Canada (Laronne and Carson, 1976). However, more recently, painted tracer pebbles have been installed in a gravel-bed river by inserting a large cylinder into the bed and replacing the sediment within with marked grains of the same size in order to calculate sediment rates and determine the size distribution of entrained sediment (Wilcock, 1997).

Magnetic tracer pebble techniques are particularly versatile in fluvial geomorphology and the technique has widespread applications. Although natural magnetic pebbles (Ergenzinger and Custer, 1982, 1983; Custer et al, 1987; Carling et al, 1998) and natural tracers magnetically enhanced using laboratory heat treatment (Arkell et al, 1983; Moore and Newson, 1986; Newson and Leeks, 1987) have been used to monitor sediment movement through the channel during flood events, pebbles with magnetic cores in the centre have been more widely used as tracer pebbles (Hassan et al, 1991; Schmidt and Ergenzinger, 1992; Laronne and Duncan, 1992; Gintz et al, 1996; Ferguson et al, 1996; Wathen et al, 1997).

The Nahal Hebron experiment in Israel using ceramic isotopic magnets placed in natural pebbles in conjunction with a magnetic locator was the first sediment tracing experiment to use these techniques and yielded an extremely high tracer recovery rate of 93% over two flood events even though 53% of the pebbles had become buried within the bed. Success in using artificial magnetic tracer techniques in a wide range of fluvial environments, for example, a gravel-bed stream in the Scottish Uplands (Ferguson et al, 1996), and an ephemeral semi-arid gravel-bed stream in Israel (Hassan et al, 1991) demonstrates that the technique is a useful one.

However, there are drawbacks associated with the use of tracer pebble techniques. When pebbles are introduced to the channel they need time to become integrated within the existing surficial bed material. This is particularly a problem in the case of larger pebbles, especially if the bed is formed by coarse framework gravels, as in many upland gravel-bed streams (Carling, 1987). If the tracers are entrained by a flood shortly after re-introduction to the channel the results may not be representative of the behaviour of the natural bed material. The tracer may be entrained at a much lower flow than the natural bed material, which is integrated within the bed structure.

A problem specifically associated with using the painted pebble tracing technique is that searching for tracers is limited to the channel bed, whereas a very high proportion are likely to have been buried within the channel (Laronne and Carson, 1976). Conversely, the main advantage of using artificial magnetic tracing techniques in conjunction with a magnetic locator is that both surficial and buried pebbles can be retrieved following a flood, significantly increasing recovery rates. However when compared with the painted pebble techniques, inserting magnets within pebbles and using sophisticated magnetic locators is costly and time-consuming (Hassan et al, 1984). It appears that the most satisfactory method of tracing fluvial sediment is to place magnets in natural pebbles, spray them with paint and number them for ease of visual identification (Laronne and Duncan, 1992).

7.3 The Sediment Tracing Experiment

Experimental Design

A total of 700 tracer pebbles were used in this experiment: 200 magnetic tracer pebbles were placed within the channel at the top of the study reach (Figure 7.1) and 125 painted tracer pebbles were placed as lateral inputs (Figure 7.2) at four main cut bank locations along the study reach (Figure 7.3). It is appreciated that a sample size of

Figure 7.1 Site 1 - magnetic tracer pebbles placed on the channel bed, extending up to four metres upstream of section 1

Figure 7.2 Site 2 - Painted tracer pebbles placed at base of cut bank

Figure 7.3 Channel planform of Swinhope Burn showing bluffs and tracer input sites

700 tracer pebbles is relatively small when compared with the total amount of bed material within the study reach. The number of tracer pebbles used in the sediment tracing experiment is limited by the practicalities of removing large quantities of sediment from the stream and re-introducing it following tagging.

An initial grain-size survey of bed material within the study reach identified the existing composition of the bed material (Figure 7.4). The b-axis of the coarsest 20 particles on the surface of the channel bed were sampled in a metre square grid which was placed mid-channel and located on the line of each of the 101 cross-sections. In addition at 11 sections the coarsest 50 particles were sampled and at six sections the coarsest 100 particles were sampled. Three grain-size classes were selected as being representative of the coarser bedload within the study reach (Table 7.2):

Site no.	Site location	Tracer Type	Grain-size classes (b-axis in mm) and number of pebbles sampled		
			32 - 45.3	45.3 - 64	64 - 128
1	Top of reach	Magnetic	75	75	50
2	Cut-bank	Painted	50	50	25
3	Cut-bank	Painted	50	50	25
4	Cut-bank	Painted	50	50	25
5	Cut-bank	Painted	50	50	25

Table 7.2: Details of sediment tracing experiment indicating site location, tracer type, grain-size classes used and number of pebbles sampled at each site.

Particles with a b-axis smaller than 32mm were not included in the experiment for two reasons. Firstly, it is impracticable to place a magnet within a pebble which has a b-axis smaller than 32mm as they tend to disintegrate during the drilling process. Secondly, for painted tracer pebbles, it is very likely that a pebble with a b-axis smaller than 32mm will be very difficult to locate visually, once it becomes integrated into the existing bed material. Pebbles with a b-axis in excess of 128mm were excluded from the experiment for practical reasons since the sediment was removed from the field site and returned following laboratory treatment.

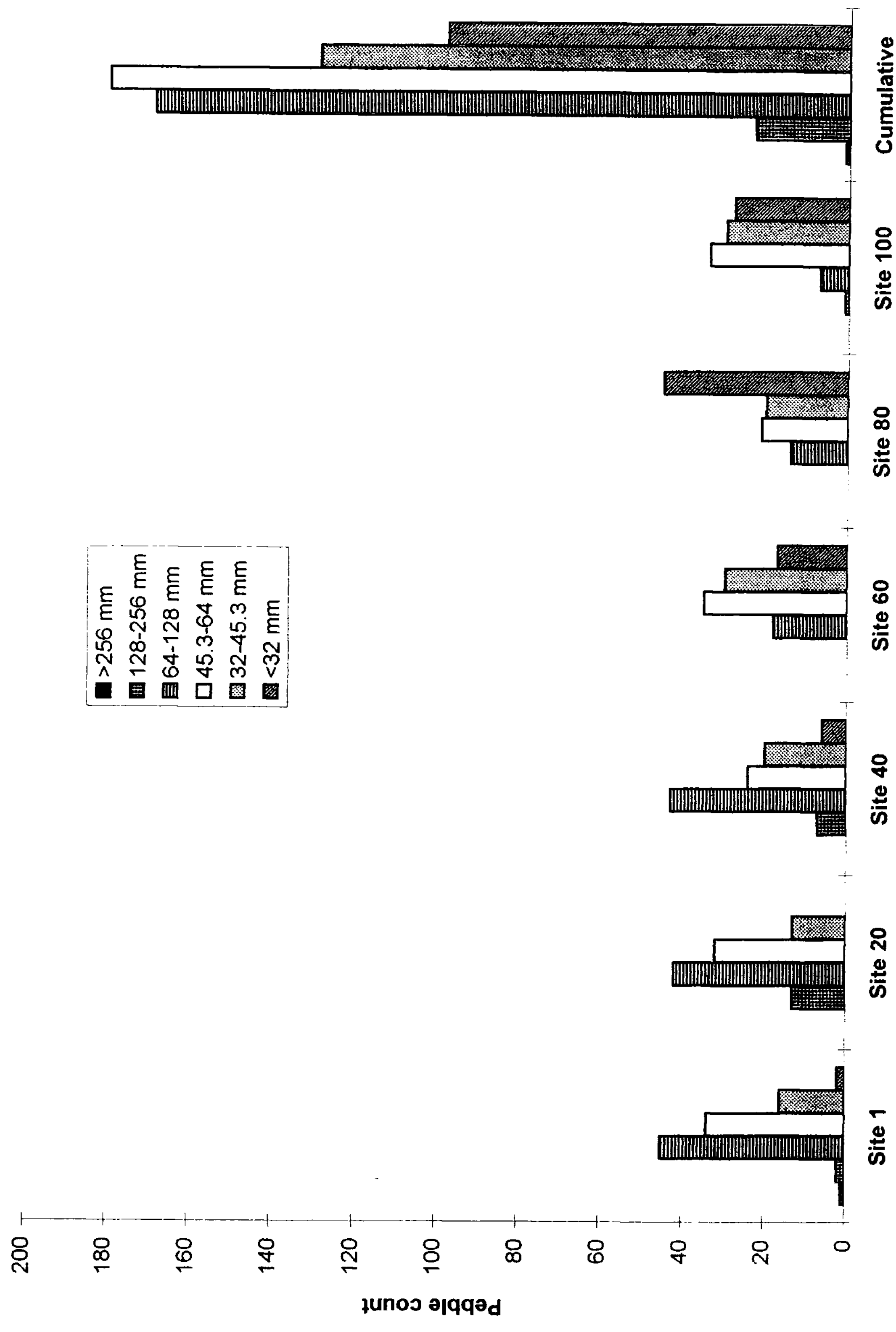


Figure 7.4 Histogram of main grain-size groups of bed material at Swinhope Burn

The painted pebble tracing technique has been used in this experiment since it is relatively inexpensive, which means that large numbers of tracers can be placed in the channel. However this method has a number of drawbacks. Searching for tracers following transport is limited to the channel bed surface since the pebbles can easily become buried following a flood event. Problems of visual identification can also arise firstly if the tracers become coated in algae or fine sediment during periods of low flow, or secondly as a result of abrasion. However, in this tracing experiment since the painted tracers remained on, or at the base of cut-banks for almost a year before they were entrained by the flood of January 8th, 1998, these problems did not arise. However, some of the pebbles had become buried within the base of the cut-banks due to trampling by cows and sheep and slow mass movements of the banks. Consequently, the tracer pebbles became integrated with the existing material of the cut-bank.

The magnetic tracing technique was used in this experiment since it is one of the most effective methods of monitoring bedload movement in upland, gravel-bed streams. The reason for this is the ability to locate both surficial and buried pebbles following a flood event using a magnetic locator, which in this experiment led to a magnetic tracer recovery rate of 78 % after one flood event. However, since the removal of pebbles from their natural location, drilling and placing magnets within pebbles is both time-consuming and costly, the number of magnetised particles in this experiment was limited to 200.

7.4 The flood of January 8th, 1998

On the 8th January, 1998, a relatively large flood event occurred within the Swinhope Burn basin. A peak over threshold value of 91.38 m^3s^{-1} was recorded on the River Wear at Stanhope, of which Swinhope Burn is a right bank tributary. A mean daily flow of 32.54 m^3s^{-1} was recorded. This compares with a peak over threshold value of 160.39 m^3s^{-1} and 49.1 m^3s^{-1} mean daily flow value for the flood in the Swinhope Burn basin which occurred on 19th and 20th February 1997 as described in the previous chapter.

Following the flood in January, 1998 the new positions of both the magnetic and painted tracer pebbles were recorded in February, 1998. Following this flood, no other high flow events occurred during the period prior to the field survey of the new tracer positions.

In the immediate post-flood period there was field evidence that both the magnetic tracers within the channel and the painted pebbles placed at lateral cut banks had been entrained by the flood. At Site 4, the peaty ledge upon which the tracers had been positioned had collapsed, due to bank undercutting, and it was evident that some of the tracers had been transported a considerable distance downstream. At Site 5, however, there was very little evidence of tracer movement. The lack of tracer movement at Site 5 may be related to the fact that tracer pebbles had become embedded in the sediment at the base of the cut-bank through animal trampling. At Sites 2 and 3, the majority of tracers had been transported further towards the channel boundary through slope processes and entrained by the higher water level during the flood, although some pebbles remained embedded in the base of the cut-bank or were within one metre of the channel margin.

7.5 Results and Discussion

The initial results of this ongoing sediment tracing experiment at Swinhope Burn are reported and discussed below. The results relate to the flood event which occurred on the 8th January, 1998. The characteristics of tracer movement in the main channel (Site 1) following the January 1998 flood are detailed in Table 7.3.

	Small 32-45.3 mm	Medium 45.3-64 mm	Large 64-128 mm
Mean	78.0	88.1	66.6
Median	78.2	84.9	47.5
Standard Deviation	51.5	61.6	51.2
Minimum	6.5	6.6	6.5
Maximum	200.8	227.8	191.0
Number moved	41	44	71
% movement	82	70	82

Table 7.3 Characteristics of main channel (Site 1) tracer movements following the January, 1998 flood, Swinhope Burn. The measurements represent distance moved downstream in metres.

Table 7.3 shows that a high percentage of within-channel tracers were entrained by the January 1998 flood, ranging between 70% for medium sized pebbles (45.3-64mm) and 82% for large (64-128mm) and small pebbles (32-45.3mm).

The mean distance travelled by the magnetic tracers ranged from 66.6 m for the largest pebbles to 88 m for the medium sizes pebbles, with the mean distance travelled by the smallest pebbles falling between the two (78 m). Standard deviation values for the different grain-size classes indicate considerable variability in the distance moved (Table 7.3). The data suggests that particles which have b-axis ranging from 45.3-64mm, once entrained are likely to be transported further. The furthest distance travelled by any tracer (228 m or over a distance of 15 cross-sections) was a medium-sized pebble. The possible reasons for this are two-fold. Firstly, particles smaller than 45.3 mm are likely to become trapped within the coarser framework gravels during transport, reducing distance travelled. Secondly particles in the range 64-128 mm are likely to be transported during brief periods of peak flow, thereby limiting the distance travelled. The dispersal of within-channel tracers in relation to pebble-size following the January, 1998 flood is illustrated in Figure 7.5. The majority of large tracer pebbles were transported relatively short distances (up to 80 m) downstream, although peaks (at 130 m and 200 m) show that some were transported much further downstream. There is a downstream decrease in distance travelled by small and medium sized tracer pebbles, and many pebbles travel relatively short distances (70 m downstream). The downstream distribution is more even, particularly for medium-sized pebbles. Peaks in distance travelled by all the tracers probably reflect patterns of deposition on channel bars on the inside of meander bends.

This discontinuous pattern of tracer deposition is illustrated in Figure 7.6 which plots the distance travelled by all tracers, irrespective of size. Tracer deposition is largely concentrated within 10-50 m of Site 1, with smaller peaks in tracer deposition occurring between 60-160 m and 190-210 m downstream.

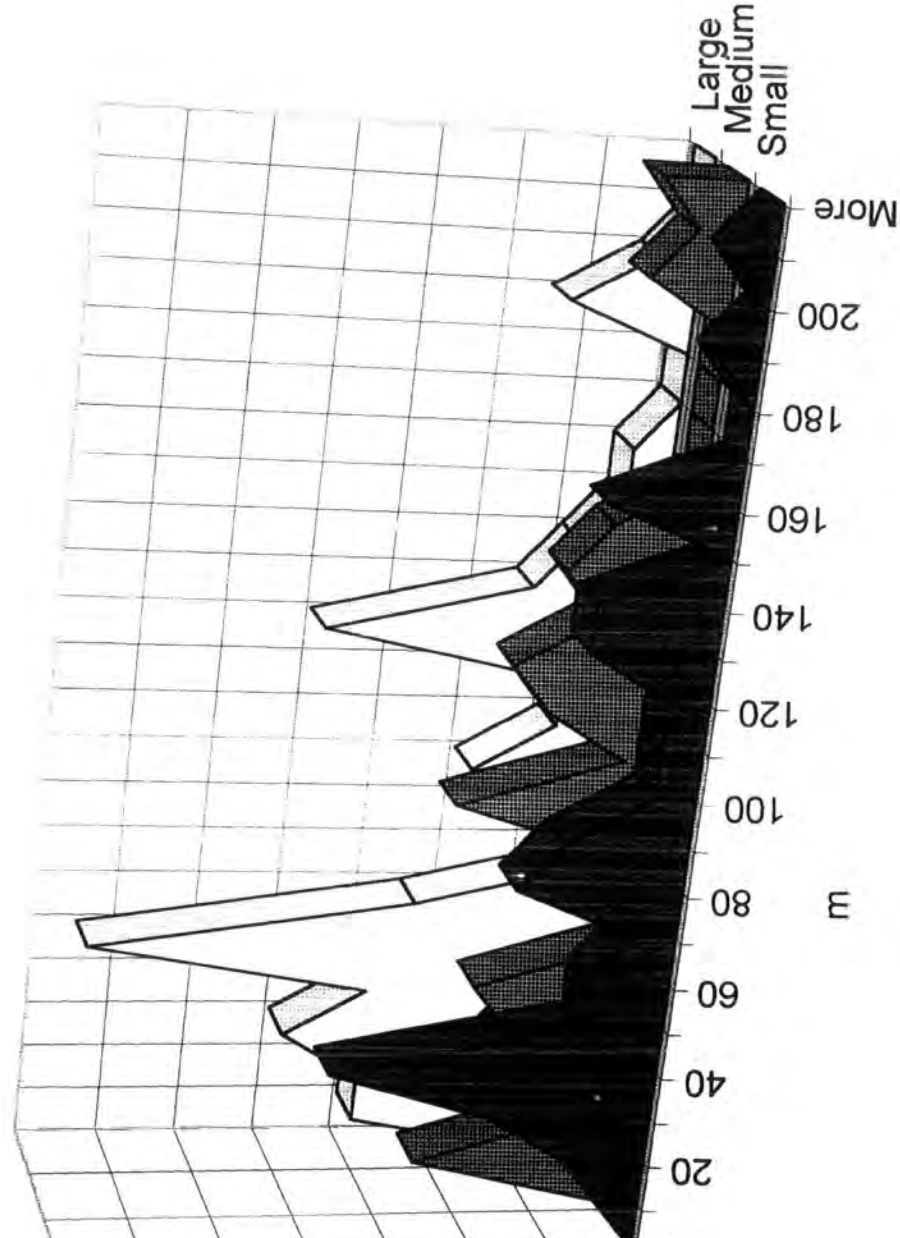


Figure 7.5 Dispersal of within-channel tracers in relation to pebble size following the January, 1998 flood

Figure 7.6 Dispersal of within-channel tracers for all grain-size classes following the January, 1998 flood

Table 7.4 shows the characteristics of tracer movements at lateral sediment input location (sites 2 to 5) following the January, 1998 flood.

	Distance Downstream →			
	Site 2	Site 3	Site 4	Site 5
Maximum movement (m)	12	82	62	1
Recovery (%)	81 (65)	90 (72)	75 (60)	107 (86)
% in main channel[1]	60	40	99	10
% moved > 1 m[1]	11	19	25	0
Notes	Slow mass movement	Slow mass movement	Bank collapse	Animal poaching

Notes:

1. Refers to % of tracers recovered.

Table 7.4 Characteristics of tracer movements at lateral sediment input locations (Sites 2 to 5) following the January 1998 flood, Swinhope Burn.

Recovery rates of painted tracer pebbles placed at lateral sediment input locations, range from 60% at Site 4 to 86% at Site 5 (Table 7.4). The low tracer recovery rate recorded at Site 4 (60%) can be explained since the highest percentage of tracers from this site were transported a distance of more than 1m. As tracers tend to become buried within the channel bed during transport by floods, the tracer is no longer visible on the surface and this reduces the tracer recovery rate of painted pebbles. Similarly, the very high rate of tracer recovery at Site 5 is understandable because none of the pebbles moved more than one metre downstream.

The downstream movement of tracer pebbles from sites 2 to 5 varies considerably. Sites 3 and 4 show the highest percentage of tracers recovered more than one metre downstream. At Site 4, 99% of tracers initially placed at the base of the cut-bank were recovered from the main channel. 25% of these were transported more than one metre from the site, the furthest tracer being recovered 62 m downstream. Bank collapse at Site 4 is largely responsible for the majority of the tracers being introduced to the channel. There is evidence to suggest that once the tracer has been moved into the main channel, it can be rapidly transported downstream. However, this largely depends on the local hydraulics and bed topography at the base of the cut-bank. At Site 4, bank undercutting during the January, 1998 flood led to bank collapse with the tracers being deposited immediately upstream of a pool on a meander bend apex. Transport through this pool explains why a high proportion of tracers were moved more than one metre from Site 4.

At Sites 2 and 3 fewer tracers (11% and 19% respectively) were recovered more than one metre from the channel, although at Site 3, a maximum distance of 82 m is recorded for one tracer. These lateral input sites are located on relatively straight sections of channel and the tracers were placed on low angle slopes. This explains why the percentage of tracers recovered from the main channel is relatively low compared with Site 4. Although slow mass movement, particularly at Site 2 encourages the tracers to be introduced to the channel (60%), transport distances during high flow events would be relatively small due to the local reach characteristics. At Site 5, the tracers were initially placed on a low angle slope at the base of a cut-bank on a straight section of channel. The probability of entrainment at this location is relatively low. Additionally, poaching (through trampling by cattle and sheep) led to the burial of a large number of tracer pebbles which is another reason for the lack of tracer movement at Site 5.

Figure 7.7 summarises the percentage and distance of tracers moved in the main channel compared with lateral sediment inputs, following the January, 1998 flood. The upper graph shows that main channel sediment inputs travelled much greater distances than the lateral sediment inputs during the January, 1998 flood. Movement of tracers placed as lateral channel inputs (Sites 2 and 5) was small due to their positioning on a straight section of channel at the base of low angle slopes where the probability of entrainment even during high flows was low. In contrast the increased distance travelled by tracers originating from Sites 3 and 4 is largely in response to their positioning on steeper slopes at meander bend locations.

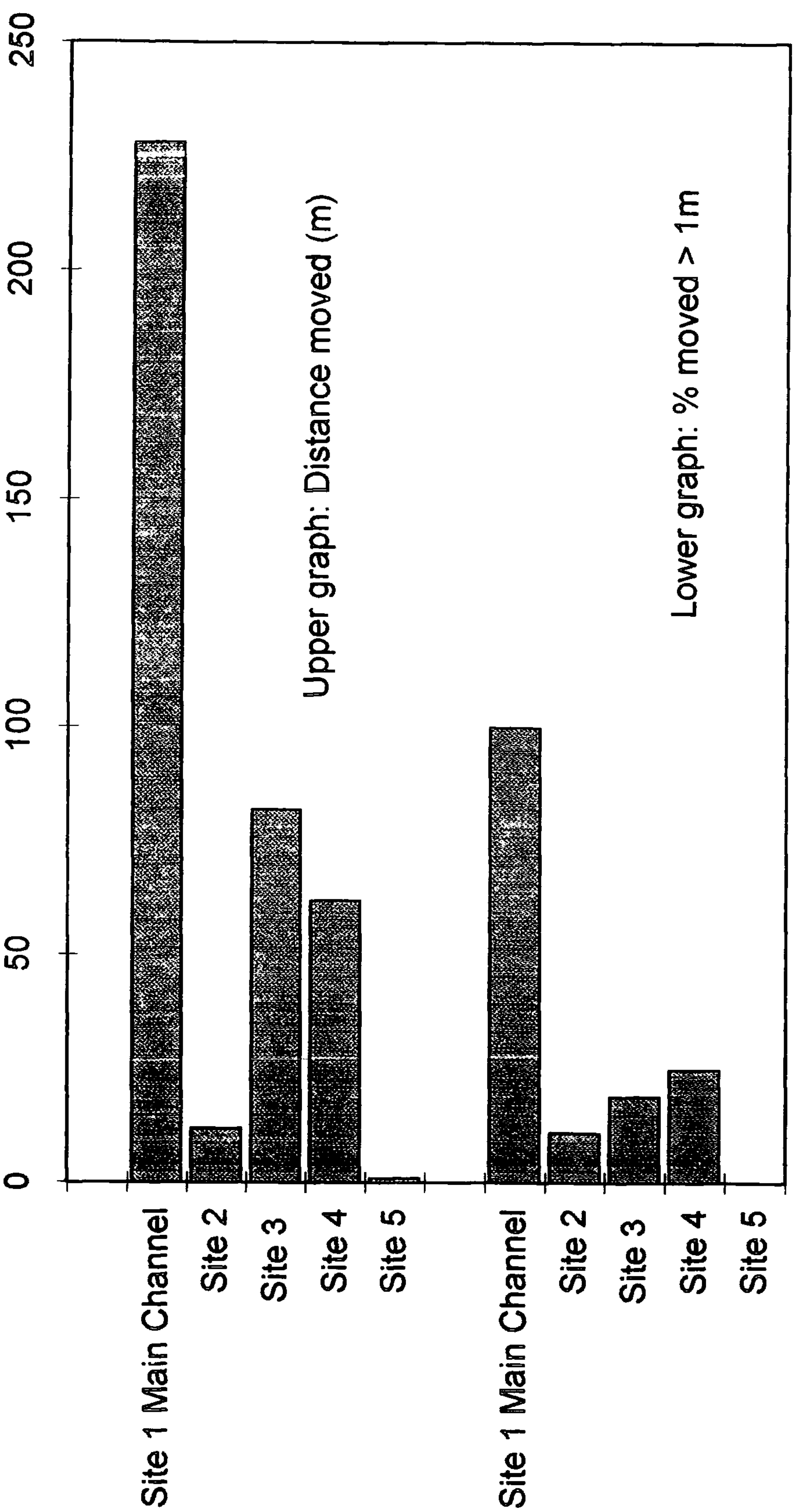


Figure 7.7 The percentage and distance of tracers moved in the main channel compared with lateral sediment inputs following the January, 1998 flood

The lower graph (Figure 7.7) shows a similar pattern whereby the majority of tracers placed within the main channel were transported more than 1 m downstream, whereas most of the tracers placed as lateral inputs were recovered within 1 m of the site.

7.6 Preliminary Conclusions

Following the January, 1998 flood event, a large number of within-channel magnetic tracers were entrained and transported with the mean distance of travel recorded at 77.6 m. The limited evidence suggests that there is no strong pattern in particle size and pebble entrainment and transport in the study reach. However, the largest pebbles (64-128mm) were transported the shortest distance. Small pebbles (32-45.3mm) were transported shorter distances than pebbles of medium size (45.3-64mm) which were transported the furthest. This pattern possibly reflects the importance of bed structure whereby small particles become trapped within the coarse framework gravels reducing travel lengths. Tracer deposition seems to be largely controlled by local conditions. The discontinuous pattern of tracer deposition identified on Figure 7.5, demonstrates the importance of deposition on bar surfaces on the inside of meander bends. The results suggest that there is active transport of bedload through the main channel during flood events

Although, it is appreciated that lateral inputs of coarse sediment influence patterns of downstream fining (Knighton, 1980; Ashworth and Ferguson, 1989; Ferguson et al, 1996; Wathen et al, 1997) and the development of cross-sectional geometry (Milne, 1979) in upland gravel-bed streams, the importance of lateral sediment inputs to the channel during flood events has not been studied in great detail. This study tests the hypothesis that lateral sediment inputs are limited and contribute minor amounts of sediment to the stream during flood events.

The movement of tracers placed as lateral sediment inputs on eroding and cut-banks along the channel was observed following the January, 1998 flood. As expected, only a small proportion of tracers placed at Sites 2 to 5 were moved during the flood. However, a major factor governing the movement of coarse sediment from cut banks to the main channel is the local channel geometry. Sites located on meander bends (Site 3 and particularly Site 4) show higher rates of tracer entrainment and transport. The maximum distance recorded for a tracer placed as a lateral input was 83 m. Once coarse sediment is introduced to the channel either through slow mass movement (Site 2), or through bank undercutting and collapse (Site 4) within-channel transport can be rapid and will approach the rate of sediment transport characteristic of the throughput load.

When interpreting these results it should be kept in mind that there is a likely reduction in sediment transport capacity down the reach. The downstream reduction in channel slope and available stream power (Chapter 5) will reduce the sediment transport capacity towards the end of the reach. Therefore entrainment of sediment from lateral sources will tend to generally decline from the top to the bottom of the reach although local conditions will remain significant.

A magnetic tracer recovery rate of 78% after one flood event demonstrates the utility of magnetic tracer techniques in upland gravel-bed streams. The technique was a particularly useful one since over half of the tracers were buried in the surface layers of the bed. For this reason, recovery rates of painted tracer pebbles placed as lateral sediments inputs were much lower.

CHAPTER 8

SUMMARY AND CONCLUSIONS

This chapter provides a summary of the main discussion points of the thesis, and illustrates how the aims and objectives outlined in Chapter 1 have been achieved. Secondly, the major conclusions are highlighted and finally possible areas of future research are identified.

8.1 Achievement of aims and objectives

The aim of this thesis was to identify the importance of flood events in contemporary and historic river channel change in Swinhope Burn, Weardale. This was achieved in Chapters 4 to 7.

The first objective was to identify the role of flood events in determining the nature and extent of historic river planform change from 1815 to the present using historical maps and air photographs and to identify possible causes of river planform change using field evidence and local historic flood documentation. A series of eight historical maps and one air photograph enabled changes in channel position and planform to be documented from the Inclosure Plan of 1815 to the present. It was concluded that the study reach has been remarkably stable over the last 180 years. However, the Tithe Map of 1844 shows that there was a temporary but dramatic change in channel pattern from a single thread meandering channel to a relatively straight channel with braiding in the upper reaches. Owing to the lack of historic flood records or rainfall and flow data for this isolated upland catchment it is difficult to identify the precise cause of the observed channel planform change. However, documentation relating to a flood in September, 1824 is of prime importance in suggesting the cause of the observed planform change. Swinhope Burn was identified as being the most severely affected area, with the 1824 storm being 'most terrific at Swinhope in Weardale, a torrent of rain having fallen which swept away Swinhope stone bridge' (Egglestone, 1874). However, a series of high magnitude floods affecting the River Wear coincides with the period during which Swinhopehead mine, located 2 km upstream of the study reach, was active. Using historical map evidence and flood and mining documentation it is concluded that it is probable that a combination of increased inputs of coarse sediment generated by a series of four major flood events from 1822 to 1828 and upstream mining activities during the period 1823 to 1846 led to the observed changes in channel pattern evident on the 1844 Tithe map. In the British uplands, historic changes in channel pattern from meandering to low sinuosity and braided channels have been linked with inputs of coarse sediment generated by mining activities (Lewin et al, 1977, 1983), floods (Ferguson and Werritty, 1980; Milne 1982a) and a combination of the two (Macklin, 1986; Macklin and Aspinall, 1986).

The second objective of this thesis is to examine downstream variations in channel form and grain-size distribution of bed material in response to flood events. In order to achieve this objective a network of 101 monumented cross-sections were pegged on the right and left banks of a 1.4 km long reach of Swinhope Burn. A grain-size survey, in which the 20 largest particles were sampled at each of the 101 cross-sections, was carried out both prior to and following a major flood event. Downstream changes in cross-sectional form and grain-size of bed material in relation to variations in channel slope, bank cohesion and bed topography were examined, which provided an insight into the structure of an upland stream. It was concluded that the irregular cross-sectional geometry of Swinhope Burn, a characteristic of many upland gravel-bed streams, and channel planform development results largely from downstream variations in bank cohesion, bed topography and the presence of bars and streamside scars. A large flood which occurred on 19th and 20th February 1997 provided an opportunity to examine the response of an upland gravel-bed stream to a specific event. Detailed analysis comparing the pre and post flood channel at Swinhope Burn identified that even though the February 1997 flood was the fourth highest peak flow recorded on the River Wear since 1958, and had been overbank along almost the total length of Swinhope Burn, very little channel change had resulted. Channel erosion was slight and on the same order of magnitude as channel deposition. There is evidence to suggest that a sediment wave moved into the upper reaches of Swinhope Burn during the flood of February 1997, causing aggradation and fining of the existing bed material. This appears to have come to a standstill mid-way along the reach forming a slug-front. However, grain-size surveys indicate that there is no significant difference between the pre and post flood grain-size distribution. It is clear from these results that the contemporary response of the channel to a major flood event is through vertical adjustments within the channel bed rather than lateral adjustment through the migration of meander bends. However, it must be noted that channel change along Swinhope Burn following the flood was highly localised, and because the study reach has a low gradient in contrast to other similar upland streams, comparison with other studies may require additional qualification.

The final objective of this thesis was to examine both within reach and through reach sediment dynamics during flood events in order to determine whether Swinhope Burn has a dynamic bed with an active sediment throughput or whether it has a predominantly static grain-size distribution. A sediment tracing experiment which was designed to determine the importance of sediment exchanges between the bed and lateral inputs from eroding banks and bluffs demonstrates the importance of within channel movements. A total of 700 tracer pebbles were used in the sediment tracing experiment. 200 magnetic tracer pebbles were implanted within the channel bed at the head of the study reach and 125 colour-coded tracer pebbles were placed as lateral inputs at four sites downstream on eroding banks and bluffs. Following a relatively large flood on 8th January, 1998, a large number of the within-channel magnetic tracers were entrained and subsequently located. This yielded a tracer recovery rate of 78%, over half of which were buried beneath the bed surface. However, in contrast very few of the painted tracer pebbles originally located at four cut-banks along the channel were entrained. Many were buried *in-situ* at the base of the cut-bank either due to animal trampling or as a result of slow mass movements of soil downslope during wet periods or after heavy rainfall. It was concluded that although there is active transport of bedload during flood events, the bed material grain-size population varies only locally and lateral sediment inputs are limited with bank erosion contributing minor amounts of coarse sediment to the stream.

8.2 Conclusions

The lack of lateral change in the channel in response to the February 1997 flood, coupled with the lack of lateral sediment exchanges to the channel during the January, 1998 flood is substantiated by the remarkable channel planform stability over a period of 180 years identified on historical maps and air photographs. The majority of channel change which occurs in response to high flow events is through vertical adjustment of the channel bed resulting from erosion and aggradation. During flood events sediment is transported through the reach with minimal lateral coarse sediment inputs from cut-banks. Adjustments in bed material size are related to local conditions.

Analysis of historical maps and air photographs of the study reach have shown that, with the exception of a temporary switch from a meandering, single thread channel to a relatively straight one, probably induced by a combination of a series of large flood events and inputs of coarse sediment from upstream mining activities, there has been very little change in the channel planform of Swinhope Burn over a period of 180 years. Field survey of Swinhope Burn over a period of two years, and analysis following two large floods has suggested that although the majority of flood-induced channel change occurred through vertical adjustments in the channel bed with very limited lateral adjustment, changes along the channel as a whole have been very small.

The low gradient of the study reach at Swinhope Burn is a key factor controlling both historic and contemporary channel response to flood events. The partial closure of the valley system by the Greenly Hills Moraine, a legacy of the Devensian glaciation, has produced a distinct step in the long profile of Swinhope Burn which has led to Swinhope Bottoms becoming a 'sedimentation zone' because of an imposed local base level. Low channel gradient along the length of the study reach and associated low stream power reduce the ability for coarse sediment transport during flood events which restricts morphological channel change. Coupled with a well-vegetated floodplain this means the channel is stable even during high magnitude overbank flooding. This explains the lack of channel response to contemporary and historic flood events and the remarkable stability of channel planform at Swinhope Burn. It appears that any major changes in channel planform, such as a transformation from a single meandering channel to a relatively straight one with a braid bar, requires the coincidence of a series of large floods in quick succession with a large change in external influences, in this case, episodic inputs of coarse sediment to the channel instigated by upstream mining activities.

8.3 Future research

In order to make firmer conclusions on the role of historic metal mining in causing the observed river channel metamorphosis on the Tithe Map of 1844, the abandoned channel preserved within the floodplain at the head of the study reach (NY 897 348) needs to be cored and samples collected for heavy metal analysis. This will help identify whether inputs of coarse, contaminated mining waste could have caused the channel to straighten and cause a braid bar to form in the upper reaches.

Likewise, sub-surface profiling of Swinhope Bottoms using Ground Penetrating Radar (GPR) and a programme of coring will allow the geometry of the basin and its alluvial architecture to be mapped. This survey would allow the depth of alluvium within the basin to be determined and may provide evidence of lake clays, possibly confirming that a temporary lake had existed at this location during the late Devensian glaciation as proposed by Moore (1994) in his research on the glacial geology and geomorphology of Weardale. A knowledge of the depth of alluvium within the basin may also provide an indication of the sediment budget and rate of deposition within the sedimentation zone of Swinhope Bottoms.

Continued monitoring of contemporary floods at Swinhope Burn in terms of both sediment dynamics and adjustments in channel form is desirable in order to determine whether high magnitude floods alone are capable of producing high rates of coarse bedload transport and major channel change or whether changes in other external factors, such as land-use, are required for the reach to become unstable. A gauging site will be established at the top of the reach (NY 898 347) in order to monitor streamflow levels during floods.

In order to put research at Swinhope Burn into a broader context, assessing the role of an imposed local base-level on downstream fining of bed material and channel planform stability in other small upland streams is required as this may provide an insight into channel planform stability over the historical period. Whereas many studies have concentrated on the dynamic nature of upland gravel-bed streams, the reasons for channel planform stability have been largely neglected. In addition, the timescales over which river channel changes in response to flood events are monitored need to be extended to include the historical period in order that observations of contemporary channel processes can be placed in a wider context. Equally, continued monitoring of river channel changes during the post-flood period is essential. If post-flood recovery is rapid and widespread following a large flood event, the geomorphic significance of the flood will be less. Conversely, if post-flood recovery is slow or highly localised the significance of the flood is far greater regardless of the magnitude and frequency of the event.

APPENDIX A

Details of historical map and plan sources used to reconstruct the history of channel change at Swinhope Burn, Upper Weardale.

Map or Plan	Scale	Published date	Mapping date
Weardale Inclosure Plan (John Bell)	10 inch to mile	1815	
Tithe Map	20 inch to mile	1844	1842
Ordnance Survey County Series First Edition	25 inch to mile	1861	1856 to 1858
Ordnance Survey County Series Second Edition	25 inch to mile	1896	1856 revised 1896
Ordnance Survey County Series Third Edition	25 inch to mile	1921	1856 revised 1896, 1919
Ordnance Survey Provisional Edition	1:10,560	1953	1940
Ordnance Survey National Sheet Lines *	1:25,000	1960	1896 to 1919 revised 1951
Outdoor Leisure Map North Pennines	1:25,000	1995	1977 to 1982
Ordnance Survey Digital Raster Data	1:10,000	Current	
County Durham Air Photograph (Aerofilms, 7/9/91)**	1:10,000	1991	1991

* Not plotted in Figure 4.2

** In addition to the 1991 Air Photograph there are 11 other Air Photographs covering the study reach. They are in various formats taken between 1951 and 1995.

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