LATE QUATERNARY VALLEY FILL SEDIMENTS IN THE RIVER TYNE VALLEY: UNDERSTANDING LATE DEVENSIAN DEGLACIATION AND EARLY POSTGLACIAL RESPONSE IN NORTHERN ENGLAND

BY

Lynda Yorke

M.Sc., UNIVERSITY OF DURHAM, 2000

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF

THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

Department of Geography

THE UNIVERSITY OF HULL

January 2008

Abstract

This thesis reconstructs part of the deglacial and early post-glacial history of the River Tyne Valley, Northumberland. Data has been gathered through description and interpretation of sedimentary sequences and stratigraphies from quarry and cut-bank exposure, the development of a geochronology was attempted, based on optically stimulated luminescence (OSL), and broad-scale geomorphological mapping using traditional field mapping and appropriately scaled digital elevation models from NEXTMap Britain.

A complex suite of landform-sediment assemblages indicates both active and stagnant ice operated in the Tyne Valley over a prolonged period during deglaciation. The sedimentary and landform sequence records the ice stagnation and disintegration signature of the Tyne Valley ice in response to ice starvation as the major Irish Sea ice stream capture the ice source feeding the Tyne Valley ice as it responded to millennial scale climate change. Ice advance down the Northumberland coast resulted in subsequent impoundment of the lowlands, which were ice-free, leading to the development of glacial lakes. The most extensive being Glacial Lake Wear. Thus, two-phases of activity, stagnation and impoundment, best characterise deglaciation in the Tyne Valley. Problems dating quartz samples with OSL, (low sensitivity and weak fast component), has meant the timing of ice stagnation remains unknown.

Following deglaciation, there was ~10m of down-cutting from the Devensian glacigenic surface. Three terrace surfaces lie at ~20, 10 and 8m above present river level, and formed between ~11-7ka cal. BP when the river experienced a period of extensive incision and reworking. Valley floor refilling at the start of the Holocene was punctuated by two cycles of incision and aggradation, driven by glacio-isostatic uplift, climatic change (sediment supply and discharge), readily available (glacial) sediments and continued paraglaciation.

The thesis contributes new detailed sedimentological and morphological data, and has re-evaluated existing data sets and interpretations. A number of sites have been investigated in very close detail, and the sedimentological analyses provide a much better understanding of their formational environment than morphological studies alone can do. The research in the Tyne Valley contributes to the growing body of work carried out by the extensive morphological mapping programs, with the detailed sedimentological and stratigraphic data vital for ground truthing remotely-sensed landform interpretations. The story of deglaciation in Britain is a complex one and the work here illustrates that we are far from understanding the behaviour of the ice during the last glacial period, and certainly there is not a one model fits all solution.

Acknowledgements

"Through dangers untold and hardships unnumbered I have fought my way here…" (*Labyrinth*, 1986).

I would like to thank Dr Barbara Rumsby, Dr Stuart McLelland and Dick Middleton for supervisory support. Without Barbara's words of encouragement and continued belief in me, this last year really would have been hell. Thank you for being a great supervisor and a friend, I shall miss our supervisory meetings in the coffee shop. To Dr Ian Fuller (Massey University) who first inspired me as an undergraduate to undertake this journey into academia, and for his continued support despite being located on the other side of the world, I thank you. To Martin Charlton (NUIM), I owe thanks for providing unstinting help, advice and support with spatial analyses, and guiding me through numerous problems that arose. Throughout this thesis I have discussed ideas with many different people, notably Dr Wish Mitchell and Dr Dave Roberts (UoDurham), Dr Andy Howard (UoBirmingham), and Dr Richard Chiverrell (UoLiverpool), I thank you all for your time, suggestions and comments. Finally, thanks go to Sandra Mather (UoLiverpool) for re-drafting a number of my diagrams and converting my scribbles into professional drawings.

I am grateful to The University of Hull who provided a GTA Scholarship and the Department of Geography for supporting my research agenda. The QRA and RLAHA (UoOxford) for funding OSL dates. Dr Jean-Luc Schwenninger (UoOxford) and Prof. Andreas Lang (UoLiverpool) are thanked for carrying out the OSL dating. Thanks to Dr Dirk Hoffman (UoBristol) for undertaking the U-series dating and Prof Michel Lamothe (Université du Québec à Montréal) for undertaking IRSL dating, both of whom undertook this work freely. I thank the North Eastern Geological Society and the Cumberland Geological Society (Angela Marchant Award) for supporting my research and my conference trip to Ottawa, Canada in 2007.

Cemex (formerly RMC Aggregates) and the numerous landowners (The Straker Estate and Mr. R. Johnson, West Warmley Farm, Fourstones, and Mr. and Mrs. A. Harrison, Eltringham Farm House, Stocksfield) along the Tyne Valley who granted access to work on the sites are thanked, without their kind permissions none of this work would have been possible.

The generosity of friends and colleagues, especially Sarah Fahle, Kay McManus, Simon Tate and Brendan Murphy who have provided assistance and amusement in the field, is gratefully acknowledged. A big thanks to The Bradley Gardens Nursery teashop, Crawcrook, and The Watling coffee shop, Corbridge, for supplying me with tea and cake, and a means to bribe people to assist me in the field. Field work should always involve a good tea shop!

To my friends, near and far, new and old, and to my parents I thank you for keeping me sane and reminding of what is important in life. I would like to give special thanks to Sally Hopley and Steph Jones who have coped with the ups and downs of the thesis during the last three years, and to Ben Gilna who got me over the finishing line.

Finally, never trust your supervisor (you know who you are Stuart) when they tell you bulls can't run down river terraces or jump barb-wire fences to terrorise you on fieldwork….they can!

List of abbreviations

APs - Aerial Photographs

AMS - Accelerator Mass Spectrometry

BGS - British Geological Survey

BIS - British Ice Sheet

BIIS - British Irish Ice Sheet

BP - Before Present, taken to be 1950

GCPs - Ground Control Points

DEM - Digital Elevation Model

DSM - Digital Surface Model

DTM - Digital Terrain Model

ESRI - Environmental Systems Research Institute

GIS - Geographical Information Systems

GPS - Global Positioning System

IMAU - Industrial Mineral Assessment Unit

LiDAR - Light Detection and Ranging

OD - Ordnance Datum

OSGB36 - Ordnance Survey Great Britain 1936

OSL - Optically Stimulated Luminescence

MIS - Marine Isotope Stage

PCA - Principal Components Analysis

Table of Contents

List of Figures

List of Tables

Chapter One

Introduction and Research Objectives

1.1 Rationale

The aim of this thesis is to construct a conceptual model for deglaciation of northeast England that is comparable to existing North Atlantic Seaboard models elsewhere and will enhance understanding of the behaviour of the eastern margin of the last British ice sheet (BIS) during deglaciation. The project will provide new insights into outwash and deglacial sediments and landforms, ice sheet dynamics and marginal fluctuations and, crucially, it will provide a chronological framework that allows the sequence to be integrated and associated with the late Quaternary stratigraphic and climatic record.

In the last few years there has been renewed and increased interest in understanding the deglaciation of the last BIS (e.g. QRA, 2007). Aligned with this has been the recognition that river basin response immediately following deglaciation is affected by the boundary conditions set during deglaciation and continues to be affected by regional and global change during the transition from glacial to interglacial conditions (Macklin and Lewin, 1997). Finally, the relationship of the ice sheet (and the post-glacial environment) to sub-millennial climate variations in the North Atlantic region during the last deglaciation and Younger Dryas (Loch Lomond) is of increasing interest (e.g. Lewin and Macklin 2003; McCabe and Clark 2003). The BIS was independent of the larger global ice sheets (i.e. Scandinavian and Laurentide) during the last glaciation and it is likely to be more sensitive to climate change in the North Atlantic than those larger ice sheets.

The Tyne Basin, northeast England was chosen for this study because despite the wealth of antiquarian literature, which exists in terms of regional glacial history, very little has been done in terms of detailed field research since the early 20th century. Quaternary science has moved on significantly since that time. One of the current paradigms in geomorphology is the study of large-scale landforms and a return to regional landform analysis (Baker, 2007). With advances in our understanding of Quaternary environments, global climate models and high-resolution climatic records (e.g. GISP, GRIP), the theoretical framework for undertaking large-scale analyses is in place (Baker, 2007). Furthermore, technological advances in terms of dating methods (e.g. AMS ^{14}C, optically stimulated luminescence, amino-acid racemisation, cosmogenics) and remotely sensed datasets (e.g. satellite, radar) allow investigation at both the large temporal and spatial scales. Therefore, given the current impetus in this research field, this project provides a timely opportunity to re-evaluate the deglacial history in Britain through broad-scale geomorphological mapping using appropriately scaled digital elevation models (DEM), collection of field data in recently exposed landform-sediment assemblages[*], and a chronological framework can be developed through both physical and geochemical techniques.

The main corpus of antiquarian research is restricted to reports by a number of workers, the most prolific being Lebour (1878, 1884, 1889, 1893), which document the existence of both glacigenic and post-glacial landforms and in many cases provide detailed accounts of the sedimentary sequences underlying them. Later publications by a few authors such as Francis (1970), Lunn (1980) and Douglas (1991) are based on reviews of earlier workers, which demonstrate there has been little new fieldwork in the intervening period. The most recent review of UK regional glacial landforms by Clark

[*] Thomas (1989) defines landform-sediment assemblages as a mappable unit, on a scale of 1:50000, in which relatively homogeneous morphological, stratigraphical and lithological characteristics occur.

et al. (2004a) and Evans *et al.* (2005) has highlighted the lack of empirical data available in northern England. Evans *et al.* (*op cit.*) illustrated that where data does exist it is very general, for example, maps indicate the presence of drumlins by a symbol rather than accurately delimiting the extent of each landform in the field. However, the key issue they raise is the fact that, over large areas of northern England, no maps of regional glacial landforms exist.

In terms of the deglacial history in northeast England, with the exception of studies in the Cheviots (Clapperton 1971; Douglas and Harrison 1985) and on the Durham coast (Teasdale and Hughes, 1999), there has been little attempt to understand the landform-sediment assemblages and none that integrates them into the North Atlantic climatic record. The accepted general model for deglaciation in northeastern England is wasting of stagnant and dead ice *in situ* (cf. Beaumont 1968; Clark 1970; Douglas 1991; Lunn 1980; Mills and Holliday 1998). However, confusion still exists in terms of understanding the precise mechanisms of sedimentation and the evolution of the landforms, and currently, no single model exists that explains the evidence with any satisfaction (Teasdale and Hughes, 1999). One of the key aims of this study is to clarify and understand the landform-sediment assemblages in terms of their genesis and to integrate the sequence into a model for deglaciation.

In the first part of this chapter, a review of the regional glacial history is given, the aim of it is to set out the recent geological history of the research area and to highlight the current situation regarding the landform-sediment assemblages and previous interpretations, and to illustrate the existing regional chronological framework. Through this review, the key issues and deficiencies of the current situation (summarised above) are made apparent. The second part of this chapter outlines the research objects and the hypotheses to be tested; finally, two conceptual models are proposed to represent likely depositional and/or formational conditions which resulted in the development of the landform-sediment assemblages, and which will be tested in this thesis.

1.2 Regional Glacial History

The last glacial advance of the Late Devensian (δO^{18} Stage 2) occurred during the Dimlington Stadial between 26 and 13ka cal. BP (Rose, 1985). The BIS consisted of lowland and upland centres of ice dispersal that were the main driving forces of ice sheet configuration and flow (Bowen *et al*., 2002).

1.2.1 Sources of ice, flow direction, extent and thickness

In northern England, the main ice accumulation zones were situated to the north and west of the study area, with two small localised ice caps developed in both the Cheviot Hills (cf. Clapperton, 1970) and North Pennines (cf. Catt 1991a,b). Recent palaeo-ice stream modelling (cf. Boulton and Hagdorn 2006, 2007; Hubbard *et al*. 2007) confirms the existence of a Tyne Valley ice stream and the development of ice over Northumberland but these models are unable to predict the development/existence of the smaller ice caps. Whilst models exist, palaeo-ice stream maps for northern England have not been published, although work is currently being undertaken by the University of Sheffield. However, evidence of ice flow direction can be seen through glacial erosion, occasional striae and the analyses of clast fabric and lithologies within the diamicton. The accentuation and the direction of strike of sandstone and Whin Sill escarpments in the South Tyne Valley indicate eastward ice movement (Clark, 1970) (see Figure 5.27A, B). Frost and Holliday (1980) map an extensive drumlin field in the North Tyne Valley, which indicates a southeasterly ice flow (stream) direction down the North Tyne towards the main Tyne Valley. Young *et al*. (2002) make reference to some drumlin-like features in the Morpeth district (~26km north of Newcastle) that have a general elongation in a west-east direction. The analyses of the diamicton, indicates that western ice comprised coalescing ice streams from the northern Lake District and Southwest Scotland (Galloway) flowing east-south-eastward *via* the Tyne Gap (the topographic depression connecting the [now] Eden and Tyne Valleys) and into the South Tyne Valley. Ice movement was slightly modified under the influence of structural control, but it essentially flowed eastward, parallel to the strike of the valley, towards Newcastle but not beyond as it encountered an ice stream flowing parallel to the coast and was deflected southwards (Mills and Holliday, 1998). The coastal ice stream originated from the northern ice accumulation area. Ice flowed east from the Southern Uplands of Scotland, along the River Tweed Valley and south east out of the Cheviot Hills, coalescing and flowing southwards, parallel to the Northumberland coast (Dwerryhouse 1902; Raistrick 1931; Taylor *et al*. 1971; Lunn 1980; Mills and Holliday 1998; Smith 1994). As a consequence of these ice streams, northern England was completely inundated by ice and thus the Tyne basin lies entirely within the limits of the last BIS (Figure 1.1).

The surface elevation of the ice sheet declined away from the source area and modelling of the last BIS by Boulton *et al.* (1977, 1985) suggests the upper surface of the ice probably lay at between 750 and 1000m above OD over the Tyne Valley. In view of the effects of isostatic depression, Smith (1994) estimated ice thickness in the Wear lowlands to be ca. 1km; ice thickness in the lower Tyne Valley was probably of similar or slightly greater thickness (Mills and Holliday, 1998).

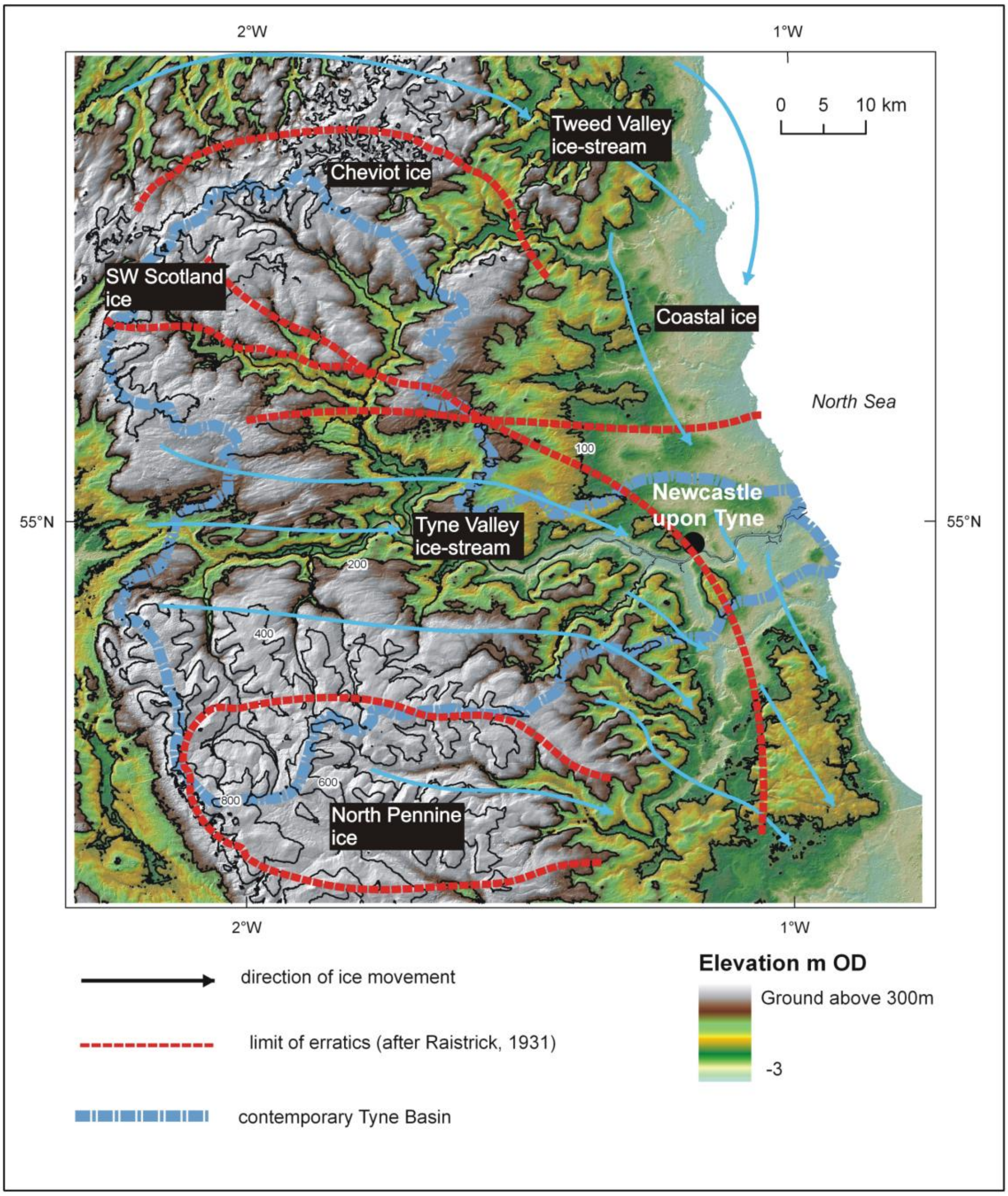


FIGURE 1.1: Map of northeast England showing the direction (blue arrows) of the ice-stream flow from source areas, and the limit of clast erratics (dotted red line) associated with the ice-streams, which potentially delimit the spatial extent of the ice-stream watersheds (after Raistrick, 1931). (Base elevation map is 25m DEM downloaded from www.landmap.ac.uk). Contour heights on the map are given in m OD.

In terms of the spatial extent of the last BIS, its southern margin is still poorly delimited due to the absence of clear end moraines, but it is generally accepted that it extended down to the north Norfolk coast in the east (Clark *et al*., 2004b) (Figure 1.2). The York and Escrick moraines in the Vale of York (Gaunt, 1976) are usually taken to mark the maximum ice limit, although there has been debate as to whether a lobe of ice extended to Doncaster (cf. Clark *et al*., 2004b). The tills of Lincolnshire and Yorkshire mark the limit of the Devensian ice and are dated to the Dimlington Stadial (Penny *et al*. 1969; Rose 1985; Clark *et al*., 2004b). In northeast England, ice marginal positions have been extrapolated from work by Sissons (1964, 1967) and West (1963, 1967) in the borders; however, nothing has been attempted to verify these. Clark (1970) speculated on the position of the landward extent of the coastal ice sheet in this area, although it is unclear what his evidence for it is. The margin essentially follows the boundary of the high ground and the coastal plain from Berwick and extending southwards to the west of Newcastle upon Tyne (Figure 1.3). However, to date, there are no actual recorded ice margin locations in northeast England.

1.2.2 Deglaciation

Bowen *et al.* (2002), in a general overview of the last BIS, suggested deglaciation was characterised by shrinking ice sheet margins that contracted back to most of the original centres of ice dispersion. Recently, McCabe and Clark (2003) and McCabe *et al*. (2007) have produced revised models for deglaciation of the last BIS from the Irish Sea Basin, northwest Ireland and eastern Scotland. McCabe *et al*. (2007) present evidence that the onset of deglaciation was earlier than previously thought in eastern Scotland (cf. Hall and Jarvis, 1989), with initial deglaciation of the area before 21.0ka cal. BP and two subsequent ice-marginal fluctuations at <20.2ka cal. BP (possibly 18.2ka cal.

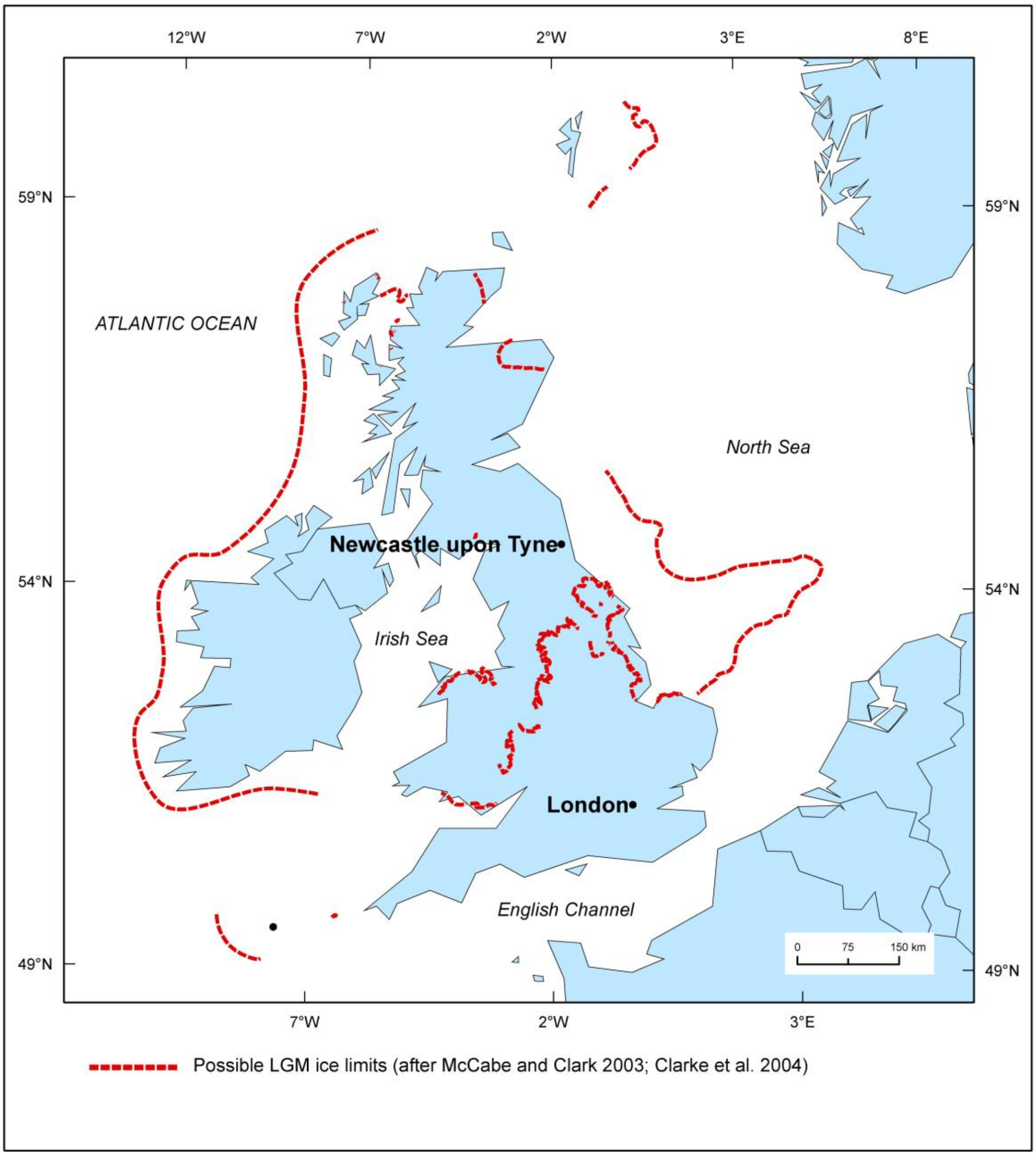


FIGURE 1.2: Map of UK and Ireland showing the accepted limits of the last British-Irish ice sheet (source: BRITICE, 2004).

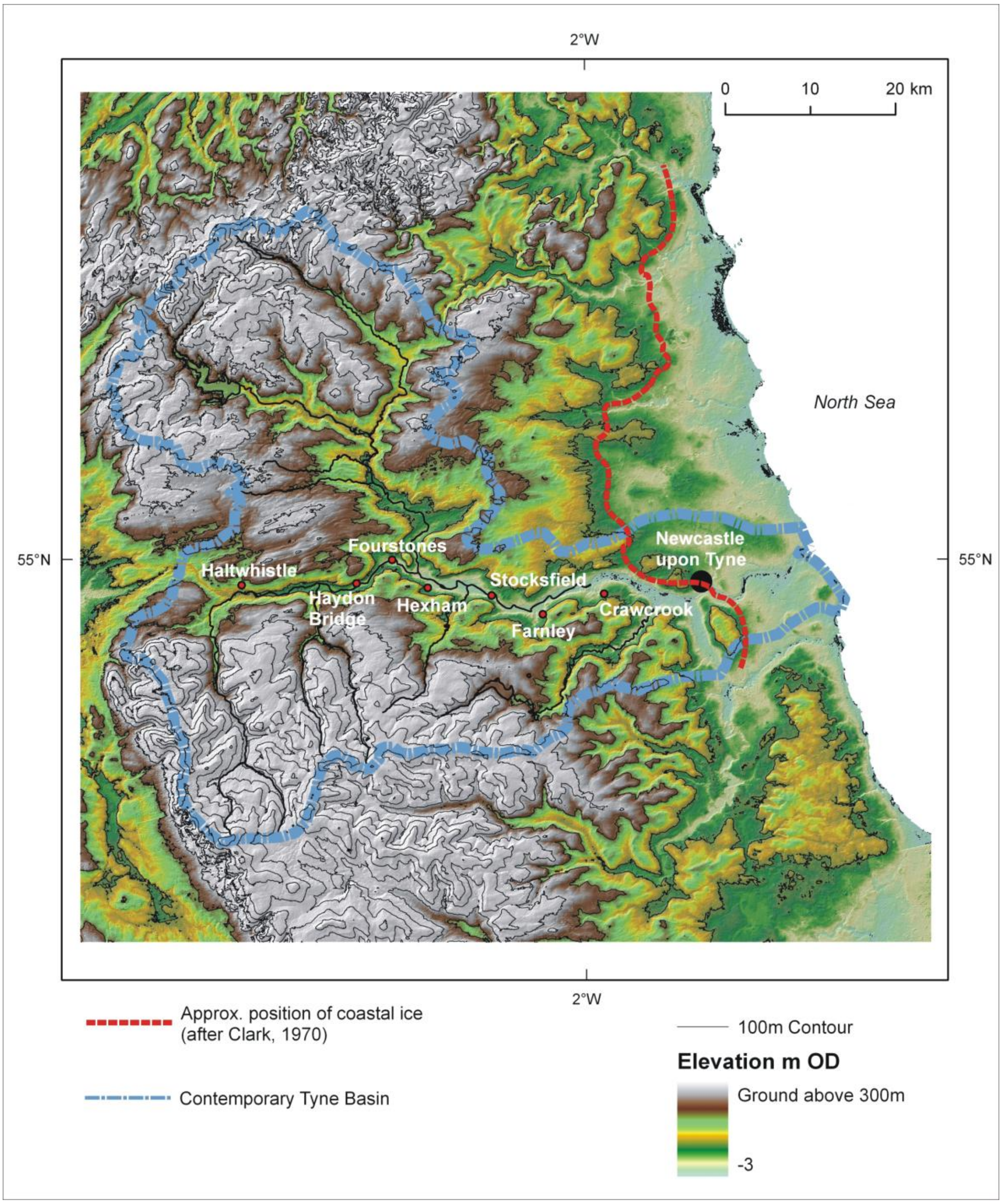


FIGURE 1.3: Map of northeast England outlining the Tyne Basin and showing the location of Clark's (1970) proposed coastal ice-margin. (Base elevation map is 25m DEM downloaded from www.landmap.ac.uk).

BP) and between *c*.17.5ka cal. BP and 14.5ka ^{10}Be BP. Their models demonstrate regional synchrony in the timing of ice sheet fluctuations, which suggests a common response to regional climatic change during deglaciation. However, in northeast England a deglacial chronology does not exist. Clark (1970) suggested that following the Last Glacial Maximum (LGM; conventionally taken as 21ka cal. BP) deglaciation was simultaneous across the region, characterised by ice wastage *in-situ* and the different responses of the western ice, which withdrew, and the northern ice, which continued for a time to be nourished and retain its thick, dense mass. There are no recorded positions of terminal, push or ribbed moraine (see Bennett and Glasser, 1996 for definitions) to mark the position of ice retreat (i.e. major standstill) or fluctuation. This, of course, may be indicative in itself pointing to *in-situ* ice stagnation and downwasting or rapid retreat with sediments widely distributed (cf. Bennett and Glasser, 1996). Work carried out by McCabe (1996), McCabe and Clark (1998) and Knight (2003) in the Irish Sea Basin and northwest Ireland, has demonstrated the value in investigating glacigenic landform-sediment assemblages (e.g. drumlins, eskers, outwash terraces). They have used this data to reconstruct ice sheet extent, locate ice margins and understand its behaviour. Nevertheless, in northeastern England this type of approach has not been carried out and although a corpus of borehole logs exists for the lower Tyne Valley, which could be interpreted as indicating early deglaciation and subsequent ice-marginal readvance (cf. N.E. Inst. Min. Eng. 1878-1910; Armstrong and Kell 1951; Land *et al*. 1974), it is futile to postulate on such low resolution data. As Cumming (1977) suggests, the sequences most likely represent evidence of reworking and redeposition of outwash sediments from further up stream.

Kendall (1902) first recognised evidence for deglaciation in the landscape in the Cleveland Hills and proposed a conceptual model for deglaciation characterised by glacial lake development. This model then formed the basis for subsequent workers to interpret deglacial sequences. Therefore, early 20th century deglacial models for northeast England were based largely on Kendall's work (1902). Dwerryhouse (1902), Herdman (1909), Smythe, (1912) and Anderson (1940) all envisaged active ice sheet retreat during deglaciation and the development of lakes and overflow channels.

Smythe (1912) proposed that the lower reaches of the rivers draining the retreating ice front were impounded by ice flowing down the coast, resulting in a series of ice-dammed lakes receiving meltwaters and sediments, which were subsequently incised as the lakes drained. In the Tyne Valley, Dwerryhouse (1902) suggested ice retreated into the valley from the high ground, whereby the ice blocked meltwater drainage flowing along the tributary valleys (River East and West Allen, Devils Water, River Derwent, Figure 1.4), resulting in the formation of lakes. They identified successive lake levels, which were related to a falling or withdrawing ice margin, and overflow channels that providing connections between adjacent lakes. Landforms were identified as deltas due to their coarse gravel sedimentology and lacustrine sequences appeared to provide conclusive evidence for the interpretation. Anderson (1940) suggested that as the ice-margin retreated north from the Durham area, drainage was impeded to the south, resulting in the formation of ice-dammed lakes.

As Mills and Holliday (1998) observe, the main difficulty with the glaciolacustrine theory is the assumption that large volumes of meltwater were impounded by impermeable ice dams with progressively falling lake levels cutting overflow channels. Peel (1951, 1956) initially drew attention to the flaws in the theory but it was Sissons (1958) who demonstrated that the features identified as overflow channels were in fact formed subglacially. Sissons (1958) concluded there was a lack of lake marginal

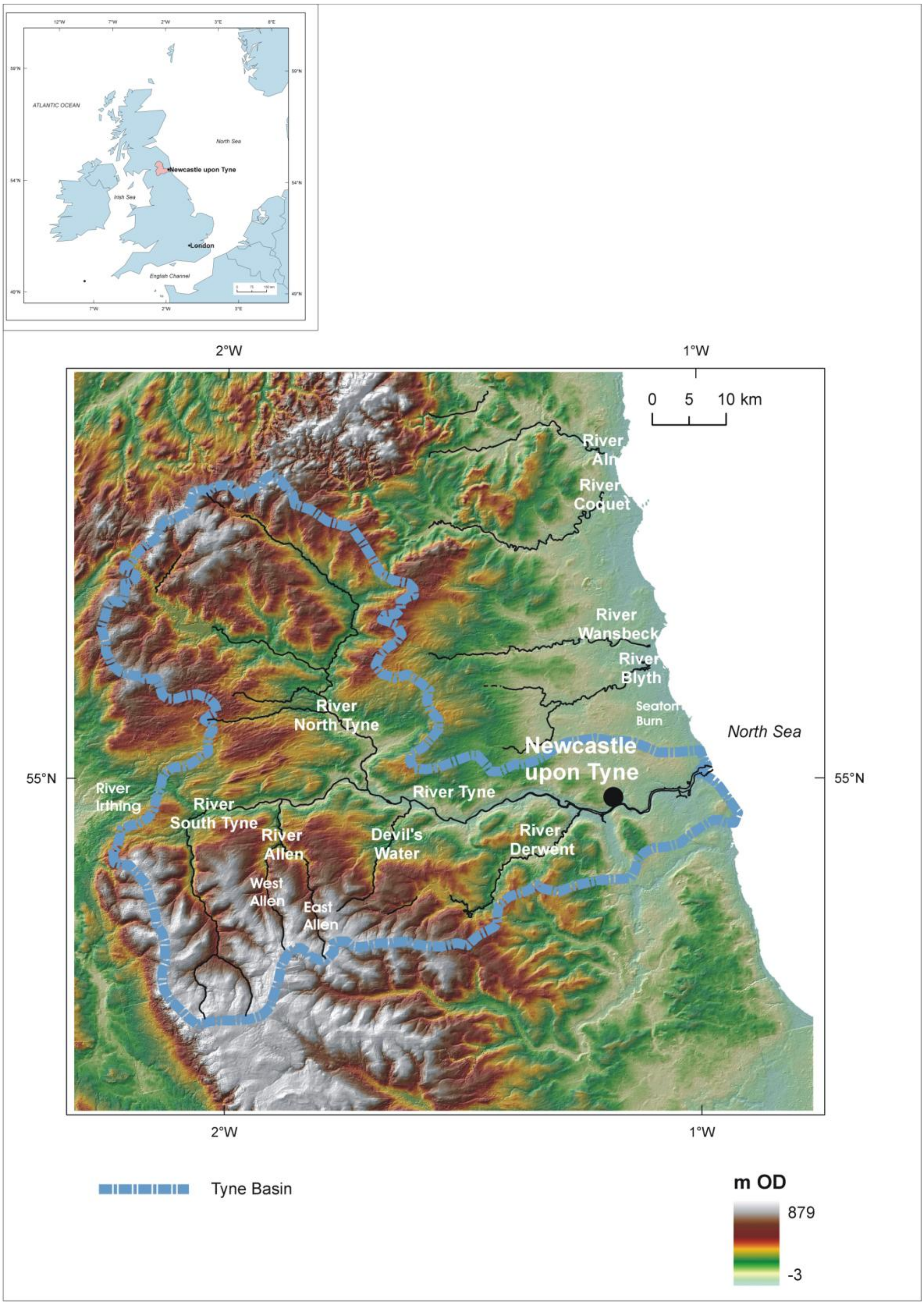


FIGURE 1.4: Map of northeast England outlining the Tyne Basin and showing the location of tributary streams, and other rivers mentioned in the text. (Base elevation map is 25m DEM downloaded from www.landmap.ac.uk).

features or clear evidence for deltas, proving Kendall's (1902) glacial lake theory was not required to explain channel formation and suggesting glacial lakes were not the main component of regional deglaciation. Subglacial meltwater drainage was initially controlled by the hydrostatic head and pressure of the ice but subsequently as the ice degraded, under more local, topographic control (Clark 1970; Mills and Holliday 1998). Through the work of Clapperton (1967, 1970, 1971) in north Northumberland, who interpreted the complex of sand and gravel landforms (e.g. eskers, kames; cf. Common, 1954) in terms of ice wasting and stagnation *in-situ*, the glacial lake deglacial model was finally defunct.

Clark (1970) developed the idea of coastal ice penetrating into the lowlands, suggesting that during the early stages of deglaciation coastal ice thickened and advanced just as the western ice was stagnating and withdrawing. Lambeck's (1993) ice sheet model, which predicts a number of short-lived surges of the British North Sea ice lobe, provides support for deflection of the coastal ice inland. This resulted in the development of an ice-free lowland area, between Hexham(?) and the coastal ice in the lower Tyne Valley. This area received meltwater from the western ice front, which although reduced, continued receiving nourishment from the uplands. Despite coastal ice erratic limits being know it is still unclear where this ice margin lay or how far it penetrated inland (Teasdale and Hughes, 1999). However, although little detailed work has been carried out on the extensive coastal till exposures that crop out both north and south of the Tyne Estuary, Teasdale and Hughes (1999) have shown that the till is related to the coastal ice sheet.

Clark's (1970) model suggests that the ice-free area became a sediment trap, with restricted drainage in the lower valley and resulting in a suite of glacigenic depositional landforms underlain by glaciofluvial and glaciolacustrine sediments. Smith (1981; 1994) provided further support for this model with notion of the development of a major glacial lake towards the end of deglaciation as a result of the inhibition of meltwater movement near to the present coastline, known as Glacial Lake Wear. The lake is thought to have extended up to 50m above OD, and evidence of lake bottom beds have been recorded in the lower Tyne east of Newcastle (cf. Land, 1974). Hughes and Teasdale (1999) record significant thicknesses of glaciolacustrine sediments in the Durham lowlands that provide further evidence for a glacial lake, with a high stand at 43m above OD.

Mills and Holliday (1998) more generally advocate that in association with a diminishing and decaying ice sheet, smaller proglacial water bodies would also have developed as temporary obstructions impounded escaping meltwaters. Regional deglaciation is envisaged as widespread down-wasting of stagnant and dead ice, and associated with the development of proglacial lakes. Landform-sediment assemblages associated with deglaciation are suggested to be widespread in all river valleys of northeast England (Lunn 1980; Allen and Rose 1986; Douglas 1991). These significant sediment bodies have been variously interpreted as proglacial outwash and ice contact features (see section 1.2.3 for discussion), suggesting high concentrations of debris was carried in the basal ice (cf. Bennett and Glasser, 1996).

1.2.3 Landform-sediment assemblages

Research in the Tyne Valley dates back to the late 19th century, and focuses on the features that crop out along the valley and sediments that underlie them. This earliest work was restricted to a few publications by G. A. Lebour, a Professor of Geology at Durham College of Science, Newcastle upon Tyne (Lebour, 1878, 1884, 1889, 1893), who reported his observations and findings. Lebour (*op cit*.) recognised a series of deposits that he subdivided into glacial (upper drift sands and gravels) and fluvial (old river gravels and sands) based upon their sedimentology and morphology. The upper drift sands and gravels formed conspicuous mounds on the ground between 90-150m OD, and the old river gravels and sands were derived from the pebbly drift that infilled the valleys up to 100m OD (the terraces become less distinct above this level); where the deposits were clearly terraced, a fluviatile origin was postulated. Lebour's research was centred in the South Tyne and mid Tyne Valley, and here he distinguished a series of stepped terrace sequences that are well developed on both flanks of the South Tyne Valley between Haltwhistle and Warden Hill, along the Tyne Valley between Hexham and Corbridge, and crop out through the Tyne Gap (Figure 1.5). Their composition varies along the length of the Tyne; in the mid to lower Tyne (between Corbridge and Wylam) the terraces are predominantly composed of sand; however, in the lower South Tyne Valley (above Warden Hill) coarse gravel predominates, and the gravels are less well-sorted above 100m OD. In the South Tyne particularly, Lebour (1878, 1889) observed that the terraces did not always keep attitudinally separate courses, and that they could be seen to merge with one another. Earlier work by Burns (1875) noted a suite of gravel terraces that lay along the Tyne Gap Valley (west of Haltwhistle) to the River Irthing Valley (Figure 1.3), though the number of terrace units or the altitude of the surfaces was not given. He recognised that the River Irthing (which now drains into the Eden system) previously formed part of the river Tyne system during deglaciation, draining the ice margin, and flowing across the present day watershed to enter the South Tyne south of Haltwhistle. He suggested the terraces were cut into glaciofluvial deposits as the meltwaters flowed eastwards from the ice front, through the Tyne Gap and into the Tyne Valley. Raistrick (1931) also suggested that during the later stages of deglaciation spreads of sand and gravel were deposited in the middle

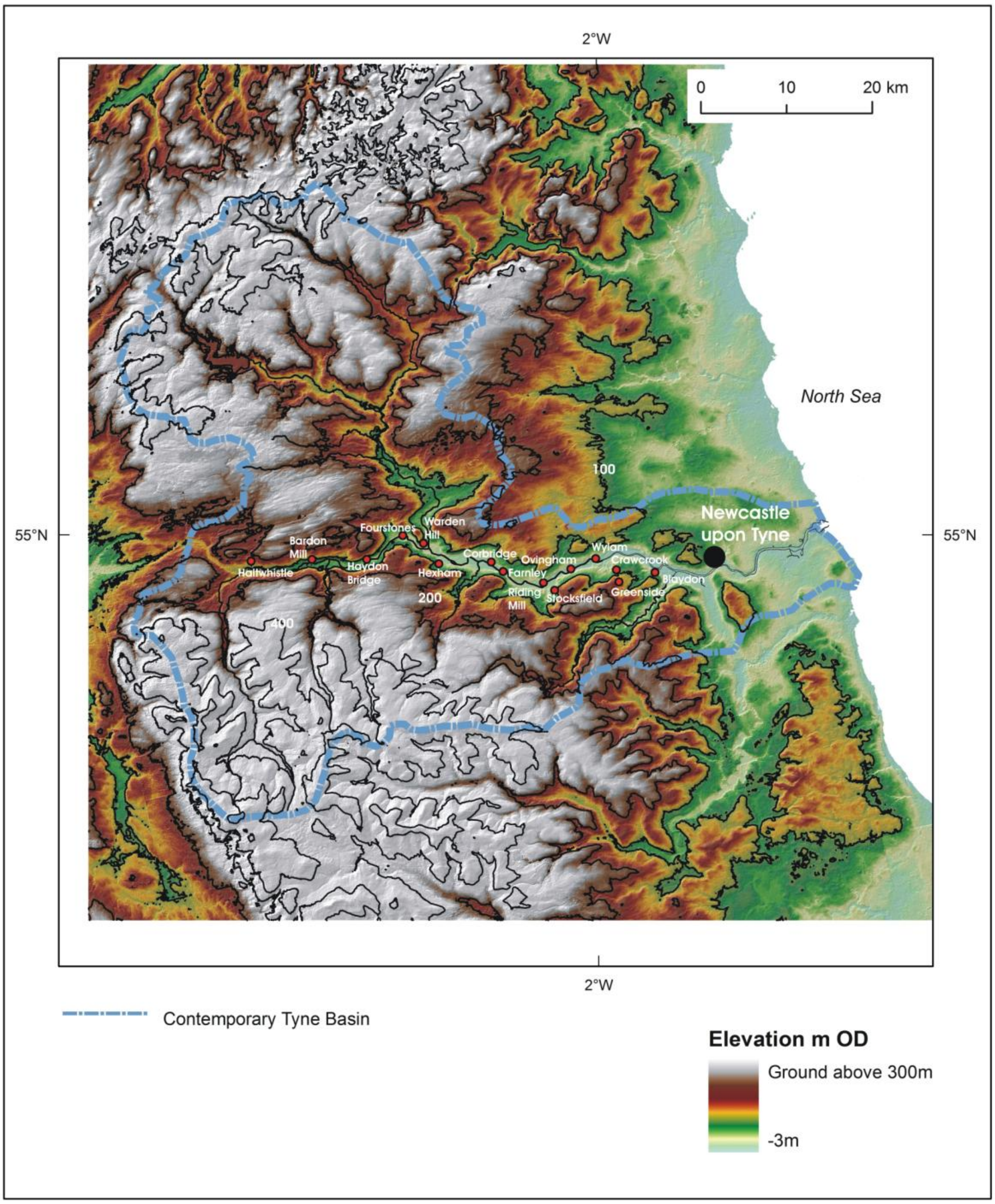


Figure 1.5: Map of northeast England outlining the Tyne Basin and showing the location of towns mentioned in the text. (Base elevation map is 25m DEM downloaded from www.landmap.ac.uk). Contours are given in m OD.

Tyne Valley by meltwaters reworking diamicton and morainic deposits from the western ice-margin draining through the Tyne Gap.

The terraces along the mid Tyne Valley are much lower and fewer in number than those which occur in the South Tyne Valley (Lebour 1884). The recent (Holocene) alluvial terraces lie up to 2m above present river level; the higher, older terraces are more denuded and less distinguishable. Lebour (1893) identified a sequence of terraces between Corbridge and Riding Mill (Figure 1.4). Cut-bank sections revealed the basal sequence comprised boulder clay (diamicton) overlain by sand and gravel with interbedded with laminated fine silty clay. Lebour (*op cit.*) suggested that the formation of the higher terraces represented incision into the valley infill. Woolacott (1905) also suggested that the terraces in the South Tyne Valley were produced by incision through the drift (glacigenic) infill.

In terms of the landform-sediments assemblages in the lower Tyne Valley and its tributaries, previous investigations focused on the identification of discrete assemblages, accompanied by only limited sedimentological investigations. Francis (1970) recorded exposures near Greenside (Figure 1.4), describing ripple cross-laminated sands with a west-east palaeocurrent direction and gravels towards the top of the sequence, which he interpreted to be deposited by a braided stream that was in part in contact with a stagnant ice margin. Further exposures in such deposits recorded by Francis (1975), in a ridge south of the current quarry at Crawcrook, described horizontally bedded sands overlain by irregular beds of sand with lenses of gravel, which in places were sharply tilted to almost vertical. Francis (1975) interpreted this sequence as part of a proglacial glaciofluvial system depositing material on the surface of a wasting ice mass and attributed the undulating topography to collapse associated with the melting of buried ice. Across the valley from Crawcrook, Allen and Rose (1986) investigated undulating mounds in the Derwent Valley (Figure 1.3;1.4). They interpreted these sediments as sub-glacial, deltaic sands and gravels deposited at the margins of a major proglacial lake, of unknown date and extent. They argued that the deltaic drainage system continued to operate after the lake had drained, resulting in the aggradation of glaciofluvial landforms, while stagnant ice remained in the valley floor. The most recent review of the landform-sediment assemblages in the lower Tyne Valley was undertaken by the British Geological Survey (BGS; Mills and Holliday, 1998), and based on a combination of new field survey augmented by synthesis and review of the subsurface investigations undertaken by the Industrial Mineral Assessment Unit (IMAU; Giles 1981; Lovell 1981). This led Mills and Holliday (1998) to conclude that the sand and gravel mounds were aggraded by seasonal streams issuing from stagnant ice within a subglacial environment.

Clark (1970) and Lunn (1980), in more widespread reviews of the Quaternary of northeast England, envisaged widespread down-wasting of stagnant ice in the Tyne Valley, with meltwater issuing from the margins depositing extensive spreads of sand and gravel. They interpreted the extensive undulating landform-sediments assemblages as kames, deposited in an ice-contact environment. From Hexham, along the South Tyne Valley to Haydon Bridge (Figure 1.4), a complex of mounded landforms crop out above the valley floor terraces (Lunn, 1995). Lunn (*op cit.*) suggests these are glacigenic, ice-contact landform-sediment assemblages aggraded in supra-, en- and sub-glacial settings, subsequently incised by meltwaters flowing into the Tyne Valley *via* the Tyne Gap (Figure 1.1).

Although pre-Holocene river terraces have been recorded along the Tyne Valley, little is known about them in comparison to the work carried out on the Holocene cut/fill units. Indeed, recent discussion of the river terraces and their sediments in the Tyne Valley is restricted to a few paragraphs in the BGS memoir for the country around Newcastle (Mills and Holliday, 1998), and to two volumes of the mineral assessment reports (IMAU) for the country around Blaydon and Hexham (Giles 1981; Lovell 1981) (Figure 1.4). Two river terraces (T2 and T1) were identified along both flanks of the mid Tyne and lower South Tyne Valley, between the Allen confluence, near Bardon Mill and Corbridge (Figure 1.4). Their surfaces in the South Tyne and Tyne Valley lie at between 78 and 30m OD (T2) and 87 and 34m OD (T1) respectively. However, in the South Tyne T2 and T1 lie at 10-8m above present river level, and in the Tyne valley T2 and T1 lie at 20-0m above the present river level; sand and gravel that crop out above these surfaces were interpreted as glacigenic (Lovell, 1981). Clearly there is some ambiguity between the two reaches, and there has not been consistent classification criteria applied to identification of terraces. Sedimentologically, there is very little to distinguish the terraces into members or associations (cf. Bridgland, 1986). The IMAU borehole logs, taken through some of the terraces, provide some indication of the sedimentary sequence. Both T2 and T1 generally comprise two facies: sands, and clayey sands and gravels, in cyclic sequences up to 20m in thickness (although many borehole logs provided incomplete records), overlying a basal diamicton. Downstream of the broad alluvial basin that extends between Hexham and Corbridge river terraces are also present. The ground surface of these terraces lies between 27 and 18m OD (8-4m above present river level) but they have not been differentiated on IMAU (Lovell, 1981) maps as individual or separate terraces as their lateral extent was not determined. A small number of borehole logs were recorded for these undifferentiated terraces, revealing they comprise a single facies: well-sorted sand and coarse gravel, up to a maximum thickness of 7.3m below the soil (Giles, 1981). There are no further river terraces identified downstream of Ovingham (NZ 085 637), lower Tyne Valley on the IMAU maps.

Peel (1941) recognised major terraces in the North Tyne Valley, recording two laterally extensive terraces at approximately 10-15m above the present river level. The IMAU survey (cf. Lovell, 1981) does not differentiate separate terraces by height, rather they record that terraces lie between 67 and 50m OD. Peel (1941) also recognised the altitudinal complexity of the terraces that Lebour (1889) made reference to in the vicinity of the South Tyne/North Tyne confluence, stating "…terraces are multiple and difficult to group by height" (Peel, 1941, p. 17). However, Peel (1941) did recognise an upper terrace group in the confluence area at approximately 30m above the present river level. In terms of their sedimentology, Lovell (1981) commented that the sediments underlying the North Tyne terraces were much thinner and less gravelly than either T1 or T2 in the South Tyne and Tyne Valleys. Given that these terraces are higher than those identified and dated to the Holocene by Passmore and Macklin (1994; 1997; 2000), it could be assumed the terraces relate to the Lateglacial period; however, no dates or time periods have been assigned by the BGS or IMAU.

Thus far, there has been little work on the pre-Holocene river terraces in terms of trying to understand whether they relate to river development following deglaciation or if they are incised remnants of glacigenic sediments, and in particular the integration of these features within the glacial/deglacial models has not been attempted. It is well established that river terraces may be linked to changes in tectonics, climate, base-level (cf. Maddy 1997; Maddy and Bridgland 2000; Maddy *et al.* 2001) and catchment land-use (cf. Macklin and Lewin, 1997). Within the coastal zone of Northumberland, the story emerging from the denes (highly incised valleys characteristic of the northeast coast) is one of rapid uplift and incision during and immediately following deglaciation (Smith 1994; Evans 1999). It is evident in the alluvial fills (cf. Passmore and Macklin, 1994) that these processes continued to some degree into the Holocene, therefore, it could be expected that this has been transferred up valley and recorded in the terrace gradients. This would be a key indication that the changes in base level that took place during and following deglaciation, driven by climate and/or tectonics, were of sufficient duration to be recorded in the fluvial record. Peel (1941) published an unfinished report on the longitudinal profiles of the terraces in the North Tyne Valley, but also made comment on those in the Tyne Valley and its tributaries. Peel (1941) identified a major knick point at 106m OD in the current river profile of the North Tyne, also identifying the same knick point in the South Tyne and its tributaries (Allen, Derwent, Devils Water; Figure 1.3). Extending this profile down valley to the present day coast it indicates a base level at 45m OD. Such a high base level may be indicative of the combined effects of high post-glacial sea levels and isostatic depression due to the ice or possibly a consequence of glacial scouring and over-deepening of the channel resulting in incision by the river to achieve a new base profile. However, Peel also found the long profile of the Seaton Burn (one of the coastal denes; Figure 1.3) to have a knick point at the same base level, which suggests the knick point is related to a regional base level change (i.e. either eustatic, climatic or tectonic) rather than a local driver. The story of the upper terraces in the Tyne Valley and their relationship to base level change remains untold, with nothing more having been attempted since Peels' work in the late 1930s.

From this brief review of the available literature it is clear only a limited amount of work has been undertaken on the landform-sediment assemblages, and although interpretations have been proposed for sediment aggradation, no generally accepted landscape development model for deglaciation in the Tyne Valley exists. Furthermore, there is no clear model of post-deglacial development in the Tyne Valley, other than those that exist for Holocene fluvial development (Passmore and Macklin, 1994; 1997; 2000), where catchment land-use and short-term climate changes are the key drivers (Rumsby and Macklin, 1996).

1.2.4 Regional chronology

The chronology for the LGM of the last BIS was developed from a radiocarbon dated assay of moss fragments recovered from organic silts beneath Skipsea till at Dimlington, an exposure on the East Yorkshire coast designated the type site of the Dimlington Stadial (=LGM) by Rose (1985). Penny *et al.* (1969) reported radiocarbon dates from the silts of 21.7ka cal. BP (18,500 ∀ 499 ^{14}C yrs. BP) and 21.9ka cal. BP (18,240∀ 250 ^{14}C yrs. BP). Organic lake deposits at Kildale and Roos Bog overlying the glacial sediments at Dimlington, Holderness, returned ages of 16.7 and 13k ^{14}C yrs. BP (Beckett 1977; Jones 1977; Keen *et al.* 1984), bracketing the timings for the Dimlington Stadial and ice sheet fluctuations. Despite the uncertainties of earlier radiocarbon dates, these are comparable to ones published by McCabe *et al*. (2007) from eastern Scotland suggesting regional synchrony on the eastern coast of Britain during deglaciation.

Francis (1970) made inferences about the timing of deglaciation in northeast England based on the Dimlington date (cf. Penny *et al*., 1969). There are only two dates in Northumberland, and certainly one of those dates is very unreliable given the context. Whilst organic deposits have been recorded in a few kettlehole basins, which were investigated in the 1960/70s, no radiocarbon ages were derived from the organics contained therein. Bartley (1966) took a core from a lake deposit developed in a kettlehole basin associated with esker and kame topography on the north Northumberland coast (cf. Caruthers *et al.*, 1927), and Turner and Kershaw (1973) examined a kettlehole basin infill 10km south of Newcastle. Pollen zones have been attached to the sequences, which provide an indication of timings for Lateglacial and postglacial pollen development; however, it has been subsequently demonstrated (cf. Birks, 1989) that broad-scale pollen stratigraphic zones and their associated chronostratigraphic framework are no longer tenable. The arrival and progression of species is time-transgressive and therefore, it is not possible to generalise on the timings of pollen migration patterns. The two dates that exist for the timing of Lateglacial change in northeast England come from the Cheviots, north Northumberland. Radiocarbon dates of 13.5-13.2ka cal. BP (11.4 k∀100 ^{14}C yrs. BP) and 11.8-11.2ka cal. BP (10.0k∀45 ^{14}C yrs. BP) from a buried soil developed upon laminated silts and clays overlying till and from peat developed on glaciodeltaic sands and gravels in the Cheviots provides a maximum date for deglaciation in the region (Tipping, 1998). Harrison (2002) presented OSL dates from periglacial deposits developed in the Cheviots Hills. Three sequential dates were generated for a sequence that Harrison (1989) interpreted as solifluctate, deposited during deglaciation and periglacial mass-wasting following the LGM. The dates were 9.8k±1.64 ^{14}C years BP; 28.8k±10.22 ^{14}C yrs. BP; and 43k±12.21 ^{14}C yrs. BP respectively. Given the margin of error associated with the latter two dates (35% and 28%), these ages must be considered highly unreliable. The first date only provides some indication that, in the Cheviots, cold climate conditions prevailed during the Younger Dryas (Loch Lomond) Stadial.

1.2.5 Summary

Through previous work, the deposits in the Tyne Valley have been subdivided on the basis of morphology and altitude of the surfaces into relatively dated stages (Francis 1975; Giles 1981; Lebour 1878; Lovell 1981; Lunn 1980; Peel 1941). However, there is confusion about the genesis of the features that are present in the valley. It is unclear whether the terraces are kame terraces, river terraces or strath terraces related to incision of glacial deposits, secondly, whether the mound features are ice marginal deposits or glaciofluvial braided stream deposits and finally, there is no clear event stratigraphy.

There is a paucity of recent detailed sedimentological work on these glaciogenic and/or fluvial sequences within the Tyne. Combined with the poorly resolved chronology, this constitutes a significant gap in our knowledge of local deglacial and postglacial geomorphic and sedimentary response and their relationship to climatic change.

In addition, because no reliable dates exist, there is no model or chronology for deglaciation, and as such, comparison to other regional models from the Irish Sea Basin and adjacent areas and eastern Scotland or to the North Atlantic climate record (Bond *et al.* 1993; 1997; Bond and Lotti 1995) cannot be made. Therefore, it is unclear how this local sedimentological sequence fits within the current age model for deglaciation of the last BIS and whether response to millennial climate variations, sea-level change and isostatic uplift can be determined from the sequence.

1.3 Research Objectives

This research consists of a programme of field, laboratory and desk-based investigations of the landform-sediment assemblages in the River Tyne Valley between Haltwhistle and Crawcrook, a reach of approximately 40 kilometres.

The primary aims of this thesis are to (1) elucidate the geomorphological, stratigraphic and geographical context of the landform-sediment assemblages in the River Tyne Valley, (2) determine the terrestrial response during and immediately following deglaciation of the last BIS, (3) compare the sequence to the North Atlantic climatic record and (4) develop a model for deglaciation in northeastern England.

There are five objectives:

(i) To map the Late Quaternary geology and geomorphology of the landform-sediment assemblages and surrounding landscapes along the River Tyne Valley.

(ii) To clarify the genesis of the deposits, and deduce the depositional environments that produced the lithostratigraphic succession that has been the subject of much confusion.

(iii) To elucidate the deglacial chronology of the region.

(iv) To examine the mechanisms by which the terraces in the Tyne Valley formed; and with reference to the river terraces, to determine the relative importance of the factors that controlled development i.e. baselevel fluctuations, glacio-isostatic rebound, and climate change.

(v) To identify and define in more detail the sequence of events that resulted in the associated landform-sediment assemblages preserved in the Tyne Valley.

In more detail these aims can be expanded upon as follows:

(i) **Late Quaternary geomorphology of the River Tyne Valley**

Surficial geology and geomorphic units are determined by field mapping, DEM visualisation and air photo interpretation. Details provided by surficial mapping include relative succession of the major sediment units, and the location of geomorphic features such as outwash terraces, glaciofluvial deltas, outwash fans, river terraces and Holocene alluvial fills.

(ii) **Landform genesis**

Through the examination of the sedimentological and architectural characteristics of the deposits by undertaking an exposure-recording programme the origin of the landforms can be determined to elucidate the Quaternary geomorphological context.

(iii) **Deglacial history of the river valley**

To understand fully deglacial histories, evidence of deglacial landforms, depositional processes and the style of deglaciation are uncovered. The ways in which the deposits formed, the local sequence of events and the thickness of the deposits are investigated. The timing of deglacial events are determined by optical dating. Finally, correlations are made with deglacial models in adjacent areas of northern England and eastern Scotland.

(iv) **Postglacial river valley development**

The spatial extent and number of river terraces is determined by detailed field mapping and DEM visualisation, and the chronology of terrace development is determined by optical dating. The thickness of the deposits are investigated through spatial modelling, and sediment transfer rates assessed by sediment budgeting. Correlations are made with independently existing chronologies for fluvial development during the Late Quaternary in Britain, and with high-resolution marine $\delta^{18}O$ climatic records. The relative impact, if any, of climate variation, base level change and glacio-isostatic uplift during and following deglaciation on the river terrace sequence, and the spatial extent of base-level and isostatic influence will be determined by examination of terrace longitudinal profiles.

(v) **The event stratigraphy for the Tyne Valley**

The changing depositional environments that existed throughout and immediately following deglaciation of the last BIS are outlined by combination of the stratigraphic information, and deglacial and post-glacial histories. By assembling a corpus of information that, when placed within a geochronological framework, allows the comparison of the event stratigraphy for the Tyne Valley with models for the last deglaciation derived from the North Atlantic Seaboard, regional synchrony can be examined. The chronology of deglacial events and the role of paraglacial and fluvial processes on the recently deglacierised River Tyne in the postglacial history of the area are more difficult to define because they are dependent upon the success of optical luminescence dating.

1.4 Hypotheses

The hypotheses are concerned with the development of the landform-sediment assemblages. A crucial observation by both Lebour (1889) and Peel (1941) was the altitudinal discontinuities observed in the terraces in the lower South and North Tyne Valleys; this may be interpreted as development of local kame terraces which may have a more chaotic distribution pattern than river terraces. Where kame terraces are formed by meltwater streams flowing against steep valley sides they tend to form small, irregular fragments (Bennett and Glasser, 1996). Along with more recent reviews (cf. Lunn, 1980; 1995), it would appear that landform-sediment assemblages in the mid Tyne/lower South Tyne Valley could be interpreted as kame terraces formed by retreat of the western ice. However, in the lower Tyne Valley, the sequence appears more complex with kames, outwash plains, eskers, deltas and lake deposits (cf. Francis 1970; 1975; Lunn 1980; Smith 1981) suggestive of *in-situ* ice, ice proximal deposition and meltwater impoundment. Therefore, it could be concluded that locally there is discontinuity along the Tyne Valley, which may be related to the idea of ice-sheet separation at the start of deglaciation (cf. Clark 1970; Teasdale and Hughes 1999) and the impoundment of sediment charged meltwaters issuing from the western ice-front in the lower valley by coastal ice resulting in a major glacial lake and deposition. Finally, the story is complicated by the terrace sequence, which may represent post-glacial alluvial deposition or may simply be related to down-cutting (strath terraces) following deglaciation. Therefore, there are a number of key questions to be explored. The thesis will test the following hypotheses:

(i) The sediment-landform assemblages in the lower valley are underlain by proglacial sediments, developed in association with complex decay of the ice lobes; they are depositional rather than erosional features;

(ii) The highest terraces in the mid valley developed as proglacial outwash (sandur) plains, their long profiles graded to glacial Lake Wear when the lower valley was blocked by coastal ice. The terraces formed through incision following the disappearance of the coastal ice and Lake Wear; and

(iii) River terraces developed following deglaciation, their long profiles are graded to base-levels below OD, sediments aggraded on the valley floor in response to paraglacial reworking and were incised in combination with glacio-isostatic rebound, declining sediment supply and hydrologic change during the Bølling-Allerød Interstadial, and Younger Dryas Stadial.

These three hypotheses have been represented as two conceptual models (Figure 1.4). The thesis investigates which, if any, of the models are applicable by applying detailed sedimentological and morphological analysis of the landform-sediment assemblages along with the development of a precise chronological framework, at a high-resolution temporal scale.

1.4.1 Conceptual models

These models outline different interpretations for the accumulation and development of the sequences, and relate to different processes operating during deglaciation. This clearly illustrates the complexity and the contradictions of the sequences along the Tyne, and raises uncertainty about whether a single model of development does exist. The models are generalised, based upon currently available literature; they can be refined and may be supplanted by a single more complex model once the sequence has been successfully deciphered. It cannot be assumed that the sequences along the Tyne Valley represent a continuum and, therefore, they should not be seen as spatially or temporally contemporaneous. The models do not attempt to present a land-sediment system model for development. The landform-sediment assemblages represent disparate episodes in the development of the Tyne Valley and are separated both spatially and temporally.

The first conceptual model (Figure 1.6) shows the evolution of the undulating topography in the lower Tyne Valley. There are two phases of development: ice compression/thrusting at the boundary between the Tyne and coastal ice sheets leads to the development of an ice-cored topography in the lower Tyne. Sediments are initially deposited in a supraglacial context within this ice-constrained topography as glaciofluvial sediments (cf. Bennett and Glasser, 1996). As deglaciation continues, the ice is degraded and sedimentation continues to be deposited as proglacial outwash burying the ice. Deposition ceases prior to melt-out of the buried ice, resulting in the

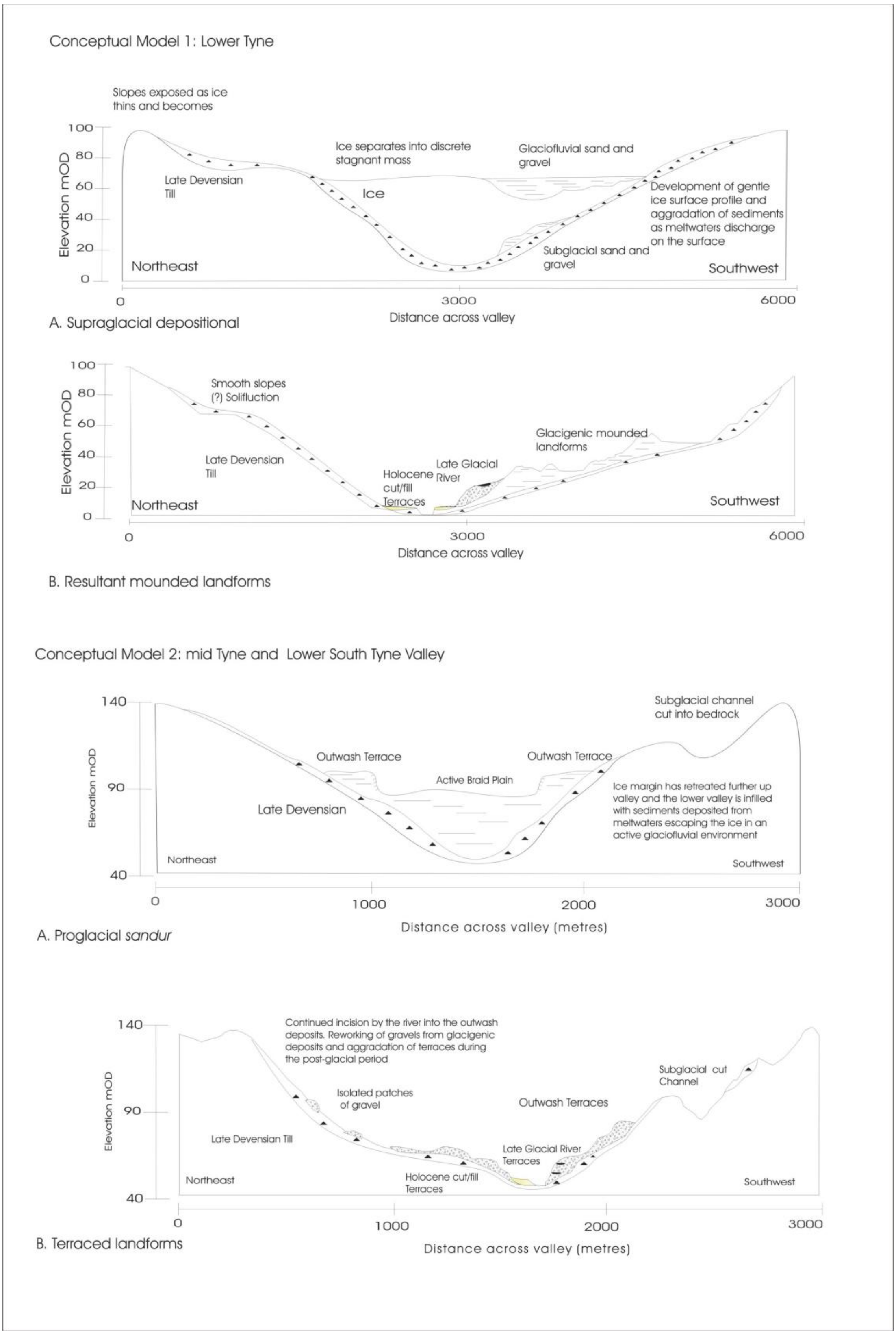


FIGURE 1.6: Conceptual models showing the development of the landform-sediment assemblages in the Tyne Valley. Model 1 is underlain by ice and develops in an ice-proximal environment. Model 2 is developed in a proglacial environment and develops as a consequence of postglacial incision.

uneven surface topography as the sediments slump and spread out as they are lowered down.

The second conceptual model (Figure 1.6) shows the evolution of the terraces in the mid Tyne Valley and lower South Tyne Valley. Initially deposition takes place along the ice-margin as the ice retreats westwards, forming kame terraces high on the valley side. Retreat continues and an ice-free area develops that is infilled by outwash as an extensive sandur develops, draining the ice margin which has retreated across the watershed into Cumbria and vast amounts of sediments and meltwater are delivered. Initial incision is contemporaneous, leading to the development of a series (or staircase) of outwash terraces. Following the complete disappearance of ice, the river re-established itself, and with continued high discharge, it reworked the readily available sediments, depositing and then incising the valley floor to form Lateglacial river terraces.

In order to test the models and determine whether the hypotheses are an accurate interpretation of the evolution of the landform-sediment assemblages, the sedimentary sequences underlying the landforms were investigated to establish the depositional environment, and to determine the genesis of features i.e. whether they were glacigenic or postglacial in origin. The mapping programme identified the spatial distribution of depositional (e.g. ice-contact) and erosional (e.g. fluvial) landforms. Finally, the development of a chronological framework helped to ascertain when the sediments were aggraded.

1.5 Thesis structure

The thesis comprises seven chapters. Chapter one provides a background to the research and is an introduction to the research problem, setting out the aims and objects of the project. Chapter two is a review and critical assessment of the existing literature related to each aspect of the hypotheses, including the geochronological methods used, and provides a context in which the research questions have been framed. Chapter three is an introduction to the study area, providing a general background to the Tyne basin and discussing how the sites were chosen for this study. Chapter four sets out the methods used to investigate and gather data on the landform-sediment assemblages so that the hypotheses and conceptual models can be tested. Chapter five is a systematic presentation of the results. Chapter six is a discussion of what the results mean and the possible interpretations. It draws together the various research threads into a coherent picture and links back to ideas and hypotheses that were nurtured in chapter one. Chapter seven presents the conclusions of the research, illustrates the contribution and significance of the findings and raises some ideas for future research.

Chapter Two

Literature Review

2.1 Introduction

In chapter 1, it has been demonstrated that confusion exists in terms of understanding the glacigenic landform-sediment assemblages in the research area, and associated with this, is how to begin to disentangle the deglacial history. Conceptual models have been proposed in chapter 1 to represent the evolutionary sequence of development. However, the models cannot be tested until the genesis of the landform-sediment assemblages, previously interpreted as both ice-proximal and proglacial, have been clarified. Therefore, the first section of this chapter concentrates on establishing the morphological and sedimentological characteristics of the principal glacigenic landform-sediment assemblages from key studies on Pleistocene and contemporary depositional sequences. The individual landform types are described, followed by a detailed description of the sedimentary facies associated with each type. By determining the probable depositional environments and interpreting the landforms present, it is possible to build up a pattern (or model) of depositional response and landscape development to ice retreat or stagnation, against which the conceptual models can be judged.

The second section of the chapter is concerned with postglacial terrace development. It has been acknowledged that following deglaciation, postglacial valley alluvial deposits can be related to sedimentation induced by the adjustment to non-glacial conditions but conditioned by those previous surroundings (cf. Church and Ryder's, 1972, paraglacial model). A brief synopsis on the mechanisms of formation is followed by a review of

Lateglacial terrace development in upland (glaciated) river valleys in Britain. This will establish whether development is region wide, as has been shown in the Holocene fluvial sequences (cf. Macklin and Lewin, 1993) possibly relating to a climatic driver, or are limited in their proliferation and more likely related to localised response following deglaciation.

The chapter is concluded by an overview of the dating methods used to develop chronological frameworks for long-term sequences. A more substantial review of the dating techniques applied in this study is presented in chapter 4.

2.2 Glacigenic Landform-Sediment Assemblages

In order to interpret the landform-sediment assemblages that characterise the study area, to understand how they relate to deglaciation and, therefore, to determine which of the conceptual models proposed best represents the evidence, it is necessary to understand fully the depositional processes and sedimentary facies and how they change spatially and temporally, and to recognise the resultant landforms that develop. Landform-sediment assemblages associated with deglacial environments record the nature of glacial retreat/response during deglaciation, and occur in two distinct environments: (1) ice-contact or ice marginal, where sediments are deposited sub-, en- and supraglacially or at the ice margin, and (2) proglacial, where sediments are deposited at and beyond the ice margin though they are controlled by meltwater and sediments delivered from the ice. There is some overlap between the two environments at the ice margin, and it is often difficult to distinguish the sedimentary environment in this transitional zone (Johnson and Menzies, 2002). Both fluvial and lacustrine processes are operating at and beyond the ice margin, and the landform-sediment assemblages reflect the different processes active in each environment, but they are also conditioned by the response to ice dynamics, local topography, regional and local climate, sediment availability and volume, and the hydrological regime (Menzies, 2002).

Close to the ice front, sequences are often chaotic, comprising coarse, poorly sorted sediments but with increasing distance from the ice front, as glaciofluvial processes dominate, sediments become rapidly finer and better sorted. In general, glaciofluvial sediments comprise moderately to well sorted sands and gravels. Sequences reflect lateral and vertical changes in grain size both spatially and temporally with an overall downstream fining due to selection and abrasion. Glacial lakes are a sediment store and where glaciolacustrine sequences occur, they generally comprise fine grained sands, silts and clays though they may contain coarse gravel and ice rafted debris dependant on proximity to the ice. The facies reflect the sedimentary processes operating in the lake such as mass movement, current or suspension deposition and ice rafting and, therefore, will also vary spatially and temporally.

2.2.1 Development and Morphology

This review briefly describes the development of ice contact, ice marginal and proglacial landforms and attempts to provide a general description of their morphology by re-examining some of the key sedimentological studies papers on glaciofluvial and glaciolacustrine systems from Pleistocene and contemporary environments.

2.2.1.1 Ice contact landforms

Deposition takes place within, alongside, beyond and above the ice as a result of the direct action of the ice (e.g. moraines) and from glacial meltwaters (e.g. eskers). Ice contact features are indicative of the position of the ice margin or discharge routes within the ice itself, whereas ice marginal features are indicative of ice sheet response during retreat and they reflect marginal fluctuations or downwasting.

Eskers:

These are ice-contact fluvial deposits, which develop in supraglacial channels and in englacial and subsequently lowered, or in subglacial, tunnels. The identification of esker types will help clarify the mode and pattern of deglaciation in the Tyne Valley. In general, they are characterised by sinuous ridges with undulating sharp or flat-topped crests, and the sequences are often characterised by post-depositional collapse structures (e.g. high-angle reverse faults) due to removal of supporting ice or melt-out of buried ice (Banerjee and McDonald 1975; McDonald and Shilts 1975; Delaney 2002). Eskers are time sequential, becoming younger towards their source as the ice recedes, and only become visible once the ice has disintegrated. General models of esker development have been proposed (Figure 2.1), and a number of esker landforms can be recognised: (1) single; (2) braided [anastamosing]; and (3) beaded [long or short] (Banerjee and McDonald 1975; Brennand 1994; Warren and Ashley 1994; Bennett and Glasser 1996). However, general models and simple descriptions do not always capture the complexity and diversity of these landforms.

Single esker ridges (Figure 2.1A) form when the channel or tunnel suddenly becomes blocked, e.g. by an ice fall, and sediments are deposited. The mechanisms by which sediments are aggraded in englacial or subglacial tunnels are poorly understood. Sedimentation (see section 2.2.2) within tunnels is modelled by applying a theory developed by deposition of solids within pipes, where little or no transport takes place at low velocities and all particles are moving in suspension (no sorting) at very high velocities (cf. Bennett and Glasser, 1996). Sudden blockages cause deposition, whereas a constriction will only increase velocity and entrain the sediment. Subsequent lowering or retreat of the ice reveals the esker ridge, which indicates meltwater routes within, above and below active ice but is not indicative of ice marginal positions *per se*.

Where a tunnel exits into a lacustrine basin, beaded eskers (segmented ridges or subaqueous fans; Figure 2.1 C, D) develop (Rust 1977; Gustavson and Boothroyd 1987; Donnelly and Harris 1989). Morphologically, Benn (1996) described a subaqueous fan with an asymmetric cross-profile with a gentle ice-proximal slope and a steep ice distal slope. Fans often form as a series of elongate ridges (long beads) or ridge and swales (short beads), reflecting their time-transgressive nature (Warren and Ashley, 1994). However, fans are often modified due to subsequent deposition or post-deposition erosion (cf. Paterson and Cheel, 1997) and may prove difficult to identify in the field. Subaqueous fans develop either: (1) as flow enters the lake at an intermediate depth as inflow (i.e. similar density to surface water but less dense than basal water) and deposition takes place (Smith and Ashley 1985; Johnson and Maizels 2002); or (2) by a combination of underflow (i.e. density is greater than the lake water) and suspension settling (Rust 1977; Sharpe 1988). Subaqueous fans may develop into complexes where a series of conduits form along the ice margin and the fans coalesce to form a moraine (Figure 2.1E), such as those recorded by Lunkka and Gibbard (1996) at an ice lobe margin in Finland. The presence of subaqueous fans represent former ice margin positions; they are indicative of deposition within standing water and are crucial in interpreting the mode of deglaciation.

Brennand (1994) describes braided eskers in Ontario, Canada, where formation is more complex and there are a number of possible explanations. Braided channels may result from: (1) catastrophic subglacial floods (cf. Shaw *et al.*, 1989); (2) through a cross-cutting pattern of time-successive channels; or (3) lowering of a supraglacial outwash fan that has deeply incised sediment-infilled channels which are inverted during melt-out to form a braided esker (Bennett and Glasser, 1996).

Supraglacial eskers form from subaerial sedimentation within ice-walled channels (Figure 2.1B) that are subsequently lowered when the ice ablates. Where ice-wall channels are formed in an interlobate area, ice was probably stagnating (Warren and Ashley, 1994), although Russell *et al.* (2001) have suggested that stagnating ice is not a prerequisite for ice-wall channel development. They demonstrated that contemporary ice-walled deposits in Iceland developed as a result of tunnel collapse and ice-margin break up due to high-magnitude *jökulhlaups* [glacial lake out-bursts], thus, implying that the ice was active rather than stagnating; this has implications for modes of deglaciation. Both braided and supraglacial esker ridges indicate the complexity that may arise within and above the ice, possibly indicating high-magnitude events but their presence in the landscape is more clearly related to processes operating within the ice, suggesting that during deglaciation the ice was active.

Previous workers (see chapter 1) identified possible eskers in the Tyne and Derwent Valleys that they interpret to represent active retreat. However, eskers should be viewed as part of the larger glacial landsystem (cf. Benn and Evans, 1998), and their association with surrounding sediment complexes examined in order to understand fully their origins and determine the mode of deglaciation. Eskers found in association with ice marginal topography in the Dee Valley (cf. Brown, 1993) indicate a complex pattern of ice retreat associated with both active and dead-ice.

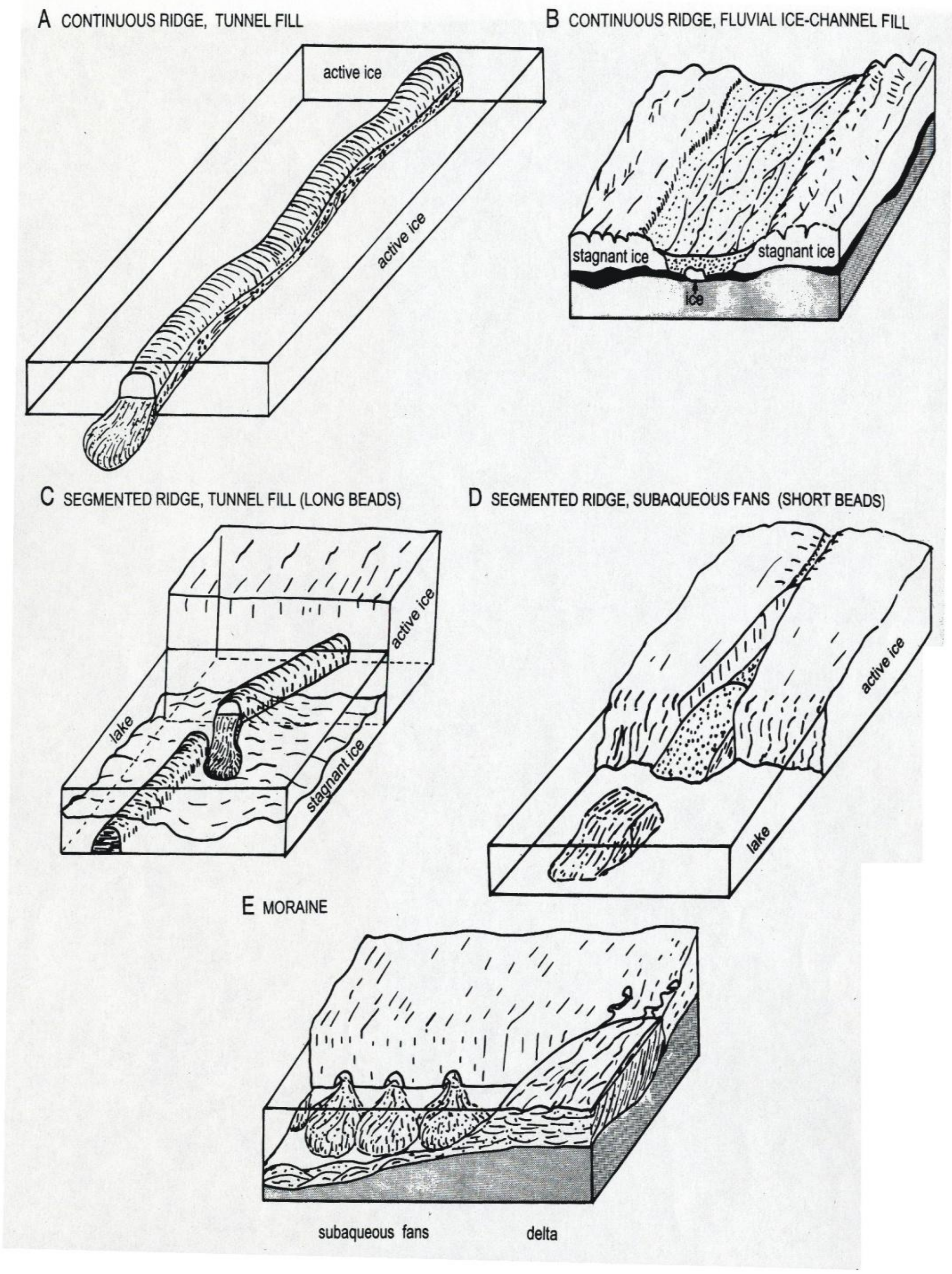


FIGURE 2.1: Sketch illustrating the depositional environments of five-types of ice-contact ridge landforms (taken from Warren and Ashley, 1994).

Kames:

Kames are defined as landforms comprising glaciofluvial sediments deposited on, within or under ice, which are a collection of irregular mounds and ridges, often developed in association with kettleholes (Bennett and Glasser 1996; Johnson and Maizels 2002). In the older literature, the term 'kame' has been applied to a complex of different glacigenic landforms (e.g. eskers, outwash plains and fans) with obvious confusion ensuing (cf. Francis, 1975 for discussion); however, Huddart (1999) has suggested the term kame has no genetic significance. The current usage of the term refers to both ice contact and ice marginal depositional and erosional landforms that comprise both glaciofluvial and glaciolacustrine sequences (cf. Bennett and Glasser 1996; Huddart and Bennett 1997; Delaney 2002; Knight 2006). Therefore, the term 'kame' does not represent a type of landform-sediment assemblage that has a distinctive morphology or sedimentary sequence, but is rather a generic descriptive term for landform-sediment assemblages in a deglacial environment rather than a genetic one. Thus, where 'kamiform' landscapes have been described in the Tyne Valley (cf. Lunn, 1995) they do not indicate a mode of deglaciation but refer to a collection of glacigenic landforms that remain to be fully investigated.

Kame terraces:

In contrast to kames, kame terraces are more precisely defined in morphological terms. They refer to terraces that develop between the lateral margin of the ice and the valley side but also form at the glacier snout where it rests against a reverse slope (Figure 2.2C, D), although they are not associated with a specific sediment assemblage because they may comprise both glaciofluvial and glaciolacustrine sediments (Bennett and Glasser 1996; Britinas *et al*. 2004). Kame terrace morphology is dependent upon: (1)

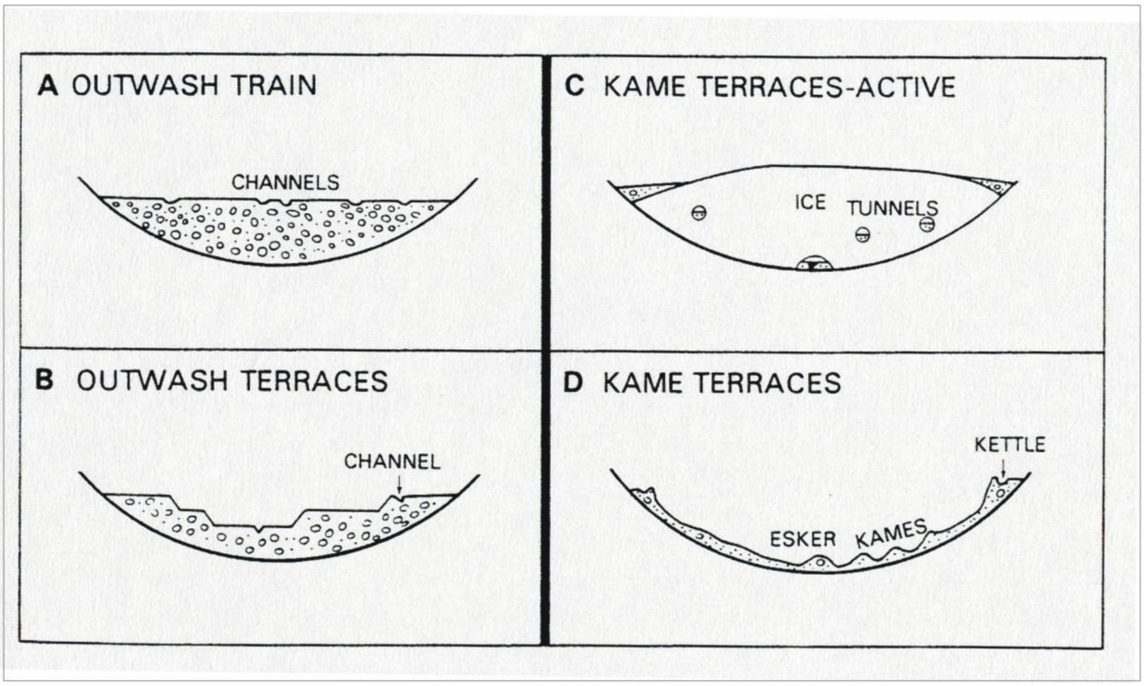


Figure 2.2: Sketch illustrating differences between kame and outwash/river terraces (taken from Gray, 1991).

size of meltwater stream; (2) steepness of ice margin, their surface gradient usually reflecting the ice-margin slope in the down-ice direction; and (3) angle of slope or hillside against which ice rests (Gray 1991; Bennett and Glasser 1996). Kame terraces are characterised by five key features: (1) they are depositional not erosional landforms; (2) they are higher at the valley centre than at the valley side; (3) they are not altitudinally matched since each valley side operates as a largely independent hydrological system as well as because of variations in microclimate related to valley aspect; (4) they contain subsidence structures at their margins when viewed in section (e.g. extensional faults and folds) due to the removal of ice support or melt-out of buried ice; and (5) they have an irregular upper surface and have ice disintegration features (i.e. kettleholes) along their front margins (Bennett and Glasser 1996; Johnson and Maizels 2002).

Ice disintegration landforms:

Kettleholes are ice disintegration features associated with glaciofluvial deposition at/near the ice-margins and in ice proximal locations due to buried ice; they commonly occur in association with kame terraces, outwash fans and in the proximal zone of sandur plains. They are ice-contact features that develop as hollows when an ice block has melted *in-situ* or from collapse of marginal ice masses buried beneath proximal proglacial deposits (Maizels, 2002). They are indicative of ice stagnation and wastage. Their size varies from a few metres to a kilometre or more in diameter and some are over 20m deep (Gray, 1991). Meltwater streams divert into the kettleholes, with deposition proceeding rapidly due to standing water within them and the associated velocity reduction resulting in small infilled basins comprising lacustrine sediments (Bennett and Glasser, 1996).

From this review, ice-contact and ice-marginal landform-sediment assemblages comprise both glaciofluvial and glaciolacustrine sediment units. Kame terraces and esker ridges are morphologically distinct, and through their ice-contact nature and association with kettleholes and proglacial outwash deposits have meant that they have often been collectively described by the generic landsystem term as kames. The identification of eskers and kame terraces in the field area is indicative of active ice retreat, the location of the ice-margin, the presence of standing bodies of water, and localised ice stagnation at the ice-margin.

2.2.1.2 Proglacial landforms

Beyond the ice-margin, the proglacial environment is dominated by meltwater and sediments issuing from the ice. Outwash deposition frequently starts on or builds back over the ice-margin, incorporating buried ice. The morphologies of ice proximal outwash landforms are, therefore, dependent upon rate at which buried ice melts relative to evolution of the drainage system. Post-depositional melt-out results in surficial deformation by subsidence structures i.e. kettle holes (cf. Price, 1973) or kame and kettle topography (cf. Stone, 1959). Whereas, with contemporaneous melt- out there is little surficial evidence of subsidence but significant faulting and tilting of beds is seen within the sedimentary record (Bennett and Glasser, 1996).

Initial deposition takes place as meltwater emerges from the ice margin, the pressure gradient which drives it drops and velocity falls, leading to deposition of entrained sediment. The proglacial fluvial regime is characterised by large-scale irregular and regular variability in runoff magnitude. Meltwater flow to the proglacial environment is characterised by fluctuating discharges and sediment loads, driven by diurnal and seasonal ablation cycles, related to long-term ice sheet behaviour as the ice ablates and the size of the ablation zone changes (Maizels, 2002). The combination of abundant sediment availability and a meltwater regime capable of transporting sediments beyond the ice margin results in the development of extensive sandur plains dominated by braided rivers (Smith, 1985). Deposition in the proglacial environment can be divided both spatially and temporally into proximal, medial and distal zones of outwash (Krigström, 1962). Morphologically, the development of outwash landforms is dependent upon: (1) location in respect to ice margin; (2) presence/absence of buried ice; (3) total volume of sediment input within meltwater; (4) the topography and accommodation space of the receiving area; and (5) the nature of the distribution processes (Bennett and Glasser 1996; Maizels 2002).

Outwash fans:

These are deposited at or immediately beyond the ice-margin, forming the interface between the ice-margin and proglacial environment; fans often merge into large sandur sequences. They are indicative of a relatively static ice-margin and reflect relatively rapid deposition in the proximal zone, fed by a single major glacial meltwater route to reach the ice-margin at a single point (Bennett and Glasser 1996; Maizels 2002). Their distal flank may be pitted by kettleholes; their surface contains a shallow relief formed by abandoned channels and bars (Bennett and Glasser, 1996). Fan complexes develop where a series of sub-parallel, subglacial drainage routes emerge along the ice margin and the deposits combine to form an 'apron' blanketing the margin (Bennett and Glasser, 1996). Morphologically, outwash fans have a steep ice-contact face, and grade away steeply in the down-ice direction (Bennett and Glasser, 1996). Subsequent dissection or burial of the outwash fan as the ice margin retreats and a new fan develops means surface morphology is often destroyed, thus reflecting spatial and temporal responses of the ice during deglaciation. However, by analogy to fans in contemporary proglacial environments they tend to possess a semiconical shape, similar to a gentle symmetrical cone (Goldthwait, 1989), have a restricted radial length, a planoconvex cross-profile and slope gradients between 1 and 5^0 (Kjær *et al*., 2004). Fan lengths vary from 3 to 30km (Boothroyd and Nummedal, 1978).

Sandurs:

Sandur systems are characterised by an extensive network of braided rivers (Figure 2.3A). Krigström (1962) differentiated two kinds of sandur: (1) valley sandurs; and (2) sandur plains. The former are usually associated with valley or cirque glaciers (Smith, 1985) and develop where meltwaters are confined between valley sides and sediments are concentrated on the valley sides. Morphologically they are just long narrow bodies of glaciofluvial sediment (Maizels, 2002), which exhibit a wide range of down-valley slopes and complex longitudinal profiles (cf. Church, 1972) indicative of the response

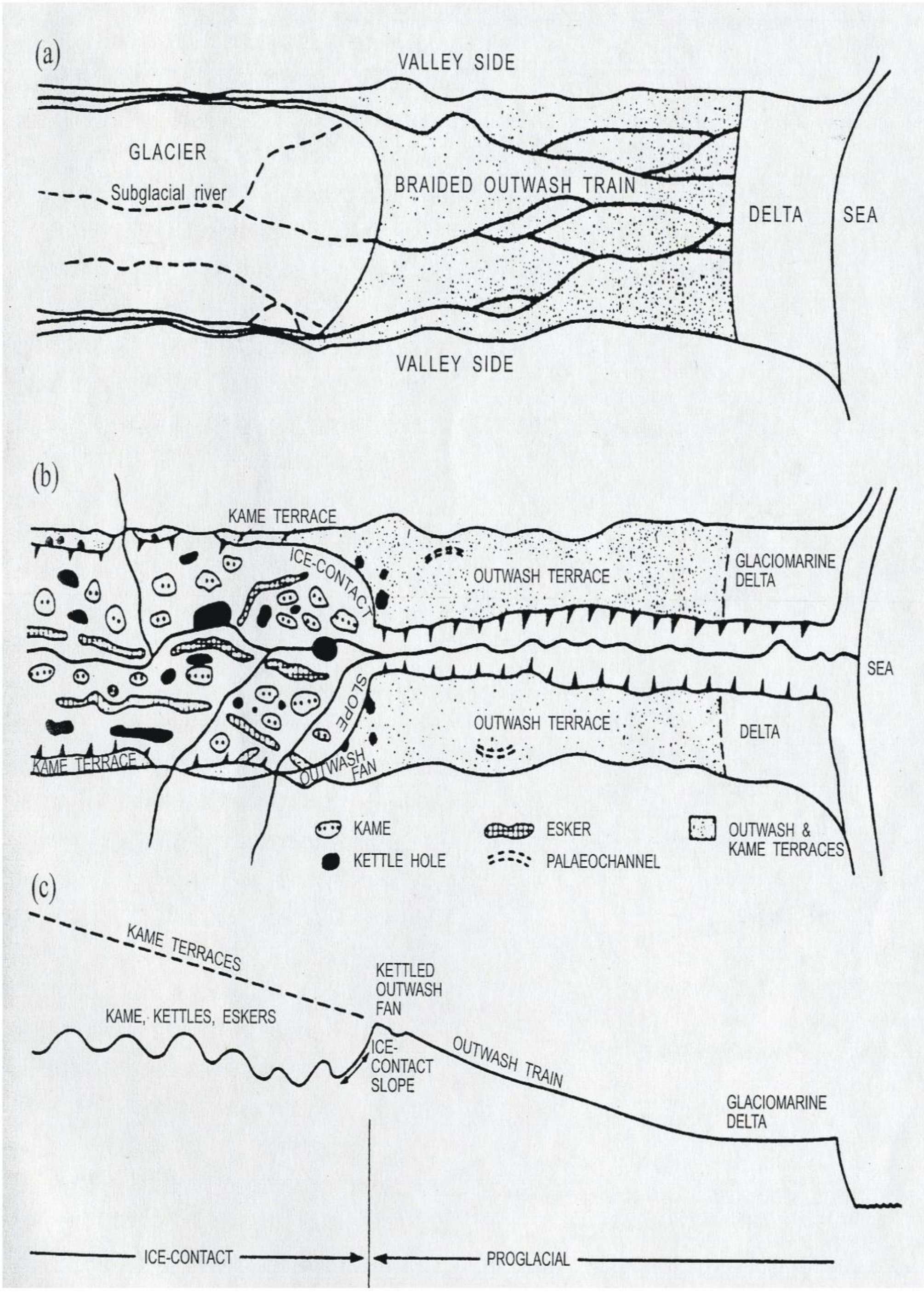


FIGURE 2.3: Sketch illustrating the depositional environments of ice-contact and proglacial landforms (a) during glaciation, (b) after glaciation, and (c) longitudinal profile of (b) (taken from Gray, 1991).

to changing conditions at the ice margin. Sandur plains form where accommodation space is not restricted by valley sides, thus the rivers are not constrained and sediment can extend laterally for considerable distances (Maizels, 2002). Sandurs have a flat surface morphology and, in general, are characterised by progressive downstream change in slope as the ice-margin retreats, giving rise to concave longitudinal profiles (cf. Church 1972; Smith 1985). The intermediate zone is characterised by frequently shifting braided channels, developed in response to changes in slope and particle size, which decrease rapidly with distance from the ice margin, and frequent fluctuations in discharge (Maizels, 2002). Significantly lower gradients occur in the distal zone compared with those of the proximal zone, demarcating a discontinuity between the two zones, possibly relating to change in sediment calibre (i.e. gravel to sands), additional inputs of sediments from the ice margin or slope adjustment to the stream power required for transport (Maizels, 2002). Along the length of the sandur, changing fluvial conditions reflect the transition from proximal to distal locations (Table 2.1).

TABLE 2.1: Summary table of principal characteristics of sandur depositional zones.

Proximal zone	Fewest numbers of channels, typically deep and narrow, and high in energy and transport capacity. Rapid sediment deposition.
Intermediate zone	Network of wide, shallow braided channels and ephemeral bars. Frequent changes in channe1 position in response to changing fluctuations in meltwater discharge.
Distal zone	Shallow, poorly-defined channels. Flow may merge into sheet floods during high discharge.

Outwash Terraces:

Extensive sandurs rarely remain intact, even during their formation, and are subject to dissection and incision as the proglacial river cuts down through the sediments, though little is known of the flow conditions that trigger incision and form terraces (Gray

1991; Bennett and Glasser 1996; Maizels 2002), and result in the formation of outwash terraces (Figure 2.2A, B). Outwash terrace sequences develop from the erosional remains of many different sandur levels. The surfaces are frequently altitudinally matched along the valley side (Gray 1991; Bennett and Glasser 1996). Dissection of the outwash plain into terraces is related to subsequent meltwater dissection as the ice-margin retreats but may also be related to a drop in a local glacial lake level or to falling sea level (Gray and Sutherland 1977; Bennett and Glasser 1996).

The presence of extensive proglacial outwash sequences is indicative of active ice, which has continued to supply sediment and meltwaters to its margins (Brown, 1993). Within the Tyne Valley, Lunn (1995) suggested that outwash sequences associated with dead-ice topography extend along most of its length, indicating both active and stagnant ice during formation. It may be that in the Tyne Valley the outwash sequence became terraced due to an actively retreating ice-margin.

Glacial Lakes:

Glacial lakes and their deposits record aspects of the deglacial history of an area (Ashley, 2002). When the existing literature from the study area was re-examined in chapter 1, it was evident that some studies had recognised a large lake, >40km^2 (Glacial Lake Wear; cf. Smith, 1994), within the study area, as well as small localised lakes and deltaic landforms. Therefore, the final part of this review focuses on the landform-sediment assemblages likely to develop in conjunction with glacial lake sedimentation that may still be visible in the contemporary landscape.

Lakes may develop in ice contact (e.g. supraglacial, englacial or marginal) and distal (i.e. non-contact) settings, and originate in a number of ways: (1) ice erosion of bedrock; (2) impeded drainage due to moraine, sediment or ice dams; (3) depressions in glacigenic deposits; and (4) valley glaciation, forming long narrow basins (Smith and Ashley 1985; Ashley 2002). The style of origin has an impact on the landform-sediment assemblages that develop (Bennett and Glasser 1996; Ashley 2002). Glacial lakes range in size from a few square metres to >100,000km^2, such as Glacial Lake Agassiz, which formed during deglaciation of the Laurentide ice sheet.

The two main landforms associated with glacial lakes are deltas and subaqueous fans. Ice-contact deltas and subaqueous fans mark the position of the ice margin, so are important for reconstructing patterns of deglaciation (Ashley, 2002). Deposition takes places as sediments enter the lake from meltwater streams as overflow (i.e. sediment is less dense than surface water so rises to the surface), interflow and underflow (Figure 2.4A). Deltas form at both active and stagnant ice-margins, either fed from subglacial or englacial tunnels or by meltwater streams originating in active ice but flowing over stagnant ice and into the lake, where sedimentation eventually builds to the lake surface (Ashley, 2002). In distal lakes, Gilbert-type deltas (cf. Gilbert, 1890) are the primary depositional landform, forming when fast moving sediment-laden streams enter a standing body of water resulting in a decrease in velocity and the capacity to transport coarse sediment (Figure 2.4B).

Within the Tyne Valley, based on the existing deglacial models (stagnation and downwasting), it might be expected that marginal lakes developed along the valley sides as the ice contracted into the valley bottom. Given the lower Tyne Valley was subsequently impounded by active (coastal) ice and Glacial Lake Wear developed, lacustrine sequences must be present in the Tyne Valley. The formation of deltaic sequences may be recognisable in the Tyne Valley as active ice in the west continued to

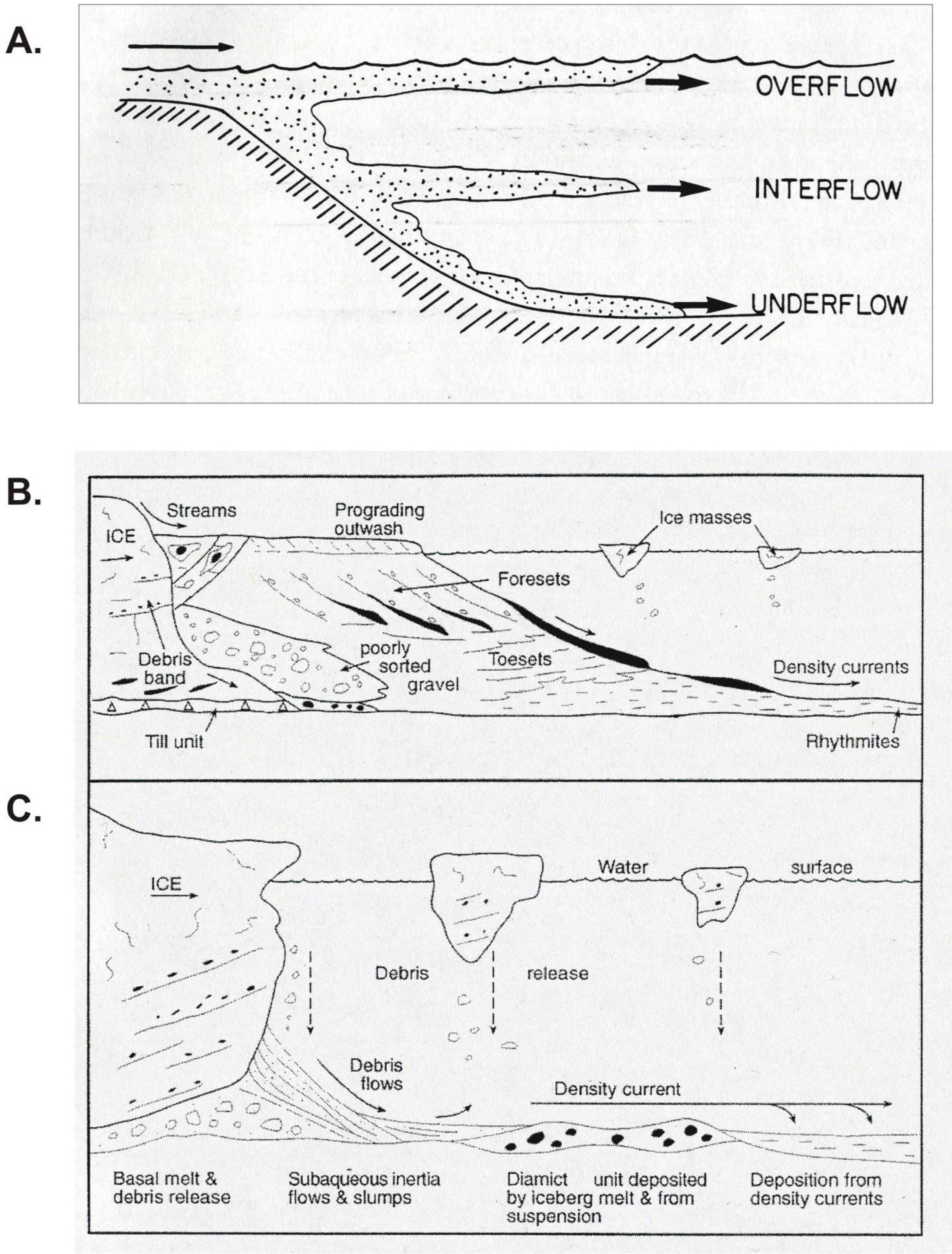


FIGURE 2.4: Sediment flow into glacial lakes, and development of deltas and subaqueous fans. A. Sketch illustrating the principal types of sediment flow into glacial lakes (taken from Smith and Ashley, 1985). B. Sketch illustrating the Gilbert-type Delta formation. C. Subaqueous fan formation in glacial lakes (taken from McCarron, 2003).

deliver meltwaters and sediment. Preservation is dependent upon limited postglacial erosion.

2.2.1.3 Summary

In general, development and morphology of glacial landforms assemblages is controlled by proximity to ice, and features may be underlain by both glaciofluvial and glaciolacustrine sediments. Prolific aggradation, such as sandur plain sequences often bury earlier depositional landforms, such as moraines, eskers, and ice contact features (kames), masking the pattern of deglaciation. Furthermore, glacigenic landforms should not be interpreted based simply on their contemporary surface expression. Indistinct surface morphology is often the result of subsequent degradation due to post-depositional erosion and not related to landform genesis. Thus, understanding the sub-surface sedimentary sequence, as well as surface morphology, is important if landforms are to be correctly interpreted.

2.2.2 Ice-contact sediment facies

Eskers and subaqueous fans:

Facies within eskers represent the depositional conditions that existed within the tunnel and at the outlet, and these can be highly unpredictable; therefore, esker sedimentology is highly variable, localised and comprises both fluvial and lacustrine sequences. Whilst generalisations are difficult, many eskers comprise a core of poorly sorted gravel overlain by sorted sand and gravel (Bennett and Glasser, 1996). Finer grained sand and silt is usually carried beyond the tunnel to be deposited at the mouth or forms overbank flood deposits in ice-walled channel eskers (Bannerjee and McDonald 1975; Warren and Ashley 1994). Some key characteristics have been identified from late

Pleistocene eskers. Single ridge eskers (long and short beaded) comprise well-sorted, clast-supported, open-work cross-bedded (dipping at ridge orientation) sands and gravels, with planar bedded sands in between and silt drapes on some cobbles overlain by matrix-supported gravels. Ridge flanks comprise cross-beds of silt to medium sand, ripples, small trough cross-bedding, climbing ripples, de-watering structures (flame) and clay drapes (Banerjee and McDonald 1975; Warren and Ashley 1994; Glanville 1997). Esker depositional sequences represent the different flow regimes operating in the tunnels (see section 2.2.1). Sequences range from well-sorted cross-beds, to clast-supported boulder gravel, matrix-supported boulder gravel, openwork gravel with no apparent change in size from bottom to top of the sequence (cobble and boulder clusters are prolific) or completely unsorted glacigenic debris deposited when the highest flow velocity drops rapidly (Warren and Ashley 1994; Ashley and Warren, 1997). Supraglacial or ice-walled esker ridges tend to comprise more fluvial-type facies, characterised by channel gravels, gravel bars and overbank fines (Shaw 1972; Warren and Ashley 1994). These types of esker can be differentiated from tunnel deposited eskers based on presence of features of subaerial origin, such as mud cracks. Furthermore, the presence of antidune bedding is indicative of open channel flow as surface water waves do not form in tunnel flow, although there are exceptions (cf. Banerjee and McDonald, 1975).

Sedimentologically, subaqueous fans (beaded eskers) can be distinguished into proximal, mid and distal facies (Figure 2.4C). The transition between environments may be as little as 100m, and sequences often have high vertical and lateral variability (Rust 1977; Paterson and Cheel 1997; Delaney 2002). Proximal facies comprise poor to moderately sorted, massive to weakly stratified, cobble and pebble gravel, with intercalated sands and climbing ripple cross-laminated sand and silt and clay drapes

(Eyles and Clark 1987; Fard *et al*. 1997; Delaney 2002; Ashley 2002) and are attributed to high discharge, debris-laden flows (cf. Baker 1978; Maizels 1989; Carling 1996). Rapid deposition in the proximal area is reflected by chaotic bedding with large clasts dispersed in sand and associated dewatering structures due to sudden loading of waterlogged sediments and expulsion upwards (Cheel and Rust 1986; Delaney 2002). Mid fan facies comprise stratified pebble gravel and sands, mass flow deposits, dune and climbing ripple/ripple cross-bedding silty sand, deformation structures (e.g. ball and pillow structures) and occasional dropstones (Rust 1977; Postma *et al*. 1983). The facies are attributed to subaqueous sheet flow, high sediment concentration density underflows and lower, less competent flows and suspension settling and ice rafting (Rust 1977; Cheel and Rust 1986; Sharp 1988; Lunkka and Gibbard 1996; Paterson and Cheel 1997). Distal fan facies comprise cyclic fining-upward sequences of massive or laminated fine sands, climbing ripples and silt and clay drapes, which eventually become indistinguishable from lake bottom sediments (see section 2.2.2.2) (Delaney 2002; Ashley 2002).

Fans are often buried by subsequent lake sedimentation and are more likely to be identified through their sedimentology recorded in borehole logs rather than through surface expression. Esker ridges are easily identified through their surface morphology; their sedimentology is indicative of flow conditions at deposition, and enables the differentiation between tunnel and open channel formation. Based on the mode of deglaciation in the Tyne Valley, and the presence of lakes, eskers are likely to be found in association with lacustrine sequences, should be easily identifiable by their distinctive morphology (sinuous ridges, bifurcating pattern) and probably occur in association with other ice-contact (kames) landforms.

Deltas and glaciolacustrine facies:

Lake sedimentation reflects both marginal landforms such as deltas and standing water sedimentation, and the facies that develop represent both ends of the continuum. As discussed in section 2.2.1, sedimentation in lakes is influenced by water density stratification, primarily related to temperature but also suspended sediment concentration that develops within the lake. Thermal stratification (Figure 2.5A), common in distal lakes, results in a lower density top layer (epilimnion) above a colder, higher density bottom layer (hypolimnion). Sediment stratification (Figure 2.5B), common to ice-contact lakes, may develop as sediments loads are high and the increase in suspended sediment concentration with depth leads to density separation.

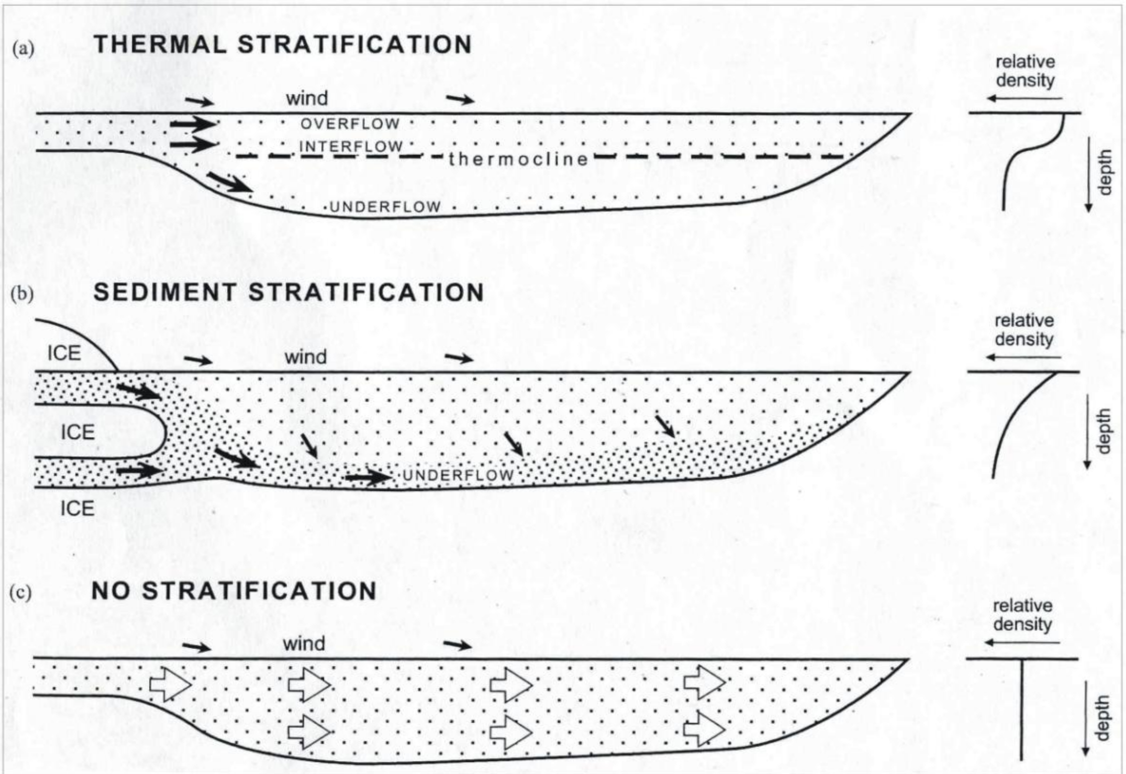


FIGURE 2.5: Density stratification in ice-contact lakes. A. Thermal stratification. B. Sediment stratification. C. No stratification. (taken from Ashley, 2002).

Sediment accumulates subaqueously on the slope of the lake margin, with the finest sediments being carried further down the slope into the deeper water (Smith and Ashley 1985; Ashley 1995). Deltaic sequences typically comprise three units: topsets, foresets (dips up to 33˚) and bottomsets, which represent alluvial, delta slope and delta toe deposits respectively (Smith and Ashley 1985; Johnson and Maizels 2002) (Figure 2.5B). Where there are pronounced underflows (Figure 2.5A), the sediment-laded meltwater rapidly descends into the lake as a series of turbidity currents, which may inhibit the build up of large deltas. Persistent underflow results in a series of subaqueous fans (discussed above) in front of a small delta (Bennett and Glasser, 1996). The rate of delta progradation is controlled by: (1) sediment supply (related to proximity to ice, which decreases as the ice melts); and (2) lake basin depth (deep basins are associated with ice contact lakes and shallow basins with distal lakes) (Smith and Ashley, 1985). Distal lake deltas tend to comprise finer grained and better sorted sediments in comparison with ice contact lakes (Johnson and Maizels, 2002). Topsets form when meltwater enters the lake, depositing coarse sediments due to the abrupt drop in velocity as the stream meets the lake. Typically, they comprise matrix supported or open-work (imbricated) massive or crudely bedded poorly sorted cobble to boulder gravel, medium to coarse sands (Shaw 1972; Brown 1994; Benn 1996). As the delta emerges from the lake, normal fluvial processes dominate as braided rivers flow over the surface. Sheet-flood gravels are deposited as extensive longitudinal bars during high but relatively shallow flow (Boothroyd and Ashley 1975; Church and Gilbert 1975; Thomas 1984; Brazier *et al*. 1998). Laminated sands and rippled silty sands develop in association with gravel units through sand bar or mega ripple migration. Scour structures, probably cut by high discharge events or relocation of distributary channels, are infilled with gravels representing low relief longitudinal bars (Thomas 1984; Brown 1994; Maizels 2002). Ice-contact or proximity is reflected through intercalated diamicton (McCabe *et al.,* 1987; McCabe and Eyles 1988; Lønne 1993).

Coarse-grained sediments carried beyond the edge of the topsets and into the lake basin by underflows (either quasi-continuous and surge currents, and intermittent slope avalanching), give rise to the upper delta foresets (Smith and Ashley, 1985). Foresets are usually steeply-dipping (20-30˚), tabular bedsets (clinoforms) that are commonly rhythmically bedded, and fine upwards from boulder and pebble gravel to coarse sand (Gustavson *et al*. 1975; Thomas 1984; Smith and Ashley 1985; Brown 1994; Benn 1996; Bennett and Glasser 1996; Brazier *et al*. 1998). Mid delta slope foresets (<10˚) form from rapid sedimentation from density underflows carrying a turbid mix of sand, silt and clay. Facies mostly comprise sands deposited as type B climbing ripples capped by silt drapes deposited under waning flows (Ashley *et al*. 1982; Nemec and Steel 1984; Smith and Ashley 1985; Eyles *et al*. 1987). Finally, on the lower delta slope, lower delta foreset rhythmites (i.e. silt-clay sequences, classic 'varves') develop that are almost indistinguishable from lake bottom facies. Bottomset facies comprise rhythmites, often intricately folded and faulted with water escape features distorting the coherence of the beds (Smith and Ashley 1985; Benn 1996; Glanville 1997; Van der Meer and Warren 1997; Spooner and Osborne 2000; Ward and Rutter 2000; Hambrey *et al.* 2001; Maizels 2002).

Sediments deposited in the lake basin are the equivalent of lower delta foreset rhythmites. They can be deposited at any stratigraphic level within a lake, and are the main facies that are associated with lakes. Coarse laminae are deposited from quasi-continuous summer underflows, whereas fining up clays represents sedimentation during the rest of the year by overflow-interflow currents. Annual rhythmites are thin, regularly bedded and contain no current structures, whereas rhythmites generated from surge currents represent rapid, continuous deposition. They are differentiated from annual rhythmites because there are no breaks within each rhythmite and the thickness of each layer is proportional (Smith and Ashley, 1985). Rhythmites may be deposited as climbing ripples, multiple normally graded layers, and thick multi-laminated beds with occasional clay layers (Smith and Ashley 1985; Maizels 2002). Where lake basin deposits are interspersed with gravels at the lower delta front/lake bottom interface, they are interpreted as traction current deposits or due to increased flow at a low lake stand (cf. (Ashley 1975; Clemmensen and Houmark-Nielsen 1981; Eyles *et al* 1987; Brazier *et al*. 1998; Spooner and Osborn 2000; Ward and Rutter 2000).

The occurrence of dropstones (i.e. individual boulders or gravel pods) within laminated or rhythmically bedded sediments is indicative of lake deposition and rainout from iceberg rafting in the lake (Benn 1996; Hambrey *et al.* 2001). Diamicton and chaotically bedded gravels within the lake basin sequence represents debris flows issuing from the ice margin or subaqueous mass-flow deposits due to slumping (McCabe *et al*. 1987; McCabe and Eyles 1988; Lønne 1993; Brazier *et al*. 1998; Ward and Rutter 2000).

The mode of deglaciation, and previous work, suggests the existence of lakes during deglaciation. Sequences on the valley sides, above the Holocene valley floor (HVF), may contain records of lacustrine sedimentation. Borehole logs may provide further information on the sequence in the valley bottom.

2.2.3 Proglacial sediment facies

Outwash fan and sandur (braid plain) facies:

Deposition at the ice-margin can be a chaotic affair as sediment-laden meltwaters debouch from the ice-front and develop into proglacial outwash fans and braid plains (Bennett and Glasser, 1996). Sedimentation of the fan can be divided into three zones, comprising an upper, mid and lower fan facies (Figure 2.6) and reflect a proximal to distal assemblage (Boothroyd and Ashley, 1975). The main morphological features associated with outwash fans and sandur plains are: (1) channels (pools and riffles) and interchannel areas (bars); and (2) transverse ribs (McDonald and Banerjee, 1971). Facies change from coarse gravels deposited in single, deep channels to sands associated with bar migration deposited in braided streams, to a more meandering system, dominated by sand (Figure 2.7). The distinction between outwash fan and braid plain is not clear, and the braid plain essentially develops by extension of the outwash fan, thus facies in the fan and braid plain are the same. The proximal to distal transition reflects a spatial relationship, where facies change systematically from an arbitrary proximal location (i.e. ice-margin, outwash fan apex, meltwater source point) to a point some distance away.

Proximal zones are associated with intercalated diamicton indicative of debris flows off the ice, and burial and faulting (syn- or post-sedimentary faults, synclinal fold or sag) through melt-out as the fan progrades (Fraser 1993; Thomas and Montague 1997). The proximal zone is characterised by high magnitude-low frequency floods and falling stage flows (Boothroyd 1972; Gustavson 1974; Boothroyd and Ashley 1975), and thus the proximal (upper fan) facies are dominated by large-scale tabular cross-stratification (longitudinal bars) cobble and boulder gravel, and sandy gravel, intercalated with sheet-like beds of sandy pebbles (infilled large channels), and infrequent lenses of parallel

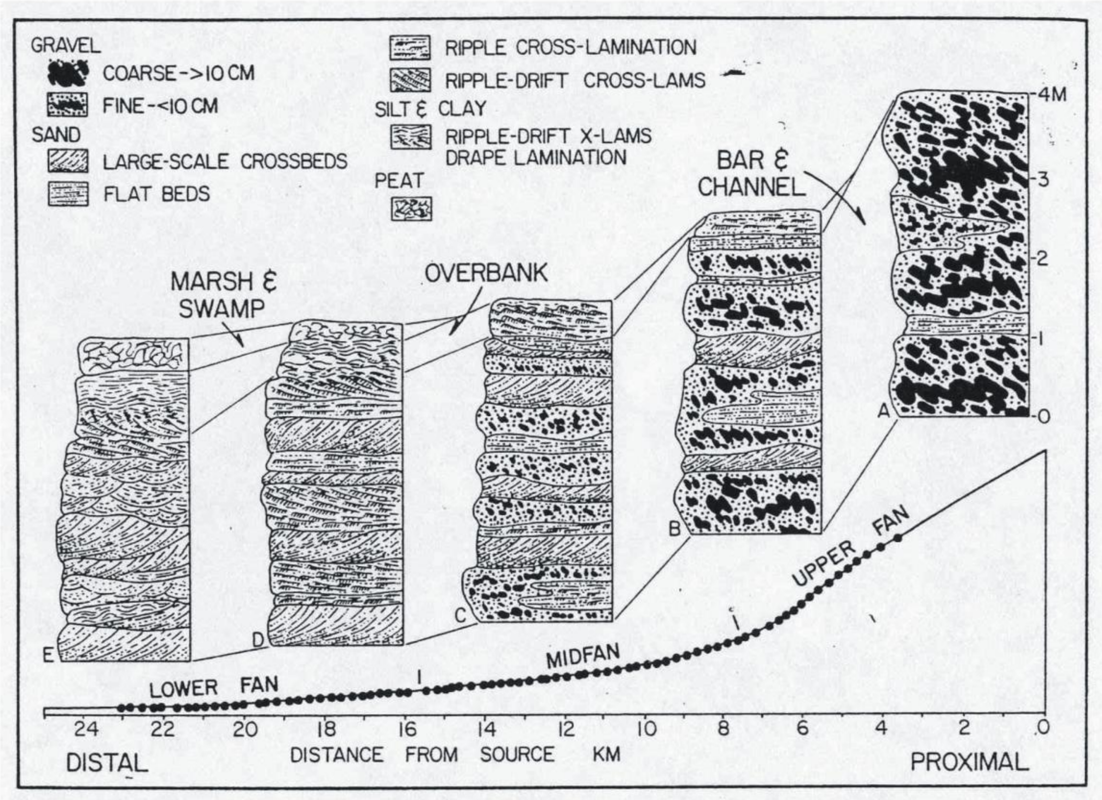


FIGURE 2.6: Sketch illustrating downstream variation in facies and sedimentary structures of an outwash fan and braid plain (taken from Boothroyd and Ashley, 1975).

laminated and rippled sands (McDonald and Banerjee 1971; Costello and Walker 1972; Eynon and Walker 1974; Smith 1974; Rust 1975; Miall 1977; Bluck 1979; Fraser and Cobb 1982; Morrison and Hein 1987; Thomas and Montague 1997; Aitken 1998; Zielinski and van Loon 2003; Knight 2006). The intermediate (mid fan) zone is dominated by is upper flow regime, shallow sheet flows which merge at high discharges in shallow, wide channels, and thus the facies exhibit a marked downstream decline in maximum clast size, and sedimentation takes place in shallow sandy channel systems, (Boothroyd and Ashley 1975; Thomas and Montague 1997). Facies comprise sand and fine gravel in small-scale bedforms reflecting the development of longitudinal, linguoid or transverse and lateral, side or point bars. Scour channels are infilled with gravel and pebbly sand, and ripple laminated silty sand aggraded under low flows (particularly in backwaters). Overbank facies reflect falling stage flood flows, and comprise horizontally bedded or massive sands, cross-bedded pebbly sand, and climbing ripples (McDonald and Banerjee 1971; Costello and Walker 1972; Bluck 1974; Eynon and Walker 1974; Miall 1977; Fraser and Cobb 1982; Thomas *et al*. 1985; Aitken 1998). The distal (lower fan) zone comprises progressively fining sediments, is almost exclusively dominated by sand and represents a gradual transition to a more conventional fluvial system (Bennett and Glasser, 1996). Under low flow conditions, meltwater is concentrated in a single channel, with braided streams developing during peak flows. The distal facies comprises large-scale troughs infilled with gravely sands, and ripple laminated sand and silts formed through migration of longitudinal bars, bar top aggradation during moderate and low stage flows and through overbank deposition (Boothroyd and Ashley 1975; Miall, 1977; 1978; Rust, 1978; Fraser and Cobb 1982; Thomas and Montague 1997; Aitken 1998; Maizels 2002; Zielinski and van Loon 2003; Knight 2006).

Given the dynamic nature of the proglacial environment, sedimentation in any of the zones may be disrupted by events of low frequency and high magnitude, such as catastrophic glacial outburst floods (i.e. jökulhlaups). Therefore, facies may include massive, poorly sorted, non-graded or inversely graded sediments attributable to extreme events (cf. Maizels 2002; Marren 2002).

It was suggested in chapter 1 that the western ice withdrew from the main Tyne Valley during deglaciation, thus an extensive braid plain must have developed if the western ice remained active, allowing the ice-free zone in the LTV to infill with sediment. Sequences may extend for some distance up valley, dependent upon subsequent destruction by postglacial reworking or quarrying.

2.2.4 Lithofacies models

To understand glaciofluvial sedimentation both Miall (1977, 1978) and Rust (1972, 1978) have proposed depositional models for braided river deposits based on modern and ancient sequences. These models reflect proximal to distal depositional changes in sedimentology and have been widely applied to sequences in proglacial zones. Miall's (1978) lithofacies models recognise six principal facies assemblages in gravel- and sand-dominated braided rivers (Figure 2.7). The Trollheim (cf. Rust, 1978) and Scott models (cf. Boothroyd and Ashley, 1975) are proposed for proximal braided stream and fan deposits, reflecting debris flows. The Donjek model (cf. Williams and Rust 1969; Rust 1972) reflects most types of cyclic gravel deposits, and the South Saskatchewan type (cf. Cant and Walker 1976; Cant 1978) comprises cyclic sand sequences. The Platte model (cf. Smith 1970; Boothroyd and Ashley 1975) contains non-cyclic sequences. Finally, the Bijou Creek model (cf. McKee *et al.*, 1967) reflects high energy flow conditions, and flash floods in sandy braided rivers. The latter types reflect more distal locations but the exact location of the Platte and Bijou creek types is not clear. By applying these models to the sediment sequences under investigation, the sequences can be more fully understood in terms of their depositional environment.

2.2.5 Summary

The origins of the landform-sediment assemblages in the Tyne Valley are currently undetermined, and conceptual models have been put forward as scenarios to be tested. Sedimentologically, glacigenic sequences are complex and represent a number of depositional episodes rather than a single event (cf. Thomas, 1985). It is widely acknowledged that distinguishing between glaciofluvial and glaciolacustrine sediment

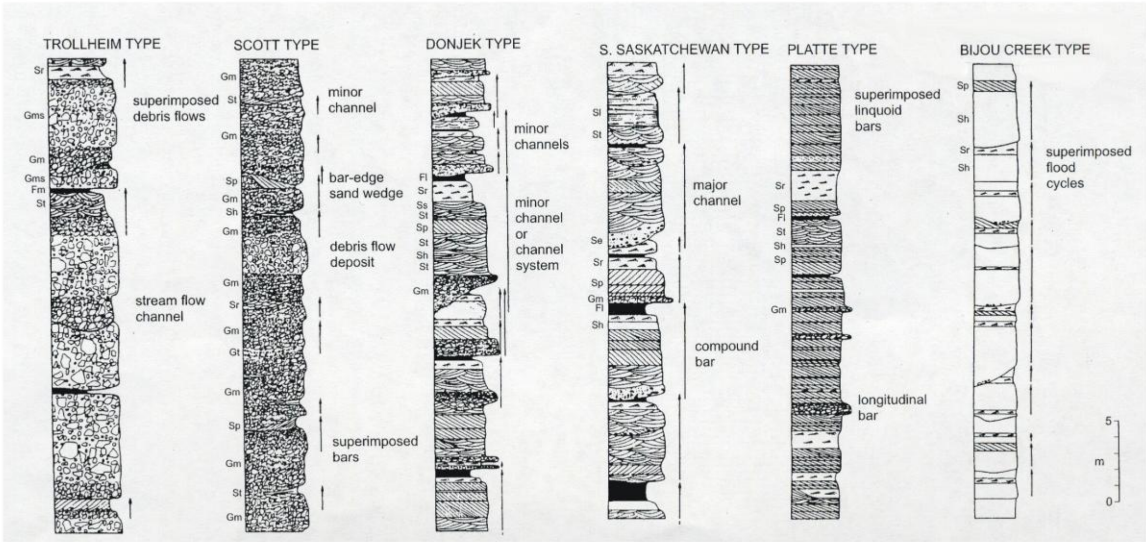


FIGURE 2.7: Lithofacies models. Miall's (1978) vertical facies profiles for braided stream deposits (taken from Miall, 1978).

facies is difficult (cf. Bryant, 1991) and it is clear that surficial morphology masks the underlying sub-environments, so that interpretation based on morphological expression alone is futile. Of course, any interpretation presented is still only the most plausible when all other possibilities have been excluded. This review of the principal landforms and sediment facies types associated with ice contact, ice marginal and proglacial environments, provides a context in which the landform-sediment assemblages which outcrop along the Tyne Valley can be interpreted. Critical examination of the Tyne landform-sediment assemblages, both morphologically and sedimentologically through mapping out outcrops and supplemented with borehole date, with reference to the detailed sedimentary models presented in this review, will allow the conceptual models of Lateglacial development to be tested.

2.3 Post-glacial events

Following deglaciation, landscape development in response to postglacial climate change in river valleys is evidenced by the creation of river terraces and cut and fill alluvial units (*sensu* Lewin and Macklin, 2003), which reflect changes in previous river regimes. In a review of British rivers, Lewin and Macklin (2003) differentiated the pattern of postglacial fluvial response between upland glaciated and lowland unglaciated catchments (Figure 2.8). They suggest two phases of valley floor aggradation between deglaciation and the start of the Holocene: (1) during the immediate postglacial, and related to deglaciation, with proglacial and paraglacial aggradation; and (2) during the Younger Dryas Stadial. Incision followed both periods of development, leading to the formation of terraces, underlain by gravelly units, identified in both uplands and lowlands in the UK that were formerly glaciated by Lewin and Macklin (2003). Whilst the work by Macklin and others provides a framework for interpreting Holocene alluvial sequences, it is interesting to note that in those catchments that lie beyond the limits of the Devensian ice sheet, e.g. the Trent Valley, incision and terrace formation is not a feature of Holocene valley floor development (cf. Maddy 1999; Brown 2001; Howard et al. 2007). This suggests continued incision throughout the Holocene may be driven by isostasy in former glaciated catchments rather than just climate driven changes. The pattern of fluvial response in Scotland suggests proglacial aggradation of sediments during deglaciation followed by incision, driven by a combination of climate amelioration, isostatic uplift and baselevel (sea level) change (cf. Maizels and Aitken, 991). Up to five outwash terraces have been recorded in upland Scottish valleys (e.g. Dee, Don, Esk) that developed prior to the Bølling-Allerød period, and are related to paraglaciation. The sequences in England are less well understood, through a combination of preservation issues, poorly resolved chronologies and limited work on identifying Lateglacial terraces.

2.3.1 Mechanisms and controls on river terrace development

River terraces represent landforms created only through downcutting by the river. They represent former floodplains of the river that are now elevated or buried above the present river level and are no longer actively reworked (Petts and Foster 1985; Merritts *et al*. 1994). The cause of terrace development may be related to extrinsic controls, such as baselevel change (eustatic or local), climate change (discharge and sediment supply), glaciation (proglacial or periglacial drainage basins), tectonic activity (isostatic uplift), anthropogenic disturbances (land use or clearance) or intrinsic controls which result in the crossing of geomorphic thresholds and force change in the fluvial system.

Over the long term (glacial/interglacial), the response of any system (fluvial, glacial, etc.) to extrinsic or intrinsic control is dependent upon the sensitivity of the system and its propensity for change (cf. Brunsden and Thornes 1979; Brunsden 1993). Change in any system is initiated by high magnitude-low frequency events (cf. Wolman and Miller, 1960), however, the response to change can be complicated by the system's proximity to geomorphic thresholds (cf. Schumm, 1973) or by local factors (Vandenberghe, 1995). Within a fluvial system, each reach of a valley experiences different responses to both intrinsic and extrinsic factors (Brown 1991; Ballantyne and Whittington 1999). This is embodied in Schumm's (1973) complex response, which explains that there is a spatial variation within the basins where the river may experience both incision and aggradation simultaneously. Thus, trying to disentangle the mechanism(s) which have driven change across a river basin, when response is not a simple one, becomes fraught with difficulties because of the number of possible variables, which are not mutually exclusive and may compound the response. Furthermore, in terms of landscape development, it is also important to consider the length of time the processes are operating and the time for landscape adjustment is of a

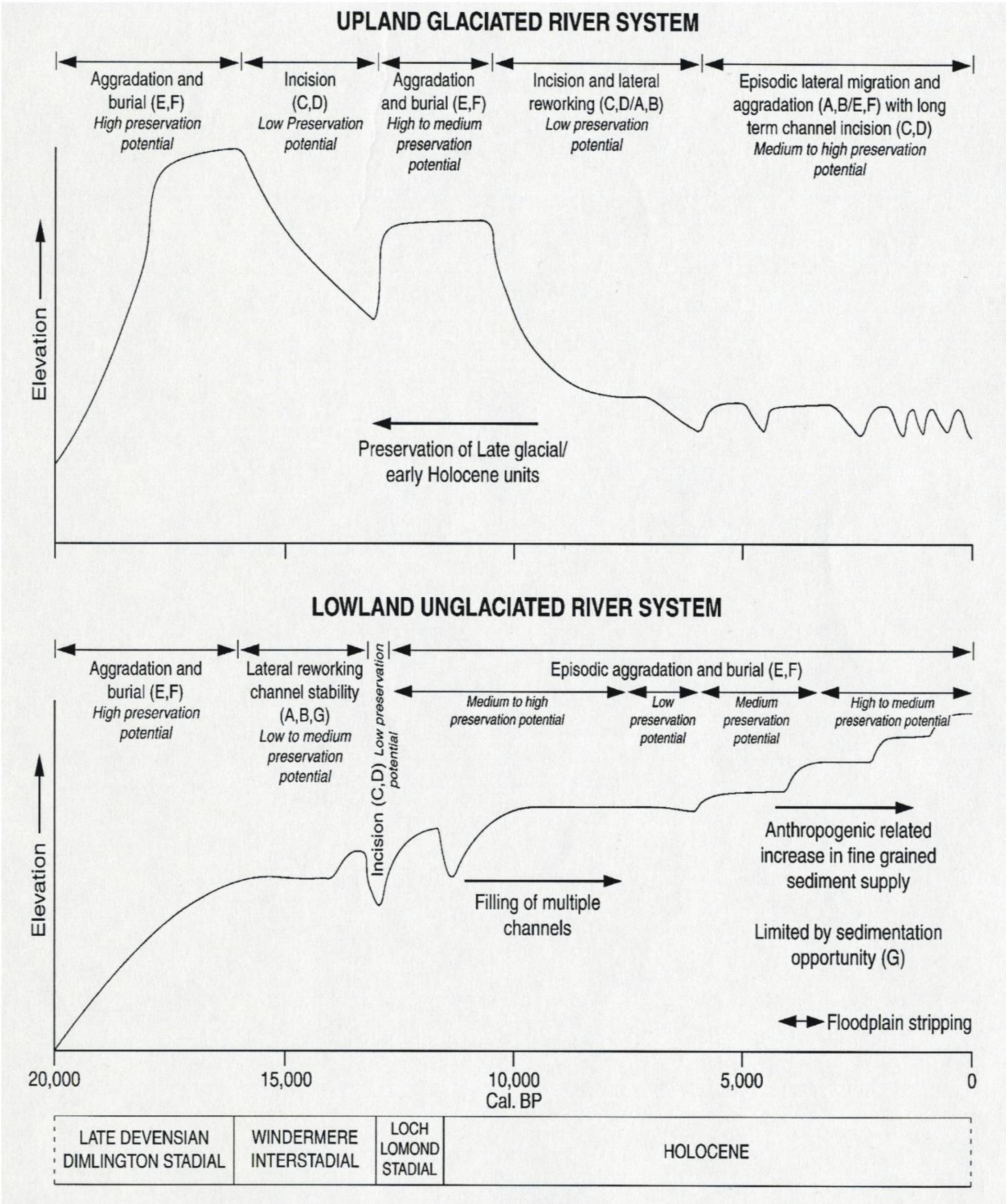


FIGURE 2.8: Response of upland and lowland rivers following deglaciation. An interpretation of alluvial unit production and preservation for the UK, over the last 20,000 years (taken from Lewin and Macklin, 2003).

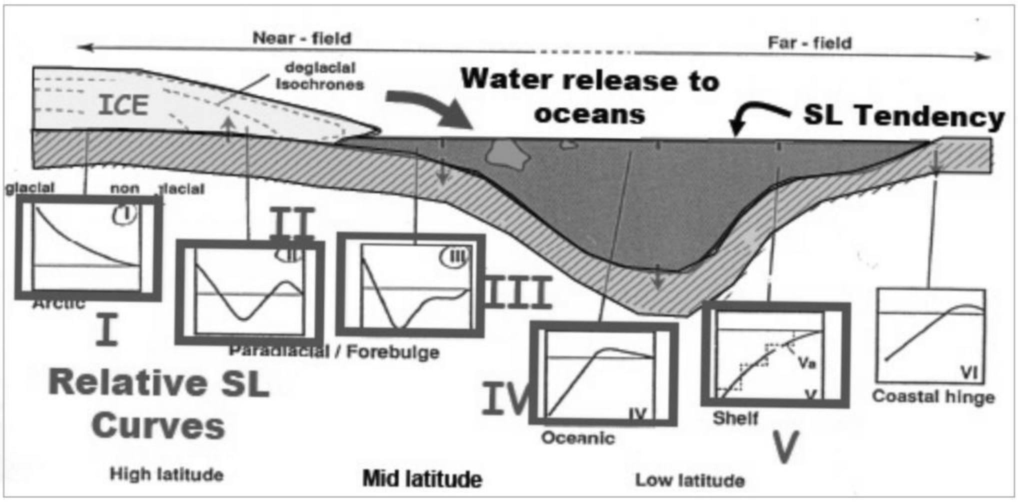
similar order of magnitude to those needed for external forces (e.g. scale of warm-cold cycle ~10k years). Relaxation time (i.e. the time taken to attain new landforms) is determined by the rate of change; however, landscapes can be described as 'over-relaxed' i.e. the previous conditions (e.g. glacial) have accomplished so much work in terms of landforms that the present (interglacial) conditions achieve little/no morphological change (Brunsden, 2001). Thus, in upland glaciated river basins, the landscape response (i.e. incision and terrace formation) following deglaciation and throughout the Holocene may be part of continuing adjustment to the previous glacial conditions rather than the prevailing contemporary ones.

2.3.1.1 Base-level

Base-level is the theoretical horizon to which rivers erode i.e. sea level (Schumm, 1993). A change in base-level is induced by a eustatic sea level fall or rise, a change in a local (glacial) lake level, or by isostatic processes (uplift or depression), which may cause the river to respond through aggradation or incision in order to adjust its long profile to the new base-level. The influence of base-level change on a river system is controlled by a number of factors; these include: (1) the direction, magnitude, rate and duration of change; (2) catchment geology (bedrock or structural); (3) catchment geomorphology (i.e. wide or narrow valley); and (4) sensitivity (Schumm, 1993). A change in base-level may result in the river profile being out of grade; and, which, however, is dependent on the factors above and how the river is able to adjust/accommodate the change. Where river terraces are preserved they could possess longitudinal profiles that are graded to former base-levels, and thus used to estimate former sea level (Törnqvist, 1998). Although Schumm (1993) points out that except under extreme conditions (not defined), changes in baselevel will not rejuvenate the entire drainage network. Thus, the effect of baselevel change is not usually demonstrable throughout the catchment, and in some cases does not extent beyond the coastal fringe. Blum and Törnqvist (2000) defined the upstream limit of baselevel change as the intersection between the modern floodplain and the Lateglacial floodplain.

During the late Quaternary, global sea level fell to -121±5m OD at the LGM (Fairbanks, 1989). Following the onset of deglaciation, global sea level began to rise, however, recorded sea level change at a location is affected by the relationship to the ice sheet, and whether the locality lies inside, at or beyond the margin of the ice (Figure 2.9A). Crustal deformation and displacement extends beyond the immediate environs of the ice margin, thus the response to eustatic sea level change is spatially and temporally variable dependant on the location relative to the ice sheet. Thus, the sea level at a location (and thus baselevel) is complicated by other factors, and so relative sea level (RSL) reconstruction provides an indication of local sea level. The general pattern of sea level change around Britain was modelled by Lambeck (1993; 1995), taking into account both glacio-isostatic and glacio-hydrostatic factors. The models predicts sea level stand still following the onset of deglaciation as the North Sea basin continued to emerge due to the delayed response of the mantle to ice and water loading (15-12k yrs BP). At 10k yrs BP, the North Sea began to advance southwards and eustatic sea level rise is evidenced by marine incursions in the southern North Sea. A relative sea level fall is then predicted due to isostatic rebound becoming more significant than eustatic sea level rise after 9k yrs BP. Global sea level continued to rise until ~6k yrs BP (Shennan 1992; 1999; Shennan et al. 1993; 1994; 1995; 2000). There is good evidence of sea level change around the British coastline, such as raised shorelines in Scotland (e.g. Sissons 1976; Cullingford and Smith 1980; Dawson 1984; Cullingford et al. 1991); submerged forests and fresh water peats along the northeast and northwest coastlines of England (e.g. Smith and Francis 1967; Gonzalez and

A.



B.

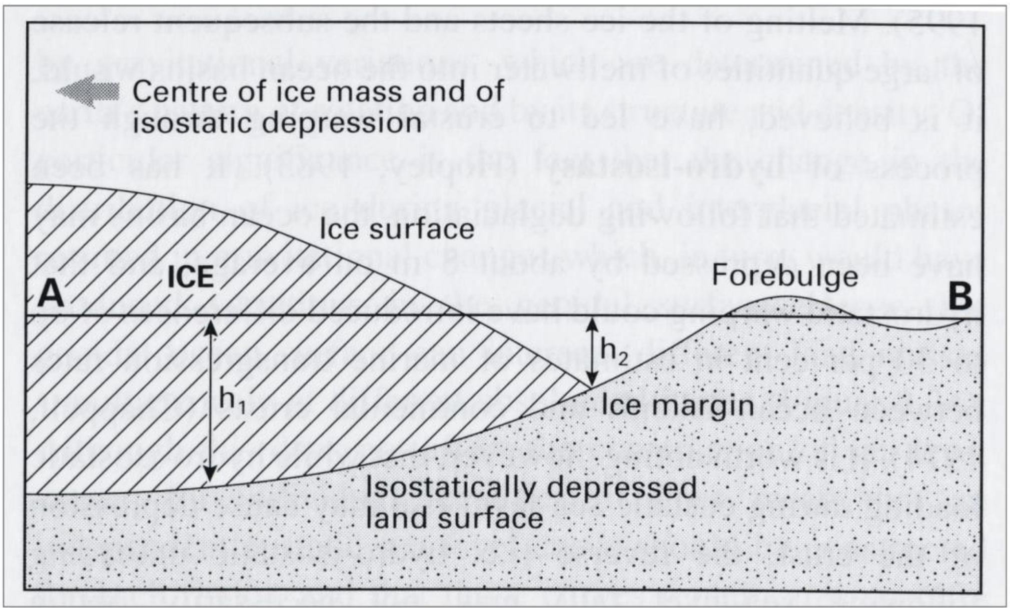


FIGURE 2.9: A. Sketch illustrating the effect of an ice mass on relative sea level. In general terms, the amount of isostatic rebound is conditioned by proximity to the ice mass centre. Relative sea level curves reflect the interplay between rising land and eustatic sea level rise (taken from Carter, 1992). B. Sketch illustrating isostastic depression and forebuldge (taken from Lowe and Walker, 1998).

Cowell 2004; Horton et al. 1999); isolation basins on the west coast of Scotland (e.g. Shennan et al. 1993; Selby et al. 2000; Selby and Smith 2007); and estuarine sedimentary records (e.g. Horton 1999; Horton et al. 2000). These records indicate that there was considerable change in base-level.

In relation to the Tyne Valley, work on the Northumberland coastal lowlands has enabled the construction of RSL curves, which demonstrate the pattern of sea level rise since deglaciation (Figure 2.10). In general, during the Holocene sea level rose from ~5m below OD at 9ka cal. BP to ~2.5m above OD at 2.5ka cal. BP (Plater and Shennan 1992; Horton *et al*. 2000; Shennan *et al.* 2000). Differences in the RSL curves between north and south Northumberland of ~8m at 8ka cal. BP reflect the isostatic effect of the BIS on the Northumberland land surface and demonstrate a within-region variation (Horton *et al*., 1999c); a trend consistent with Lambeck's models. As Mills and Holliday (1998) observe, the effect of baselevel change on river systems and their deposits in northeast England is unknown. This thesis has the potential scope to inform the debate. Dawson and Gardiner (1987), however, urge caution, suggesting that whilst baselevel change is an attractive mechanism to explain incision and aggradation, it is extremely difficult to establish direct links between episodes of terrace formation and sea level change.

2.3.1.2 Isostatic influences

As mentioned above, baselevel change can be induced by tectonic (or isostatic) uplift or subsidence, and evidence of tectonic displacement is reflected in flights of terraces or shorelines. The response is localised and related to proximity to the ice sheet i.e. centre, margin or distant. Crustal displacement beneath the ice sheet is compensated for by forebuldge beyond the ice margin (Figure 2.9B). Following the onset of

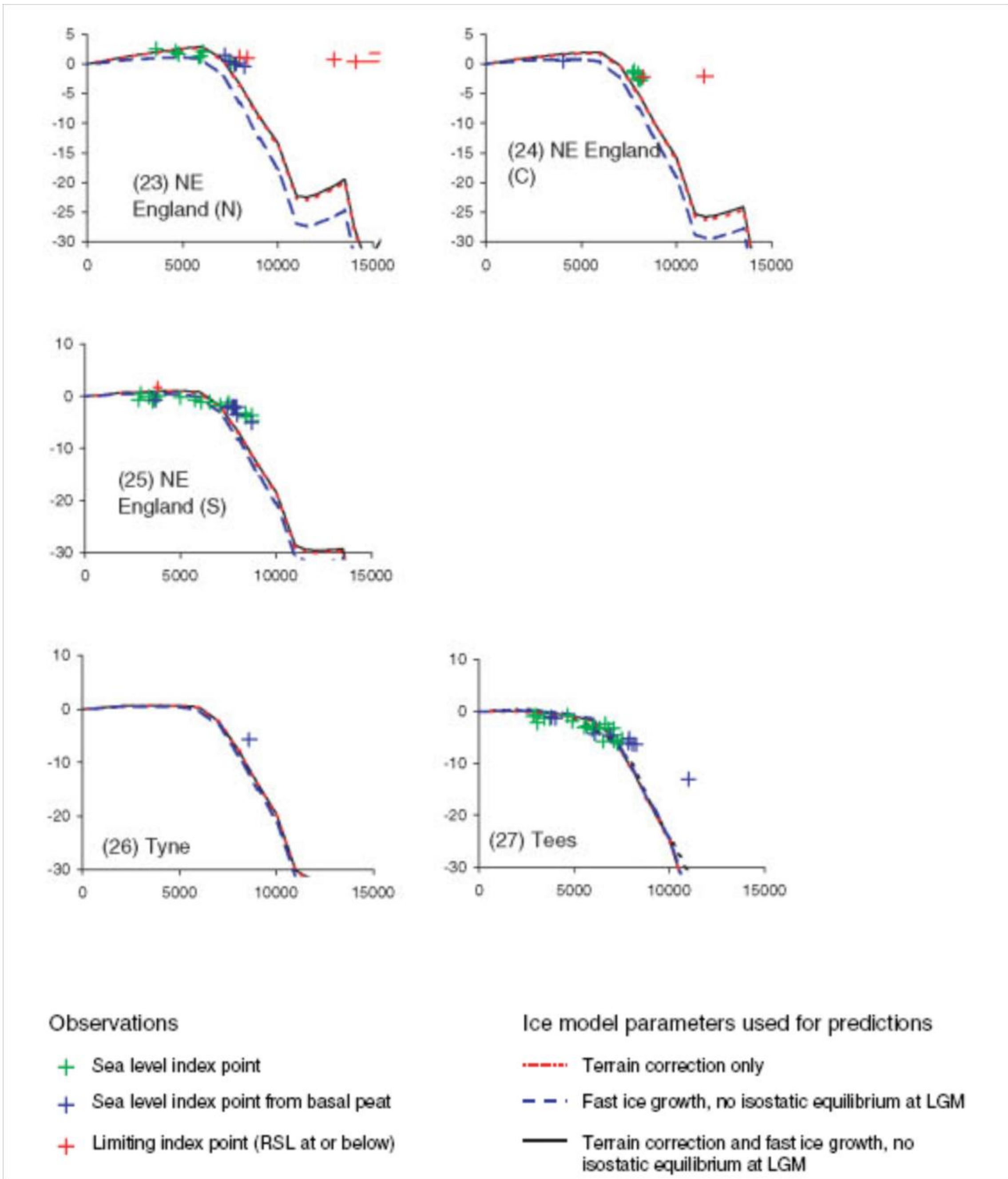


FIGURE 2.10: Model predictions for Lateglacial and Holocene relative sea level changes in Northumberland-North, –South and –Central, the Tyne Valley and the Tees Valley (taken from Shennan *et al*., 2006).

deglaciation, isostatic rebound is initially high, and this is followed by a gradual slowing until a pre-glacial equilibrium is attained (Lowe and Walker, 1997). Glacio-isostatic uplift is clearly recorded by in areas previously covered by the BIS i.e. raised shorelines around Scotland (Sutherland, 1984), and continues in highland Britain but the influence (if any) on Holocene fluvial systems has yet to be fully explored. In southern Britain, Maddy (1997) and Maddy *et al*. (2000) have suggested that the influence of regional crustal uplift following deglaciation of the MIS 12 ice sheet may have a role to play in subsequent fluvial development. They suggest that isostatic movements have been underestimated and that they continue to exert an influence on valley systems. This is clearly demonstrable in the long profiles of the Pleistocene terrace sequences that have developed in the Thames or the Severn-Avon valleys (cf. Bridgland 1983; 1994; Maddy 1989; Maddy *et al*. 1995). In terms of the last glaciation (MIS 2), little consideration has been given to glacio-isostatic uplift and terrace formation. Macklin (1999) suggests that glacio-isostatic rebound following deglaciation contributed to the formation of Lateglacial river terraces at the end of the Late Devensian and during the early Holocene, but this was not as significant as climate change. However, Evans (1999) has suggested that rivers in northeast England have continued to incise through bedrock to appreciable depths, enhanced by regional glacio-isostatic uplift. This can be demonstrated in the coastal 'denes' that are a prevalent feature of the northeast coast and are highly incised in their lower reaches. This could be related to low sea level prior to or following glaciation, or to isostatic rebound. The pattern of sea level index points in north Northumberland clearly demonstrates continued isostatic uplift through the Holocene (Horton *et al*.,1999a,c). An alternative hypothesis, therefore, is that the continued influence of glacio-isostatic uplift throughout the Holocene better explains the incision of deglacial sediments, Holocene alluvium, and river valley floors rather than waning sediment supply, climate change and anthropogenic legacies as supported by Macklin and others (Macklin 1999; Macklin and Lewin 2003; Lewin *et al*. 2005). To evaluate the influence of isostatic uplift on the Tyne terraces, long profiles need to be obtained and examined for evidence of tilting (uplift), gradient change when compared to the present day profile (baselevel change) and the existence of knick points (extent of rejuvenation). The sedimentary sequence in the lower valley fill may provide an indication of whether the coastal wedge was formed through onlapping (sea level rise) or downfilling (upstream incision) sediments. The rate and magnitude of uplift should be established (for glacio-isostatic models e.g. Lambeck, 1995) to ascertain whether baselevel change would be accommodated by the river rather than through incision. Finally, fill terraces need to be correlated from other east draining valleys in Northumberland, which would indicate whether they formed in response to the same, external control.

2.3.1.3 Climate (discharge and sediment supply)

Over the long term (~100ka), climate is considered to be the major influence on river terrace development (Vandenberghe 1995, 2002). During the Lateglacial and Holocene period, climate has changed dramatically from glacial to interglacial and consequently rivers have been affected by this. Vandenberghe and Maddy (2001) summarised the problems and issues in understanding the role of climate in fluvial development, and it is clear there is not a simple linear relationship. Climate influences discharge and runoff (i.e. precipitation) and soil cohesion and vegetation cover (i.e. sediment supply), and the various factors are intrinsically linked i.e. soil development and evapotranspiration is determined by vegetation. Fluvial response to climatic change is a lagged one, as rivers takes time to adjust their pattern and gradient to new conditions (Knox 1983; Vandenberghe 1995).

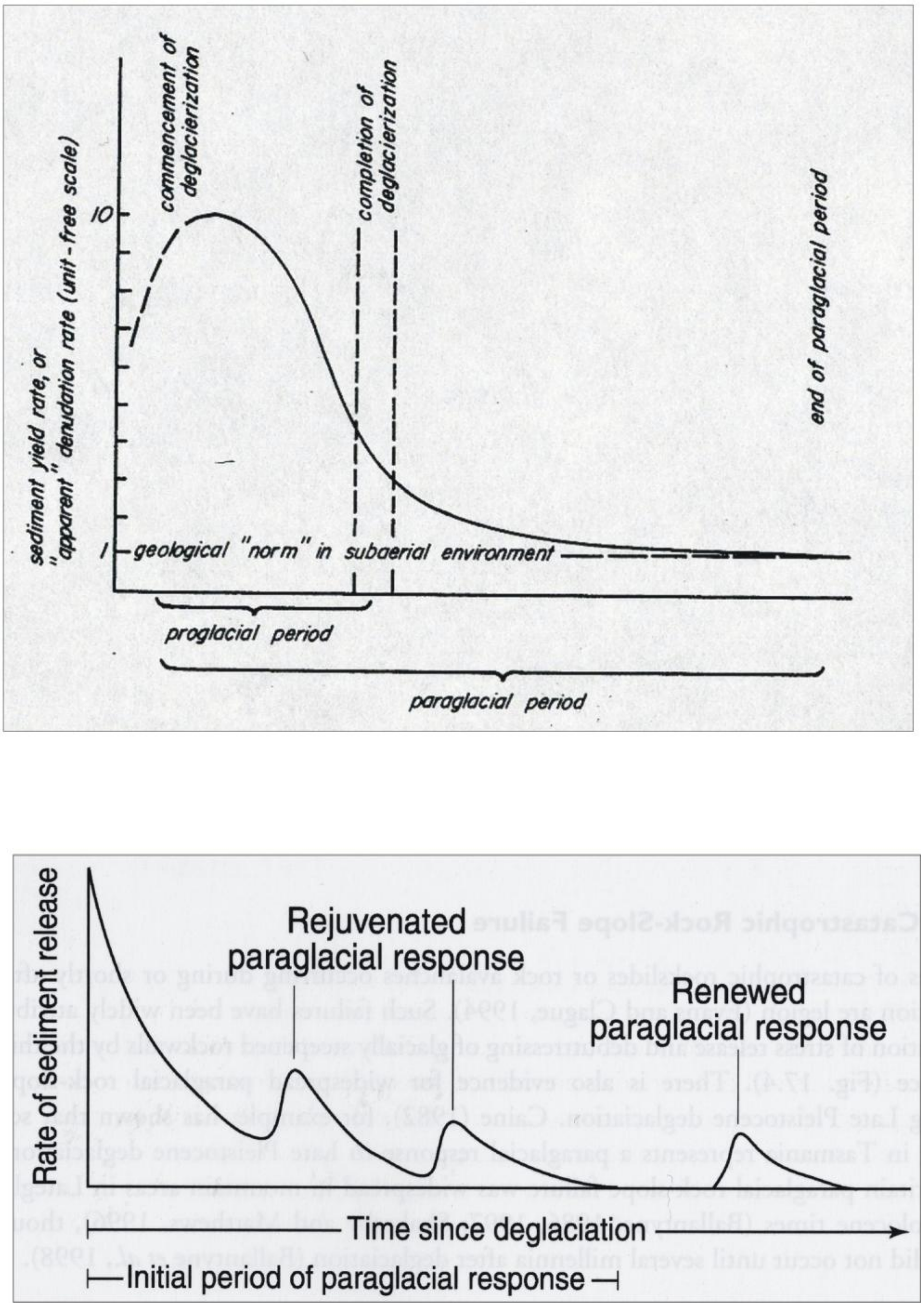
During deglaciation, proglacial conditions existed in drainage basins and rivers were affected by high discharge, due to a run-off regime driven by snow/ice melt, and high sediment availability and delivery, due to the combination of an open (sparse vegetation cover) and readily available sediments (Lewin and Macklin, 2003). This resulted in extensive erosion and aggradation by rivers. Generally, it is suggested that during the brief Bølling-Allerød period (15- 12.9ka cal. BP), it is thought slope stability, lower runoff and reduced sediment availability, due to soil development and increased (scrub or tundra) vegetation cover prompted rivers to incise their beds. However, isostatic uplift during the initial paraglacial period probably elevated the deposits high above the valley floor. The return to glacial conditions during the Younger Dryas Stadial (12.9-11.5ka cal. BP) prompted a second phase of valley floor aggradation as periglacial drainage basins responded to flashy seasonal discharges and high sediment availability, due to sparse vegetation cover and periglacial processes (Bryant, 1983). The return to warmer conditions in the early Holocene prompted incision and reworking by the rivers.

The modification of the landscape following glaciation is embodied in Church and Ryder's (1972) paraglacial model (Figure 2.11A), which suggests that in formerly glaciated landscapes 'non-glacial processes are directly conditioned by glaciation' (p.3059). Sediment release during the initial period of paraglacial response, i.e. the 'proglacial period', is high, this is followed by an exponential decline in sediment release until 'equilibrium' is reached, and the 'paraglacial period' is complete. However, the term 'paraglacial' has somewhat lost its original meaning and is now applied to both sediments and landforms as well as processes (Ballantyne, 2003). Increasingly it is recognised that glacigenic sediments contribute to alluviation throughout the Holocene, and are released and reworked by a number of processes not envisaged by Church and Ryder's (1972) model. Ballantyne (2002a,b; 2003) suggests that releases of glacigenic sediment through both rejuvenated and renewed paraglacial response, takes place as and when sources of glacigenic sediments are accessed by the fluvial system (Figure 2.11B). Whilst Macklin *et al.* (1998) suggest that the rates and patterns of Holocene sediment supply, storage and transfer are conditioned by deglaciation, and that the extent and thickness of paraglacial valley infill (derived from glacigenic deposits) and valley morphology (alluvial basins and narrow reaches) established the boundary conditions for subsequent river development they see paraglaciation as complete before the start of the Holocene. Passmore and Macklin (2001) attempted to quantify the amounts of Holocene sediment transferred and stored within a small reach of the South Tyne River. They calculated that ~50% of the glacigenic sediments had been removed by postglacial valley floor reworking; interpreting this to be evidence of paraglaciation. However, it is unclear what phase of paraglaciation this represents. Currently the paraglacial model as applied to alluvial landsystems has not been fully explored. Although it thought the initial paraglaciation period ended before the Holocene began, and thus incision is driven by climate changes (cf. Macklin 1999; Lewin *et al* 2005), it is probable that paraglaciation continued to excerpt its influence on the landsystem resulting in sediment releases throughout the Holocene.

2.3.1.4 Anthropogenic activity

In terms of the mid to late Holocene fluvial sequences, the pattern of cut and fill activity has been complicated by the increased influence of anthropogenic activity in drainage basins. Anthropogenic activities have affected hydrological regimes and sediment supply through land clearance (Neolithic farmers), agriculture (Romans) and industrialisation (19^{th} century lead mining) (Harvey and Renwick 1987; Rumsby 1991).

A.



B.

FIGURE 2.11: A. Schematic diagram of sedimentation pattern during the paraglacial period (taken from Church and Ryder, 1972). B. Effects of extrinsic perturbations on the exhaustion model of paraglacial sediment release (taken from Ballantyne, 2003).

Macklin and Lewin (1993) went as far as to state that the Holocene fluvial record was "climatically driven but culturally blurred" (p. 119). Periods of increased disturbance in the catchment have been reflected in the alluvial record, and river channel change is strongly linked to the combined influence of human activity providing readily available sediments and increasing runoff, and periods of climatic instability (Ballantyne 1991; Tipping 1992; Macklin 1999).

It is clear that the impact of deglaciation upon fluvial systems probably continued to exert its influence throughout the Holocene, but that its affect has been underestimated and remains unqualified (Passmore and Macklin, 2001). Relatively little is known about the nature, magnitude and rate of sedimentological and morphological response that occurred during the most dynamic period (i.e. millennial and sub-millennial scale shifts) of recent climatic (cf. Dansgaard *et al*. 1993; Bard *et al*. 1996) and landscape change following the last glacial maximum. In part this may be related to the dynamic nature of the rivers at that time, which were probably experiencing rapid lateral and vertical change and continually reworking their valley floor deposits (Lewin and Macklin, 2003), leading to low preservation potential. It can be hypothesised that the sequences would comprise outwash terraces and paraglacial river terraces that developed as a consequence of both climatically driven change and glacio-isostatic uplift. Fluvial response during the Holocene has been correlated with climate (Macklin and Lewin, 1989) and anthropogenic activity (cf. Bell 1982; Bell and Walker 1992); however, in the context of this study it is important to consider whether the pattern of fluvial response during the Lateglacial and early Holocene is conditioned by the boundary conditions set following deglaciation (i.e. coarse sediment loads and valley morphology).

2.4.1 Fluvial response across Britain

British river catchments can be divided into those which lay wholly or partly within the limit of the last, Devensian (Dimlington Stadial) ice sheet, and those that lay outside it and were not directly affected by glaciation but were subject to cold climate (periglacial) conditions. As a consequence, fluvial response in Britain is spatially variable, with some catchments having terrace sequences underlain by proglacial outwash, others that simply aggraded under a harsh periglacial climate, and those which responded to paraglacial reworking following deglaciation.

Within lowland river valleys of southern Britain, records of fluvial response during the last glacial period suggest sequences characterised by high-energy erosion and transport in braided rivers depositing gravel under a periglacial regime (cf. Gibbard 1988; Gao *et al*. 2000). Following climatic amelioration, catchments were characterised by relatively stable conditions i.e. little channel migration and low sediment transport during the early and mid Bølling-Allerød Interstadial. In contrast, however, towards the end of the Interstadial and throughout the Younger Dryas, an unstable system existed, characterised by lateral channel instability, high bedloads and braiding (Rose *et al*. 1980; Collins *et al*. 1996).

Those rivers that lay immediately beyond the limits of the ice sheet, in ice-proximal locations, responded to proglacial fluvial and periglacial regimes. Dawson (1985) and Dawson and Gardiner (1987) recorded terraces and their associated lithostratigraphy in the lower River Severn Valley. They suggested that sediments were aggraded in a proglacial environment and subsequently incised following the onset of deglaciation and the adjustment to improving climatic conditions, with reduced runoff and sediment supply from upstream. The terraces were interpreted as representing proglacial outwash terraces rather than river terraces, and were aggraded throughout the Dimlington Stadial (26–12.9ka cal. BP).

In upland river valleys that lay within the limits of the Devensian ice sheet, rivers re-established during and immediately following deglaciation. Vandenberghe (1995) suggests that at the scale of one cold-warm cycle, fluvial response occurs at climatic transitions when vegetation, soil and runoff are adjusting to climate change. Therefore, fluvial response could be expected at the start of the Bølling-Allerød Interstadial, during the return to severe cold conditions associated with the Younger Dryas Stadial and during the early Holocene as hillslopes stabilised as the climate ameliorated and vegetation re-colonised catchments. However, published studies of fluvial response during the Lateglacial climatic transition are rare.

Deglacial/Lateglacial fluvial activity during this period is likely to have being extremely dynamic. As the ice decayed, large volumes of coarse and fine grained sediment were fed into proglacial river systems, aggraded as outwash and infilled the valleys. Following the disappearance of the ice, rivers adjusted to initially high sediment loads derived from glacial outwash, and later from bare hillslopes affected by periglaciation and paraglacial reworking. Subsequent incision of the glacigenic sediments was in response to waning sediment supply and glacio-isostatic rebound (Macklin and Lewin 1993; Macklin 1999). Within UK upland river basins, late Devensian glacigenic erosional and depositional landforms have been subjected to limited postglacial reworking, and remained largely unaltered. Thus, their valleys contain moraines, eskers, tributary (paraglacial) fans etc., which have resulted in the restriction of some valley floor reaches (Macklin, 1999). It is the wider/open valley floors, which have provided the accommodation space that has enabled the preservation of Holocene alluvial units. Over the last few decades there has been considerable interest in elucidating Holocene fluvial histories in upland Britain (Harvey *et al*. 1981; Harvey and Renwick 1987; Macklin and Lewin 1989; Macklin *et al*. 1992b; Passmore and Macklin 1994; 1997; 2000; Tipping 1995; 1998; Howard *et al*. 1999b; 2000).

Although evidence for the Lateglacial is scarce, an outline of broad stages of Lateglacial terrace development can be established from work carried out on Holocene sequences within and beyond the ice margin. Evidence from sequences within upland UK sites generally illustrate that higher terrace(s) crop out above the Holocene cut and fill units. In all cases they are conventionally assumed to be Lateglacial in age, but little else is understood about the development or sedimentology of these terraces. Unlike the Holocene fluvial record, comparison and correlation between sites is difficult because the evidence for Lateglacial fluvial activity is more fragmentary, complicated by local ice histories and possible effects of glacio-isostatic rebound. As Howard *et al*. (2000, p.39) point out "The paucity of Lateglacial/early Holocene records may reflect the paraglacial development (sensu Church and Ryder 1972) of many of these river valleys leading to increase energy levels through the Holocene and erosion and reworking of sediments relating to earlier parts of this period". Despite the lack of evidence and/or research into Late Devensian fluvial response, it is possible to highlight a number of characteristics common to all the upper Lateglacial terraces recorded in the literature.

In upland areas across northern and western Britain, the upper terraces appear consistent in height above present river level, and it is assumed they are similar in age as well. In terms of northeast England, terrace sequences have been recorded in a number of east draining river valleys (e.g. Till, Aln, Coquet, Font, Wansbeck, Blyth and

Wear; Figure 2.12), and assumed to be Late Devensian in age (cf. Land et al. *1974; Frost and Holliday 1980; Lawrence and Jackson 1986; Smith 1994; Young* et al. *2002). Terrace* s*urfaces along the Rivers Blyth, Wansbeck, Pont and Font lie at 17, 10 and 3m above* present river level (Young et al., 2002). The lowest terrace is interpreted as a Holocene cut and fill unit. However, the higher terraces remain undated, and it is unknown whether glacigenic or fluvial sediments underlie the terraces.

Within the Tyne Basin, Macklin and Aspinall (1986) identified a higher terrace on the River West Allan, a tributary of the South Tyne. Its surface lies approximately 20m above present river level, and is underlain by boulder gravel supported in a matrix overlying a diamicton. Whilst no dating has been undertaken, it was assumed to be Lateglacial in age. Aspinall *et al*. (1986) acknowledge that some of the best preserved flights of terraces in the Tyne basin crop out on the valley side of the upper South Tyne Valley at Garrigill. They identified six unpaired terraces, with surfaces lying between 10-2m above present river level. The highest terrace, which is traceable ~7km downstream to Alston (cf. Macklin, 1997), comprises cobble and boulder gravel, interpreted as incised proglacial outwash. The lower terraces are interpreted as Holocene alluvial units, typically comprising cobble gravels overlain by silty sands aggraded as channel bed and overbank deposits. Reaches of the South Tyne and its tributaries have been incising through bedrock during the Holocene, a possible indication that climate change is not the primary driver of incision but glacio-isostatic uplift has played a significant role. Terraces also crop out along the North Tyne Valley, though these are lower in height than the ones recorded in the South Tyne Valley. Moores *et al.* (1999) identified an upper terrace, its surface lying between 8 and 7m above present river level, which comprises fine sand overlying sandy coarse gravel. Two upper terraces were identified in the River Rede, a tributary of the River

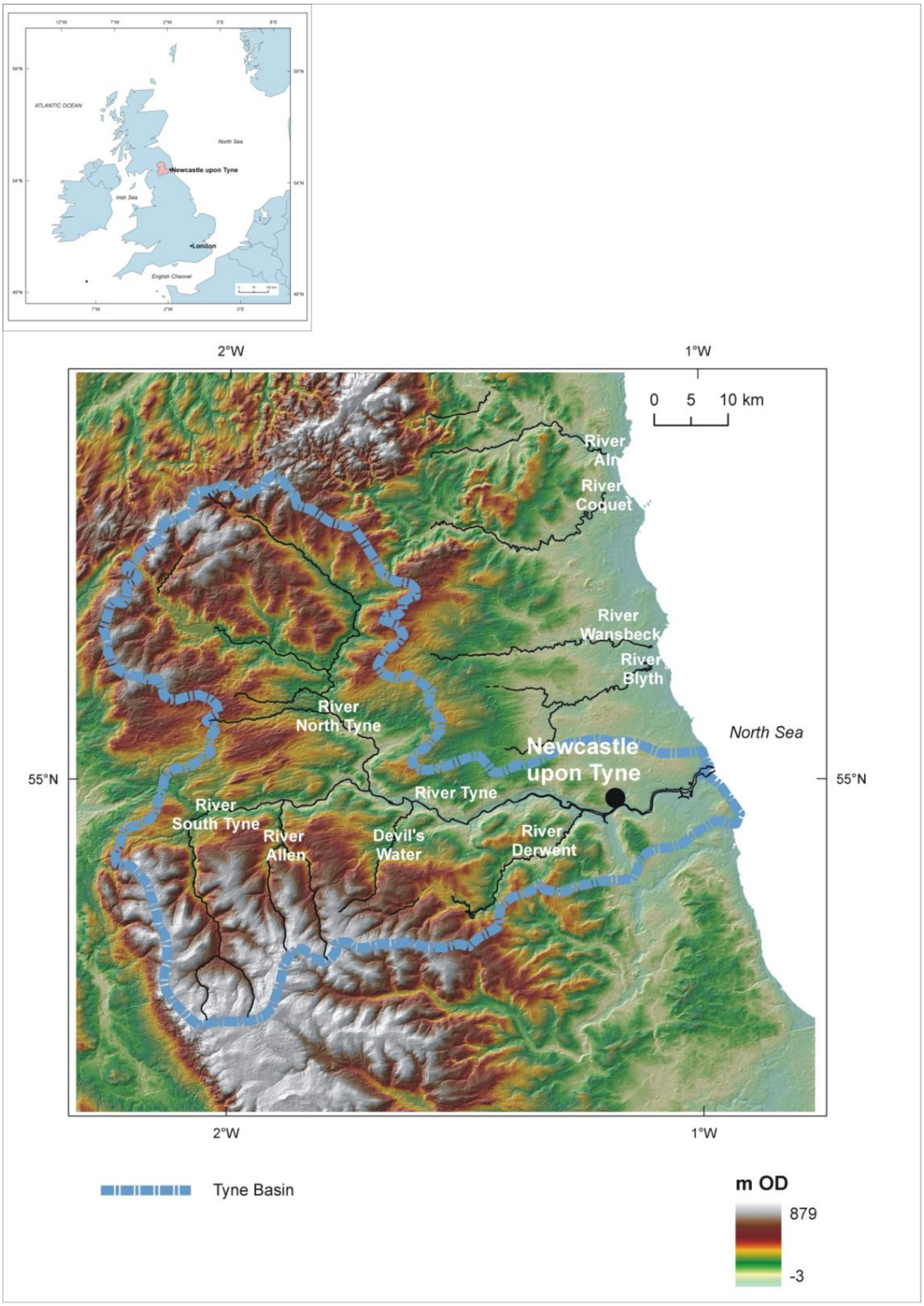


FIGURE 2.12: Elevation map (source: www.landmap.ac.uk) showing the east draining rivers in Northumberland and the main tributaries of the River Tyne.

North Tyne, by Moores *et al*. (1999), with surfaces between 6 and 5m above present river level. In the Eden basin, adjacent to the Tyne, Cotton *et al*. (1999) identified an upper terrace on the River Irthing, with a surface at ~6m above present river level. Due to its elevated position relative to dated Holocene units and to the present river level, the upper terrace is interpreted as Lateglacial reworking of the valley floor glacigenic infill. At all the sites, there is no dating control on terrace development.

From the above discussion, it is emerging that Lateglacial terraces within the Tyne Basin and adjacent catchments range in height between 20 and 6m above the present river level. The highest terraces (20-10m above present river level) appear to have developed as the rivers trenched through glacigenic sediments, probably at the close of or immediately following deglaciation. The lower terraces represent reworking of glacigenic material (a paraglacial effect) possibly during the Younger Dryas Stadial or early Holocene period. Given the height of the upper terrace above present river level, it is clear that a significant amount of incision has taken place since deglaciation. When looking at the pattern of Holocene alluvial units in UK upland catchments, previous studies of have demonstrated that a region-wide response to environmental change can be synthesised (cf. Macklin and Lewin 1993; Macklin 1999). Therefore, it is interesting to observe whether earlier episodes of incision and terrace aggradation in the Lateglacial record are broadly synchronous or if responses are conditioned by local complexity. Table 2.2 is a broad overview of the upper terraces recorded in studies of a number of small and large catchments across upland Britain. These upper terraces have been attributed to reworking of glacigenic sediments during the Late Devensian/early Holocene. Taylor *et al*. (2000) observe that the highest terraces (~6m above present river level) that crop out along the River Swale Valley, Yorkshire, are underlain by glacigenic (i.e. proglacial outwash or glaciolacustrine) sediments, illustrating that

Table 2.2: Summary table of key characteristics of upper terraces recorded in UK (England, Wales and lowland Scotland) catchments that have been assigned to the Late Devensian. All dates are relative except for the date from Kirtle Burn which is a calibrated ^{14}C age derived from an organic deposit and those from the Wharf which are U-Series derived ages from calcium carbonate coatings. NG denotes data not given in original paper. HAF denotes Holocene alluvial fill.

Reference	**River/Catchment**	**Zone**	**Catchment Size (km^2)**	**Height above present river level (m)**	**Sedimentology**	**Relative age**
Howard *et al.* 1999	Wharfe, Vale of York	Upland	244	20-10	Massive, poorly sorted gravel	~7.4-5.1ka BP
				7-5	Massive, poorly sorted gravel	~>8-4ka BP
Taylor and Macklin 1997	Swale, Vale of York Swale	Piedmont	499	10	Coarse gravel	~18-14ka BP
Taylor *et al.* 2000		Piedmont, lowland, perimarine		6	Glacigenic	Late Devensian/Early Holocene
Hooke *et al.* 1990	Dane	Piedmont	NG	15-9	Gravel	Late Devensian/Early Holocene
				6	Sandy Gravel	Early Holocene
Harvey and Renwick 1987	Langdon, Hodder	Upland	NG	8	Gravel overlying soliflucted till	Late Devensian/Early Holocene
				6-4	Gravel overlying till	Late Devensian/Early Holocene
Macklin and Lewin 1986	Rheidol	Lowland, perimarine	182	10	Gravel and glaciofluvial gravels	Late Devensian/Early Holocene
Taylor and Lewin 1996; 1997	Severn	Upland	876	7-6	Outwash gravels	Late Devensian/Early Holocene
	Afon Tanat, Severn	Upland	788	10	Outwash gravels	~18ka BP
				5	Cobble gravel, silty sands	10.9ka BP
	Afon Vyrnwy, Severn			10	Outwash gravels	~18ka BP
				7	Cobble gravel, silty sands	12.4ka BP
				5	Cobble gravel, silty sands	10.2ka BP
Tipping 1995	Kirk Burn, Kirtle Water	Piedmont	110	11	Sands over till	~13ka BP
				7	Coarse gravel	9.1ka cal. BP
Tipping *et al.* 2007	Kelvin Valley	Piedmont	NG	HAF	Gravel, sands and peat horizons	11.1.-9.4k cal. yr BP
Tipping *et al.* 1999	Upper Annandale	Upland	NG	HAF	Minerogenic	8.3-7.9k cal. yr BP
Smith and Boardman 1994	Mosedale	Upland	NG	7-4	Outwash sand and gravels	Lateglacial

trenching of the valley floor infill probably followed deglaciation but may have been penecontemporaneous. The modern River Swale is at least 17m below the till deposits, demonstrating a significant incision trend since deglaciation. It would appear that the size of catchment is not related to the development of an upper terrace or to the height at which its surface lies relative to present river level, as this appears to be fairly consistent at ~10m. However, the location within the catchment, and probably local deglacial history, exerts an influence. Upland zone upper terraces are lower in height above the present river level than those in the piedmont zone, although this could be related to increased incision in the piedmont zone, differential uplift or bedrock reaches, which would act to reduce incision (Brown 1991; 1995; Taylor *et al*. 2000).

Evidence for paraglacial (*sensu* Ballantyne, 2002; 2003) incision and dissection of glacigenic sediments, leading to the development of the upper terraces, can be seen in sequences investigated in Wales and Scotland. Thomas *et al.* (1982) investigated the Dyfi Valley, where up to five terraces were observed. The highest terrace surfaces lies at 30m above present river level and comprises glacigenic sediments, with the lower terraces comprise a veneer of fluvial gravels overlying the glacigenic sediments. The sequence confirms a high incision rate by the River Dyfi since deglaciation, essentially incising through the glacigenic sediment to create the upper terrace, with reworking and subsequent aggradation during the Lateglacial or early Holocene period to form the lower terraces. Macklin and Lewin (1986) identified a higher terrace above the Holocene alluvial units at Capel Bangor, River Rheidol. The highest terrace comprises a sequence of gravel, divided into two lithofacies. The lower lithofacies comprises ice-marginal glaciofluvial gravels, unconformably overlain by the upper lithofacies comprising low-sinuosity channel gravels. The sequence illustrates fluvial down-cutting into proglacial outwash following deglaciation and reworking/refilling of the valley floor during the climatic deterioration of the Younger Dryas Stadial. However, there has been no direct age determination of the deposits. In the Upper Severn basin, the highest terraced landforms at ~10m above present river level are underlain by proglacial outwash sediments (Taylor and Lewin, 1996; 1997).

In contrast to findings from elsewhere in upland Britain (cf. Macklin and Lewin 2003; Johnstone *et al*. 2006), Tipping *et al*. (2007) suggests fluvial activity in the early Holocene was more significant than during the mid to late Holocene. Tipping *et al*. (1999; 2007) reported Lateglacial incision of up to 15m in Upper Annandale (lowlands) and the Kelvin Valley (Midlands) probably reflecting paraglacial reworking. In Upper Annandale, terrace and alluvial fan formation provides evidence for activity during the Younger Dryas period, although there is no absolute dating. However, the earliest evidence of valley floor activity i.e. channels and peat beds was dated to between 8.2 and 7.9ka cal. BP in Upper Annadale and to between 11.1 and 9.6ka cal. BP in the Kelvin Valley. Given the amount of incision in these valleys, there is strong argument for a response to the combined influence of isostatic uplift, baselevel (eustatic) and climate driven change in all of these catchments. It can be suggested that paraglacial alluvial response (cf. Ballantyne, 2003) is demonstrable through incision of the glacigenic valley infill, and the development of terrace and fan features.

Where Lateglacial terrace sequences in England, Wales and lowland Scotland comprise one or two terraces, sequences in upland Scotland are characterised by up to five terraces (dependent upon accommodation space) from that period. The sequences comprise proglacial glaciofluvial terraces, and reflect paraglaciation and reworking; they are not simply alluvial terraces. The summary Table 2.3 shows that terrace surfaces lie up to 40m above present river level and are underlain by glacigenic

TABLE 2.3: Summary table of the key characteristics of the dissected terrace landforms that crop out in the upland valleys of Scotland. Dates are all relative. NG denotes data not given in the original paper.

REFERENCE	RIVER/CATCHMENT	ZONE	CATCHMENT SIZE (KM^2)	NUMBER OF TERRACES	HEIGHT ABOVE PRESENT RIVER LEVEL (M)	SEDIMENTOLOGY	RELATIVE AGE
Young 1976	Feshie	Upland	240	5	40	Proglacial outwash	Late Devensian
Robertson-Rintoul 1986	Feshie	Upland		2			~13-10ka BP
Maizels 1983	North Esk	Upland	NG	5	25-20	Cobble gravels	Late Devensian
Sissons 1979	Glen Roy and Glen Spean, Great Glen	Upland	NG	>20	40-30	Glaciofluvial	Late Devensian
Sissons and Cornish 1983	Glen Roy, Great Glen	Upland	NG	NG	10	Cobble gravels	Late Devensian

sediments. Maizels and Aitken (1991) demonstrate that incision of outwash deposits and terrace formation was driven by glacio-isostatic rebound during and following deglaciation. They linked terrace sequences to three main phases of geomorphic activity following the onset of deglaciation: (1) initial aggradation of outwash across the valley floor and reworking by braided meltwater channels between 14.5-13k yrs BP; (2) increased stability during 13-11k yrs BP; and (3) upland aggradation and terrace formation, and downstream incision between 11-10k yrs BP. Terraces are erosional rather than depositional alluvial sequences. Thus, sequences in upland Scotland are not related to the development of the postglacial fluvial system but rather reflect both proglacial and paraglacial development (cf. Young 1976; Maizels 1982; Sissons and Cornish 1983; Robertson-Rintoul 1986; Maizels and Aitken 1991). Glacio-isostatic uplift was significant in this region because the BIIS was thicker here, resulting in greater depression and subsequent rebound as the ice disappeared, driving proglacial (and paraglacial) rivers to incise through the valley infill.

The widespread occurrence of a higher terrace(s) appears to be ubiquitous in northern Britain, demonstrating good synchrony although caution must be urged when applying this broad-scale approach. Formation of these terraces would appear to be intrinsically linked to episodes of major incision. It is clear from the review that these upper terraces are essentially undated. Most of the terraces have been assigned relative ages based upon surface altitudes and morphology but few absolute ages have been established for any of the older terraced landforms in upland Britain. Where ages do exist, they have been derived from either radiocarbon or U-Series dating. For example, Howard *et al*. (1999b) dated calcareous cement from gravels in the upper terraces (T1, T2) at Wharfdale, Yorkshire, using U-series. The U-series ages gave a minimum date for terrace construction, dating the formation of the cement rather than the aggradation of the gravel, hence the gravels were probably deposited much earlier. Tipping (1995) estimated the age of the upper terrace in Kirtle Water, Scottish lowlands, to be post 13k yr BP based on the radiocarbon age (9.1ka±45 cal. BP) of the lower terrace. Taylor and Lewin (1997) combined ages from radiocarbon dated Holocene alluvial units with amounts of incision between terraces in a novel approach to calculate the ages of the upper terraces in the Severn Basin. They estimated the ages of the upper terraces on the Afon Vyrnwy (T1, ~7m; T2, ~5m) to be 12.4ka BP and 10.2ka BP respectively, and on the Afon Tanat (T1, ~5m) to be 10.9ka BP. These dates suggest development during the Younger Dryas Stadial. Robertson-Rintoul (1986) developed a relative terrace chronology for the Feshie, Scottish highlands, based upon the chronosequences of the soils on the terrace surface. However, the upper terraces were still only assigned relative dates of 13 and 10ka BP based on elevation above present river level. Other than these few attempts, it is clear to see that the upper terraces have not been dated. This problem is of course related to the issue that conventional dating using radiocarbon requires organics and the sediments underlying the upper terraces are mostly comprised of coarse boulder gravel, therefore, attempts in the past to date these types of sediments using conventional radiometric techniques would have been futile.

However, in the last decade there have been advances in both geochemical and other radiometric dating techniques (see section 2.5 for review), which increases the possibility of dating such sequences.

Although the dates may be relative, the difference in height between the upper and lower terraces suggests a rapid rate of incision and refilling between the end of deglaciation and the onset of the Holocene. In the Kirtle Water and the Upper Severn (Tipping 1995; Taylor and Lewin 1996, 1997), the lower of the two upper terraces are laterally less extensive throughout the basin, which suggests reworking following development of the second terrace was greater and more efficient at removing the sediments out of the system. This is further evidenced in the Dane Valley, where almost 6m of incision took place through middle terrace but less than 2m of sediment were aggraded in the lower terrace unit (Hooke *et al*., 1990). The development of the middle terrace appears to coincide with a period of extensive incision, reworking and erosion.

In those catchments that lay beyond the Devensian ice margin, the pattern of valley floor development is very different. Research in the Trent Basin indicates that the Lateglacial period is characterised by accumulations of sand and gravel (likely to date to the Loch Lomond Statial), and alluvium deposited during the Holocene. The sequence in the Trent and its tributaries is characterised by Holocene valley floor reworking rather than incision and terrace formation seen within the valleys that were formerly glaciated (Maddy 1999; Howard *et al*. 1999, 2007; Brown *et al*. 2001).

In summary, research during the past few decades has been solely focused on elucidating Holocene fluvial histories in upland Britain although there is increasing interest in Late Devensian river response (Lewin and Macklin, 2003). However, currently what is lacking in the research is the detailed investigation of the high terrace(s) that lie above the modern valley floor cut and fill units, how these relate to development of the river valley following deglaciation and what they can tell us about changing environmental conditions in the immediate postglacial period. This previous lack of interest is likely to relate both to the fragmentary preservation of the upper terraces and the issues of dating control, and possibly coupled with previous research paradigms that were more focused on small-scale Holocene climate change and alluvial development, rather than the current focus which has shifted onto larger temporal and spatial scales of investigation.

From this review of the available literature, the relative elevation of the higher terraces above present river level in formerly glaciated catchments has led to them being assigned to the Late Devensian. However, in many cases there has been no attempt to interpret or classify the terraces into stages, to date them or even to understand their mode of formation. Beyond the ice margin, river valley development is characterised by lateral reworking of the Holocene floodplain without the development of terraces. In upland glaciated basins the surfaces of Lateglacial terraces lie at between 40 and 6m above present river levels and predominantly comprise coarse boulder gravel. Given the more detailed sedimentological investigation of the terraced landforms in Scotland and Wales, it would seem that the sediments aggraded under a glacial regime but the landforms developed subsequently through erosion, incision and reworking. Although some lower elevation terraces contain glacigenic material, it is probable that they represent early Holocene (paraglacial) reworking. It is unclear, due to the lack of dating control, when the upper terraces were formed but it is most likely prior to the start of the Holocene (11.5ka cal. BP). The mechanisms driving change in the system are glacioisostatic uplift, baselevel change, local hydrological change, sediment availability and climate change. However, it is unclear how the terrace sequences in upland England relate to those factors and, therefore, a Lateglacial story remains to be elucidated.

2.5 Dating methods and chronological frameworks

The above discussion has demonstrated a critical lack of dating control on deglacial and Lateglacial landform-sediment assemblages. However, the need to constrain sedimentary sequences through age estimation is important if terrestrial records are to be correlated with high-resolution marine records of climate change (Lewis *et al*. 2001; Bridgland *et al.* 2004) and to enable them to be compared with other sequences both within the UK and region wide (i.e. North Atlantic).

Techniques used to date/determine the age of sequences in the Quaternary can be divided into radiometric methods, i.e. those based upon radioactive decay, and incremental methods (not discussed here) which are based on the measurement of regular accumulations of sediments or biological materials over time (Lowe and Walker, 1998). Age-equivalent markers can also be employed to date a sequence. By establishing an age marker at one locality (by radiometric or incremental techniques) then, by inference, equivalent horizons at other sites/within other successions can be indirectly dated. Finally, by establishing an order of antiquity from landforms or sediment units a relative sequence of events can be determined. Age determination is not without problems; the methods are all subject to unique problems and uncertainties. Therefore, even when an age has been determined, it is not necessarily reliable. The degree of reliability of the age estimates that are derived is related to precision (i.e. statistical uncertainty attached to the physical or chemical measurement) and accuracy (i.e. the degree of agreement between the true age and that obtained by the dating method) (Lowe and Walker, 1998).

Despite the availability of a number of dating methods, the main problem associated with Late Pleistocene sediment sequences is the limited availability of datable material. Sequences that aggraded during the transition between the end of the last glacial stage and the start of full interglacial conditions are not rich in floral or faunal remains, and consequently it is this absence of the conventionally datable material that has hindered the development of chronologies. However, in the last decade two radiometric techniques have increased the possibilities for dating inorganic (minerogenic) sediments/surfaces, and developing relative chronologies. First, Optically Stimulated Luminescence (OSL) dating, which targets the minerogenic (quartz or feldspar) grains, provides opportunities to date inorganic sequences especially with the advancement in luminescence techniques in the last few years (Duller, 2004; 2006). OSL can constrain the ages of the sequences; and the advantage of this technique is that it gives a primary age representative of the sediments that underlie the terrace as opposed to dates from secondary sources (i.e. organics) within the sediment (see chapter 4 for a more detailed discussion on OSL). Secondly, the potential to date gravels, boulders and rock surfaces using cosmogenic dating techniques (^{10}Be and ^{36}Cl) has been successfully demonstrated in glacial and fluvial contexts (cf. Repka *et al*. 1997; Hancock *et al.* 1999; Schildgen *et al.* 2002; Ward *et al*. 2007). This technique has not been widely applied to UK sequences as yet. Bowen (1994) and Phillips *et al.* (1996) carried out cosmogenic dating, using ^{36}Cl, on surface boulders from the Ridgacre Formation (Severn Valley), confirming the formation aggraded during the cold climate of MIS 6. More recently, Everest and Kubik (2005) applied cosmogenic nuclide analysis, using ^{10}Be, to date a Lateglacial standstill in the eastern highlands of Scotland.

Chronological frameworks can be developed using litho- and bio-stratigraphic markers for longer-term (i.e. glacial-interglacial) sequences (cf. Gibbard 1977, 1979; Bridgland 1983, 1994; Maddy 2002; Bridgland and Schreve 2004) but have little value in a study for this period.

In terms of alluvial sequences, where organic material has been preserved, ages can be derived through a variety of radiometric dating techniques, such as radiocarbon dating (^{14}C) and uranium-series dating (U-series). The development of chronologies for Holocene alluvial units has been in some ways easier because of the potential for organic material to accumulate within channels or within the sequence that can be dated by ^{14}C (Passmore *et al.*1992; Tipping 1994, 1998; Moores *et al.* 1999; Cotton *et al.* 1999; Taylor *et al.* 2000). The reliability of radiocarbon dating is affected by a number of factors, such as temporal variation in ^{14}C production, contamination and isotopic fractionation, all which must be compensated for where they are known. However, ^{14}C dates derived from palaeochannel organics may only provide a minimum age for terrace abandonment (dependant on where the palaeochannel occurs within the sequence); it is unclear whether the date reflects abandonment of the channel or complete abandonment and incision of the floodplain. Macklin and Lewin (1993) and Johnstone *et al.* (2006) both highlight the lack of reliably dated Holocene sequences, due to poor preservation of organics (e.g. oxidation issues) and channel reworking (Lang and Nolte, 1999).

An alternative radiometric technique for deriving ages for alluvial deposits is U-series dating. This technique can only be applied where sediments aggrade in areas underlain by calcareous rocks. Therefore, there is potential to date deposits by this method in many areas of the UK. Howard *et al.* (1999b) sampled cemented gravels and tufa deposits in the highest two terraces in upper Wharfdale, Yorkshire, conventionally assumed to be Late Pleistocene. The U-series dates provide minimum ages for formation of the calcareous cement and by inference sediment aggradation but cannot provide absolute dates for terrace formation. Given the nature of the precipitation process and formation of carbonate coatings on gravels, there are issues related to contamination of the tufa deposits and, therefore, the reliability of the derived age is in doubt if this is not compensated for. Murton *et al.* (2001) encountered this problem when dating tufa clasts within channel deposits at Marsworth, Buckinghamshire, known to be late Middle Pleistocene in age. The U-series dates were in discrepancy with amino-acid dates from mollusca taken from the same sequence. This also reinforces the point that, wherever possible, more than one dating technique should be applied to the sequences to establish the reliability or otherwise of the derived dates.

Finally, age-equivalence methods can involve pedogenesis, i.e. the degree of pedogenic development, as a basis for relative chronologies and a means of terrace surface correlation (cf. Harvey *et al.* 1981; Robertson-Rintoul 1986). The technique is based upon the assumption that all soil forming factors remain constant over time, although local topography, vegetation cover and climate are more important (Lowe and Walker, 1998). Taylor and Lewin (1997) used empirical calculations to estimate terrace ages by calculating incision rates between terraces, and estimating time elapsed since the LGM (18ka BP) and ^{14}C dated palaeochannels on younger terraces. Sedimentary facies analysis can be used to indicate climate change and infer a relative age (cf. Thomas *et al.*, 1982; Zieliński and Goździk 2001). Where there are no conventional dating methods possible, establishing a relative order of antiquity as a simple proxy for age chronology on the basis of terrace elevation and stratigraphic position within an overall sequence (i.e. law of superstition) allows the development of a weak chronology that at least envelopes the depositional sequence (cf. Sissons, 1976).

An important aim for this and future research is the development of a chronological framework for the sequences aggraded under Lateglacial conditions, using a reliable dating method. Whilst a number of techniques are available, most are not applicable to the sediments under investigation here. However, recent developments in dating techniques (e.g. OSL) offer the potential to date older sequences, allowing them to be placed within a robust chronological framework. As Lewis *et al*. (2001) note, this information, combined with evidence of climate change determined from the sedimentary record and any palaeoenvironmental indicators (pollen, plant macrofossils, Coleoptera) can be used to interpret the sequence in the context of rapid climatic changes known to have occurred during the Lateglacial period (cf. Bond and Lotti, 1995).

2.6 Summary

From this review, it is apparent that there is complexity in terms of clearly defining and interpreting landform-sediment assemblages, and there is no simple linear relationship between the morphology and sedimentary facies. Thus returning to the conceptual models proposed in chapter 1 (Figure 1.4), the principal landform-sediment assemblages that are predicted to outcrop along the Tyne Valley are given in Table 2.4. Based on the assessment of the sedimentary facies, it would be expected that the sequences that comprise Conceptual Model 1 contain evidence of proximity to ice in the form of faulting, kettle holes and large-scale disturbance. The sediments are likely to include coarse grained gravels, chaotic or unstratified and intercalated diamicton.

The sequence should comprise glaciofluvial sediments that infilled the degrading ice-cored ridge topography. Sedimentologically, the sediment facies underlying the sandur will comprise glaciofluvial sediments typical of braided rivers; they may show a proximal to distal depositional pattern (dependant on where the exposures occur). Sequences within Conceptual Model 2 will comprise ice-contact (kame) landforms, kettle holes eskers and kame terraces underlain by glaciofluvial and glaciolacustrine sedimentary successions. The river terraces will comprise reworked glacigenic sediments indicative of Lateglacial reworking but the Holocene terraces will comprise typical wandering gravel bed sequences (cf. Deslodges and Church, 1987) with cobble gravels overlain by silty sand sequences.

The existence of an upper terrace(s) in upland UK river valleys has been established in a range of studies, although because little or no detailed work has been carried out, it is still unclear whether they are outwash terraces, strath terraces or river terraces. There have been no attempts to correlate these upper terraces, and except for the work carried out in Scotland, it is unclear what is driving their formation. If terrace formation is a response to paraglaciation, it should be expected that the amount of sediment yield (calculated per terrace unit) will show a peak following deglaciation and thereafter decrease exponentially. Combined with climate and hydrological change, and baselevel adjustment, terraces can provide insights into Lateglacial change but thus far very little is known. Finally, a well-constrained chronology is crucial in order to establish the sequence of events with some precision and to be able to begin to make links between region wide climatic changes and local response.

TABLE 2.4: Summary table of principal landforms and their significance in landscape reconstruction (modified from Bennett and Glasser, 1996).

ESKERS	Morphology: steep-crested sinuous ridges of variable extent and size. Indicative of: the location of discharge routes within the glacier.
KAMES	Morphology: irregular collection of mounds and ridges, often with enclosed kettle holes or depressions. Indicative of: areas of outwash deposition in which melt-out of buried ice occurred after the surface had been abandoned by the meltwater streams.
KAME TERRACES	Morphology: valley-side terraces with outer edges which possess concentration of kettle holes or belts of ice disintergration (i.e. kame and kettle) topography. Indicative of: the position of the ice margin.
SANDURS	Morphology: flat surface of sand and gravel formed by braided river system. Indicative of: retreating ice margin with relatively high meltwater/sediment discharge.
RIVER TERRACES	Morphology: valley-bottom terraces with bluffs and surface palaeochannels. Indicative of: the former level of the river.

Chapter Three

Study Area

3.1 Introduction

This chapter provides an introduction to the physical characteristics of the study area. This is followed by a discussion of the approaches used to select the sites for detailed study and is concluded by a brief overview of the study site characteristics.

The Tyne basin is located in northeast England (54° N, 1° W) and the River Tyne drains the greater part of Northumberland, with an area encompassing approximately 2, 927km^2, and a mean discharge at Bywell (lower Tyne) of 44m^{3}s^{-1}. It ranges in elevation from sea level to headwater peaks of 893m OD in the South Tyne. The River Tyne is fed by two major arms: the River South Tyne and the River North Tyne (Figure 3.1). The catchment area of the South Tyne comprises the northwest part of the North Pennines, draining 800km^2. The South Tyne flows northwards from the Pennine massif, towards Haltwhistle, where it turns eastwards towards the coast. The headwater streams (e.g. Thinhope Burn) are characterised by bedrock reaches and steep gradients, <0.01- ~0.1m m^{-1} (Macklin *et al*., 1992b), decreasing to 0.0003m m^{-1} as the South Tyne emerges from the Pennine massif in the middle and lower South Tyne between Lambley and the Warden (Passmore *et al*., 1993). The catchment area of the North Tyne comprises the southern Cheviot Hills and Bewcastle Fells, draining 1118km^2. Rising from a series of small streams, it flows in a southeasterly direction to join with the South Tyne, downstream of Warden, near Hexham (Figure 3.1).

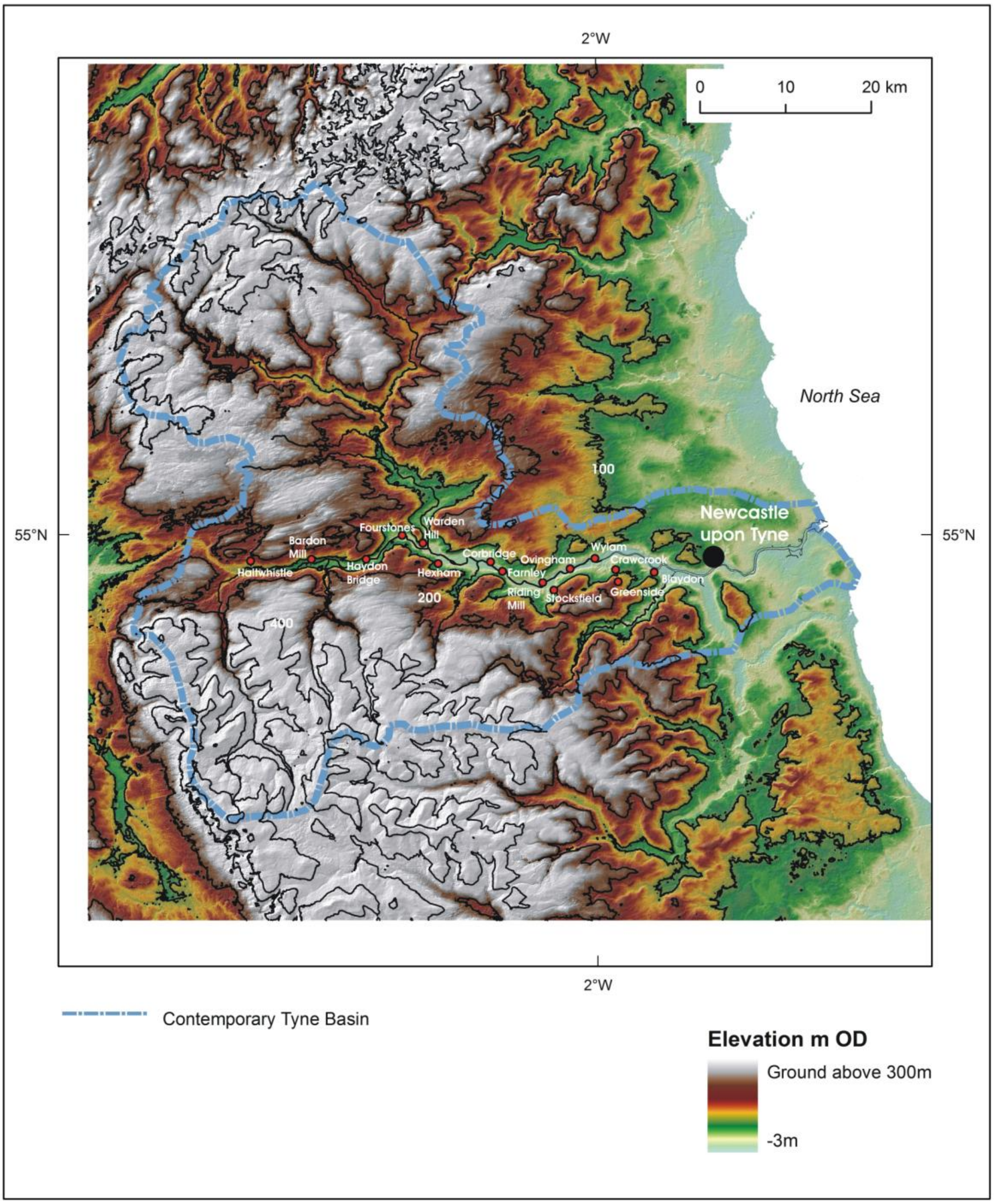


FIGURE 3.1: Elevation map of the Tyne Basin (source: www.landmap.ac.uk). Locations mentioned in the text are shown. Contours are marked in black and heights given on the map.

From here the Tyne flows eastward towards the coast, entering the North Sea at Tynemouth. The gradient downstream of the confluence is ~0.019 m m^{-1} between Warden and Farnley, decreasing to ~0.001 m m^{-1} between Farnley and Wylam. The Tyne Valley is characterised by a thick (>60m), incised valley fill of late Pleistocene sediments. The contemporary regime is characterised by flashy runoff and the planform morphology is similar to a wandering gravel bed river (cf. Church 1983), with meandering reaches that occasionally divide around large gravel islands (Passmore *et al*. 1993). The North Tyne and lower Tyne are currently regulated due to the impoundment of the Kielder Burn in 1981, to form Kielder reservoir. The climate is temperate (cool winters, cool summers), with moderate to high annual precipitation, ranging from ~2000mm in the North Pennines to ~700mm at Newcastle. Land use in the basin is predominantly rural, with significant residential and industrial areas located in the lower reaches of the Tyne basin, where Newcastle upon Tyne is the major city in the region. Forestry is significant in the uplands of the North Tyne Valley, and agriculture (pastoral and arable) takes place in the valley bottoms particularly along the main valley west of Newcastle upon Tyne. Tourism, with Hadrian’s Wall (UNESCO World Heritage Site) running across the basin, also plays a role.

3.2. Geology and topographic setting

The geology of the Tyne basin is well known and is principally dominated by Carboniferous rocks (Figure 3.2). The North Tyne rises on Silurian greywackes and Devonian (old red) sandstones, but for the most part, the river flows over Lower Carboniferous (Namurian) rocks comprising basal marine limestone and cyclic (Yoredale) sequences of fluvial sandstone, siltstone and mudstone (Taylor *et al*., northeast direction in the lower reaches of the North Tyne and South Tyne. The South Tyne rises on Lower Carboniferous (Dinantian) rocks of the North Pennines and flows

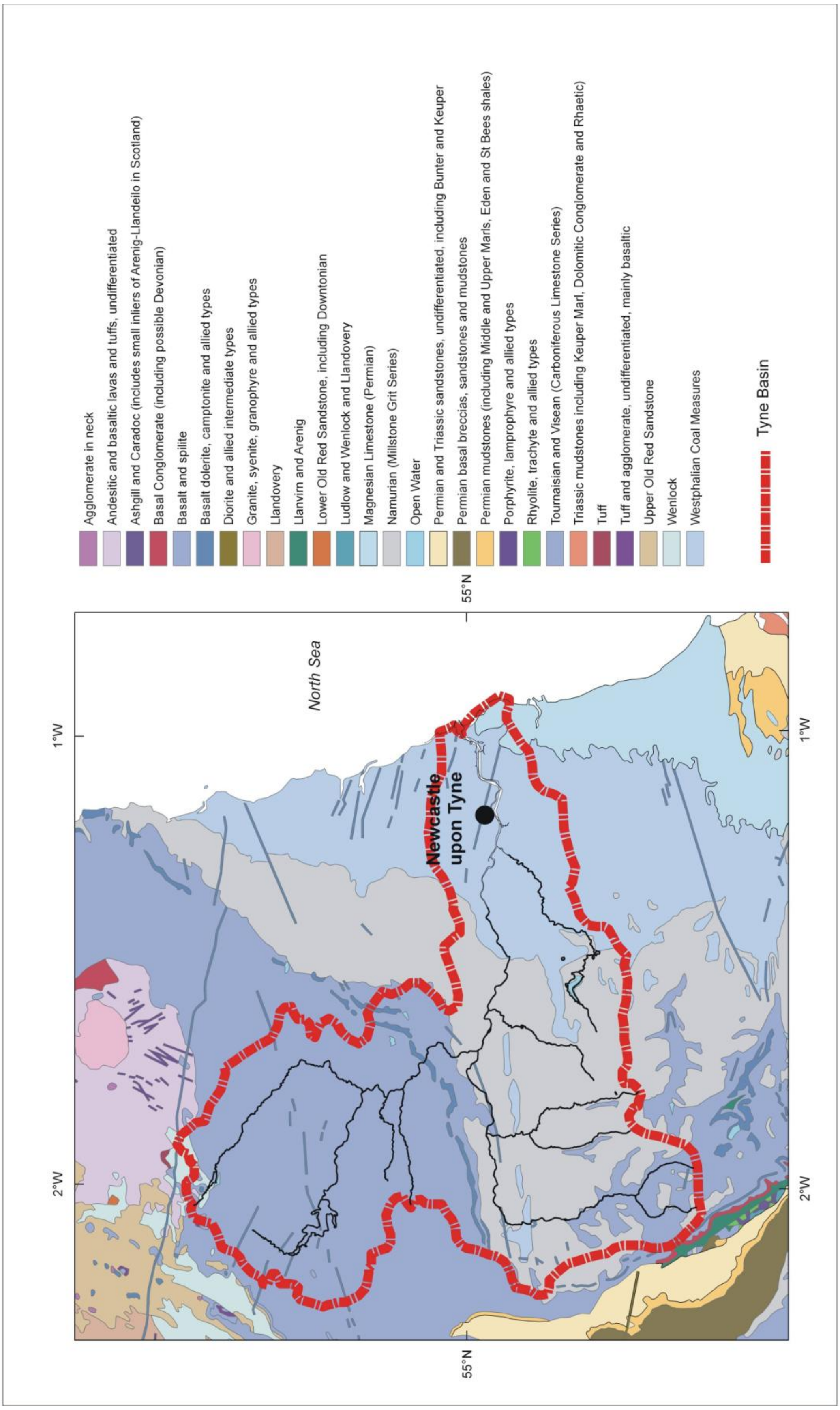


FIGURE 3.2: Solid geology of the Tyne Basin (delimited by thick red outline) and northern England (geological data from Edina Digimap http://www.edina.ac.uk/digimap).

northward across thick sequences of marine limestones and sandstones, then east over Namurian rocks. The North Pennines were affected by mineralisation during the emplacement of the (subsurface) Weardale granite, and mineral veins of lead and zinc have been worked in the South Tyne Valley, where the influence of mining can be seen in the alluvial record (cf. Rumsby, 1991). The lower Tyne flows over Upper Carboniferous rocks (Westphalian) comprising cyclic deltaic sequences of shale/mudstone, sandstone, seatearth and coal, which extend inland from the coast reaching their maximum extent near Stocksfield. Structurally, the Tyne basin is bounded by the high ground of the Alston block to the south, and the Northumberland fault-trough to the north, with the hinge zone of the Stublick/Ninety Fathom faults between (Scrutton 1992; Johnson 1997).

The rockhead channel beneath the Tyne Valley is deeply incised. Beneath the present-day Tyne lies a broad, deep trough approximately 1.6km in width and 50 to 60 m deep, which follows a route eastward before it changes direction from the Tyne into the Team Valley (The Wash, cf. Wood and Boyd 1863) at Dunston (NZ 230 621) (Figure 3.1). Downstream of the Tyne-Team confluence, the rockhead channel forms a narrow gorge (~0.5km) cut through a sandstone ridge that underlies western Newcastle and central Gateshead (NZ 255 627). Below the gorge, the channel widens to ~2-3km in width, with meanders broader than the current river, and flows directly to the present-day coastline where its base is at ~-30m OD. The deeply incised meanders of the buried channel at the Tyne/Team confluence and the lower Tyne above here are infilled with glacigenic sediments up to 60m thick on the southern flanks, suggesting meltwaters and sediments were impounded and accumulated at this point (Cumming 1971, 1977). West of Hexham, the bedrock floor is at present-day sea level, but buried beneath ~30m thick infill. Between Hexham and Ryton (NZ 152 642), the rock head has an uneven profile characterised by deep grooves and a braiding pattern, which could be both related to ice movement and preglacial fluvial development. The present River Tyne still flows along the route of the buried channel, whereas its tributaries were displaced eastward of their buried route, now possessing an asymmetrical profile with their eastern flanks cut into rock and their western side deep in glacigenic sediments (Mills and Holliday, 1998).

3.3 Surficial geology and topography

Northumberland has been glaciated several times during the Quaternary, most recently during the late Devensian, and as such the bedrock is buried beneath till, glaciofluvial sediments and more recent alluvial deposits (Figure 3.3). At the present day, bedrock outcrops in a few locations to the north of the South Tyne Valley between Haltwhistle and Warden. The scarp and dip-slope topography of outcropping Whin Sill and limestone, which has a significant southerly dip, was accentuated by the western ice sheet as it moved eastward coinciding with the strike of the rock (Clark, 1970). In the lower North Tyne Valley near Chollerford, the movement of the ice was across the strike of the rock, which resulted in flattening out of the scarp topography as it infilled the depression with till (Lunn 1980; Scrutton 1992; Johnson 1997). Locally, glacigenic sediments have created distinct undulating topography on the southern flanks of the lower Tyne basin. Terraces have been distinguished above the modern floodplain along the Tyne Valley (Giles 1981; Lovell 1981; Mills and Holliday 1998). The eastern sector of the Tyne basin is characterised by the coastal plain, an essentially flat topography due to the extensive covering of glacigenic sediments. Late Devensian periglacial processes have modified the topography, with the development of smooth, shallow slopes due to solifluction (cf. Douglas and Harrison, 1985; 1987). Occasionally sandstone bedrock highs protrude above the glacigenic covering (Anson

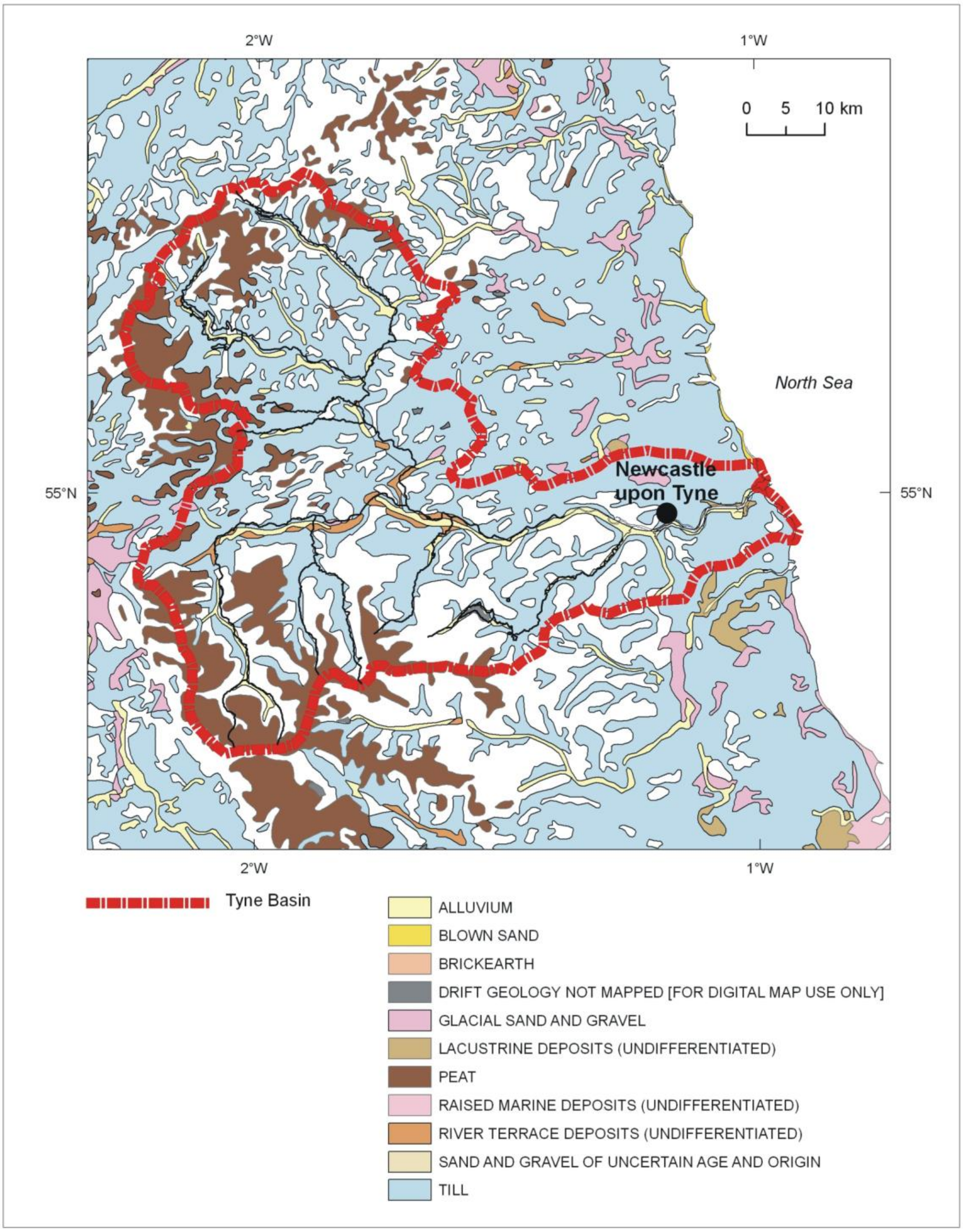


FIGURE 3.3: Surficial geology of the Tyne Basin (delimited by thick red outline) and northern England (geological data from Edina Digimap http://www.edina.ac.uk). White indicates no data.

and Sharp, 1960).

3.4 Site selection

With the objectives clearly laid out in chapter 1, it was crucial that the sites selected were the best available, in that they provided good exposure of the sediments underlying the landforms, and good spatial and temporal coverage. Therefore, the aim of the site selection process was to assess all the available data pertaining to known exposures from mineral reports to maps, and to compile a list of possible sites. Final selection of sites was made after field reconnaissance.

The approach involved a comprehensive desk-based search of the antiquarian literature, the minerals local plan for Northumberland and Industrial Mineral Assessment Unit (IMAU) minerals report and resource map (1:25000), the Soil Survey of Great Britain map for Hexham, Northumberland (1:63360), and historic Ordnance Survey (OS) 1:10560 and contemporary 1:25000 maps to identify exposures (cut-bank or historic sites) and quarry workings (current and abandoned). A complete review of the available late 19th and early 20th century literature and later review papers was made to identify where the historic sections and sites were located. Thus if they could be identified they could be re-visited, and it would be interesting to compare the conclusions of earlier works with the findings of the present research. It is worth recalling that earlier work was achieved on relatively sparse evidence compared to the amount of data currently available. However, this does not mean that the explanations are less valid, or that current interpretations are not also filled with the same speculation and uncertainties that pervaded research 100 or so years ago. A wealth of data exists in antiquarian literature but in order to evaluate the quality, and potential value in such data, sections must be re-assessed.

Several potential sites for re-investigation were identified from the antiquarian literature (e.g. Corbridge Station; The Seal, Hexham; cf. Lebour, 1889), although unfortunately none could be re-visited; due to urban expansion in the Tyne Valley, sites that existed 100 years ago are now subsumed beneath modern developments. A second problem was that in some cases it was impossible to locate the exact site or area because the authors made reference to a general locality rather than a specific location. There was a general lack of detail in the review papers such as Clark (1970). Whilst Clark makes reference to sites of potential interest, only general level maps were provided which delimit the presence of sediments or features by symbols or block highlighting, thus it was difficult to locate possible sites, and UK grid references were not given within the text.

The Northumberland County Council minerals local plan (2000) and the IMAU mineral assessment reports for the country around Blaydon (Giles, 1981) and Hexham (Lovell, 1981) were consulted to identify active and disused quarry workings and the principal deposits worked at those sites. Four quarries were identified, which were working glacigenic and fluvial deposits in the lower/mid Tyne at Greenside, Crawcrook, Stocksfield and Farnley (Figures 3.4;3.5). The most extensive site working glacigenic sediments was located at Greenside, but work had already commenced to landfill the site and thus no working faces existed. The quarry at Stocksfield was disused, work having ceased over a decade ago, and as such the worked faces were degraded. Although it was initially unclear whether the faces were intact or had been backfilled, closer inspection suggested they could provide some useful field evidence as they were unmodified. A field reconnaissance of the area surrounding the quarry provided a second exposure in the sediments underlying some interesting, undulating topography to the east of the quarry. The site at Crawcrook was active and provided access to an extensive area of undulating topography. The site comprised an active excavation, revealing new faces, and an extensive disused excavation with slightly degraded but accessible faces. The quarry at Farnley, where Passmore and Macklin (1994) investigated the Holocene valley floor sequence and had observed some Late Devensian sediments, was still active but current works were concentrated in the youngest alluvium on the valley floor. However, a field reconnaissance of the area to the west of the quarry resulted in the discovery of an extensive cut-bank exposure (*c.* 20m high) revealing a sequence of sediments underlying a terraced landform.

Exposures in landforms are often created through river bank erosion and stream cuttings, and even former railway cuttings (e.g. Wymer, 1968). The Soil Survey of Great Britain Sheet 19 for the Hexham district (Jarvis, 1977) and the IMAU mineral resources maps were used to establish where sediment types were located across the study area. This information was converted to a digital format and integrated with historic and modern OS maps (in digital format) in a GIS, so that possible exposure sites could be identified where river, stream or railway cuttings intersected with glacigenic sediment types. Through this method, a total of 27 sites were identified and visited in the field. At all but one location, dense vegetation cover either prevented access to the possible section or there was no exposure through the sediments. However, the river bank section identified on the River South Tyne at Fourstones revealed an extensive (~1.5km) cut-bank that was well-exposed and accessible. On initial reconnaissance the terraced landforms appeared to be underlain by a sequence of diamicton, glaciolacustrine sediments, and sand and gravel deposits.

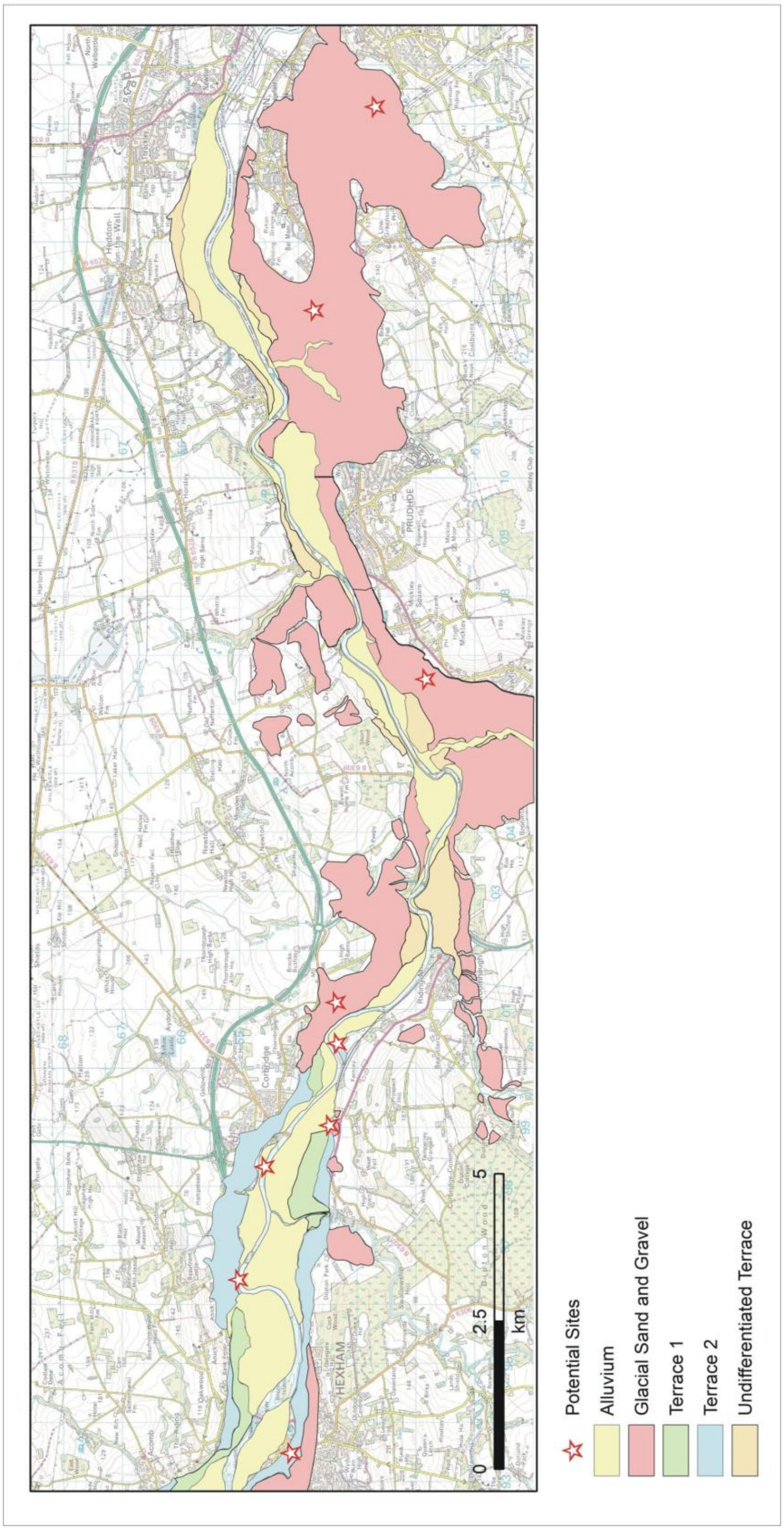


FIGURE 3.4: Possible sites identified for investigation in the lower Tyne Valley. Base map is OS 1:25,000 tiles (downloaded from www.edina.ac.uk/digimap/).

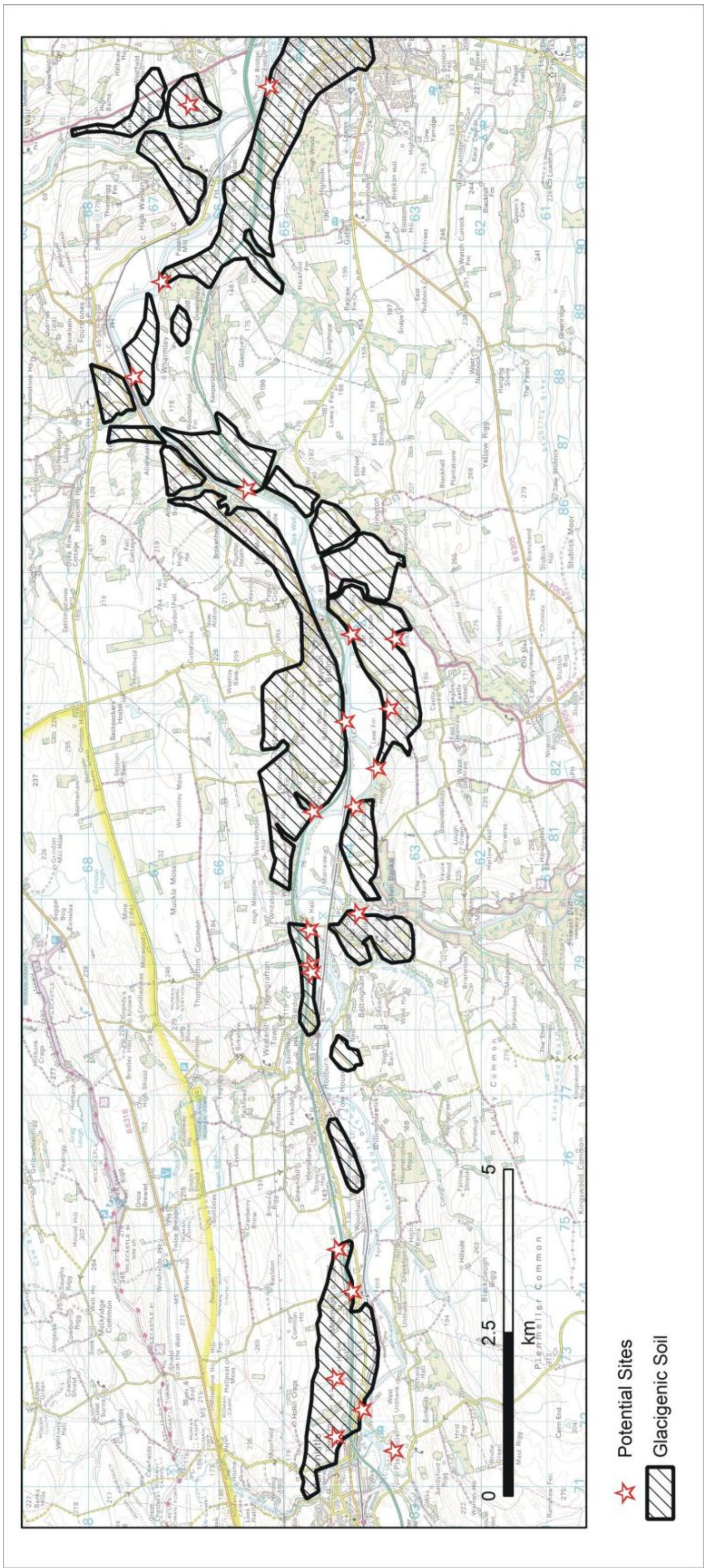


FIGURE 3.5: Possible sites identified for investigation in the lower South Tyne Valley. Base map is OS 1:25,000 tiles (downloaded from www.edina.ac.uk/digimap/).

3.5 Field sites

The site selection process produced 4 sites for detailed further study that fulfilled the criterion outlined above, and which represent the range of glacigenic landforms and contexts present in the Tyne Valley. The selected sites allowed the sediments beneath the present-day surficial morphology to be examined in detail (Figure 3.6), thus gaining a better understanding of the landform-sediment assemblage relationship. The sites at Crawcrook (NZ 412 564) and Stocksfield (NZ 064 617) provided access to undulating (mounds) topography, which extends along the southern flank of the lower Tyne Valley. The sections at Farnley (NZ 005 633) and Fourstones (NY 884 673) allowed the sediments underlying terraced landforms to be examined, which extend along the mid Tyne and lower South Tyne Valleys. Thus, study sites from locations distributed through the Tyne Valley represent a range of depositional environments and potentially different ages/time periods.

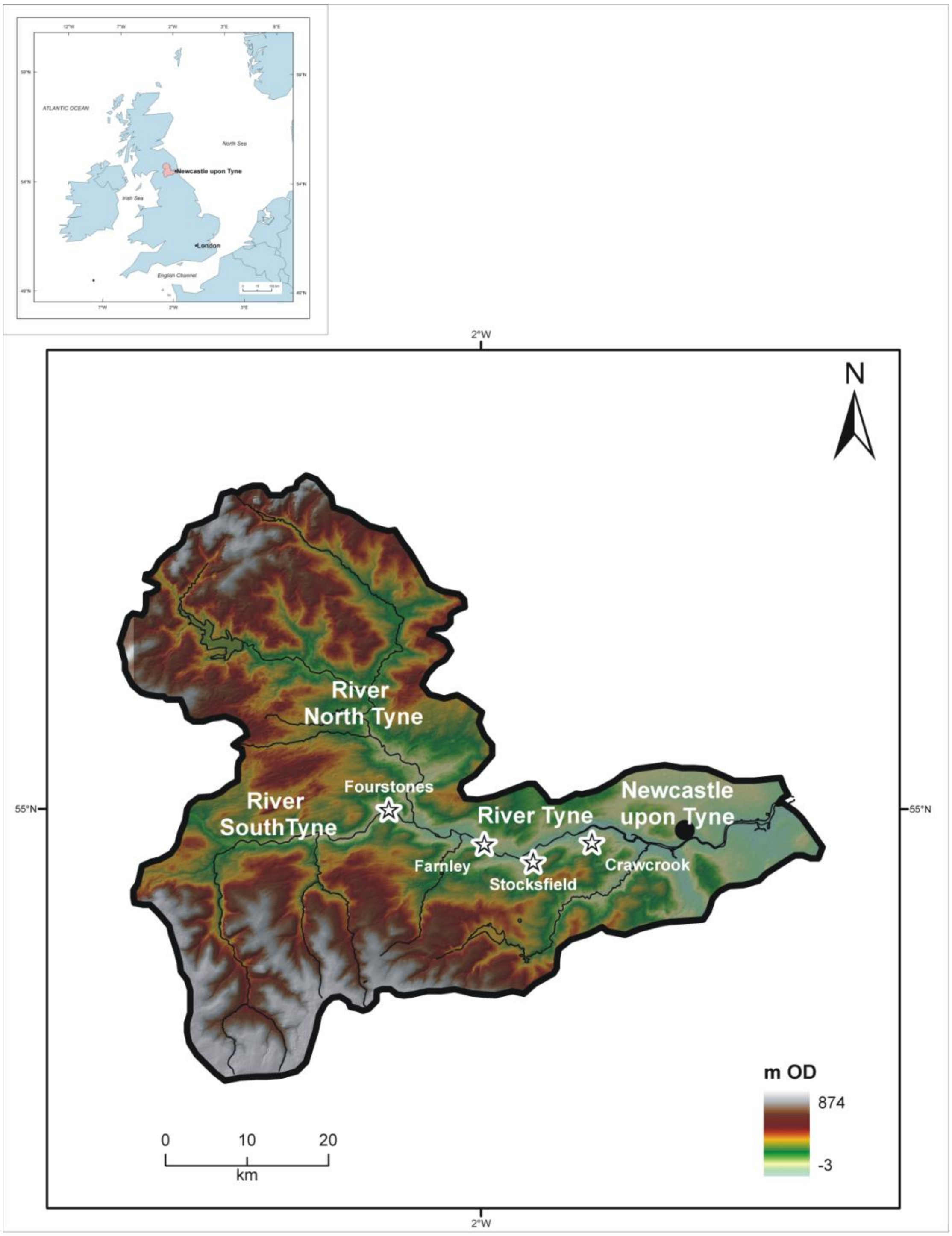


FIGURE 3.6: Location map showing the Tyne Basins and the four field sites chosen for further investigation (denoted by star).

Chapter Four

Research Methods

4.1 Introduction

This chapter outlines the field, laboratory and dating methods employed to investigate and gather data on the landform-sediment assemblages. The methods used to examine the study area included: field mapping; remote sensing and visualisation; spatial analysis and modelling; exposure measurement, recording and laboratory analyses; and geochronological dating. Each method is introduced with a brief explanation of the data that was generated and how each technique contributed to the aims of the project. Where novel methodological techniques were employed, a brief discussion on the background to the technique is given.

The chapter is divided into four sections, covering the various methods used to determine topography and morphology, sedimentology, postglacial fluvial response and geochronology. The aim of the methodologies employed was to draw together data that would contribute to the understanding of the genesis of the landscape in terms of development of the medium scale landforms and their relationship to deglaciation and/or postglaciation. This was determined from accurate recording of the features using both traditional and new approaches to mapping. The description and measurement of the sedimentary sequences enabled understanding of the aggradational environment of the sediments to be determined. Modelling of the sedimentary response since the end of deglaciation allowed the behaviour of the river and its response to external and intrinsic drivers to be ascertained through sediment budget calculations and assessment of stream competences. Finally, the development of a robust chronology using radiometric techniques allowed the sequence to be placed within the wider context of the climate changes that are known to have occurred since the LGM. Cumulatively, the collated data was then employed to test the hypotheses proposed in chapter 1.

4.2 Sedimentology

The depositional environment of the landform-sediment assemblage was determined by describing and measuring vertical profiles from sections at each of the exposure sites.

4.2.1 Photographs and field sketches

As part of the initial recording process, prior to describing the sections in detail, exposures were photographed. This enabled the identification of major features, such as bedsets, boundaries between units (erosional or exposure surfaces) and bounding surfaces, stacking patterns, depositional and deformation (ice-wedge casts, faults or collapse structures) structures, and channel features within the section, and were delimited on the photograph. Annotated photographs were taken back into the field for validation, and additional information recorded. Field sketches were then made to provide the detail necessary to help understand the depositional environments (Figure 4.1), and to supplement the logging and analysis process.

4.2.2 Logging and analysis

More than 40 sections were described and measured at sites along the Tyne Valley. Section location coordinates (UK National Grid) were established using a Garmin handheld GPS and elevation was obtained from topographic maps (1:10000). The

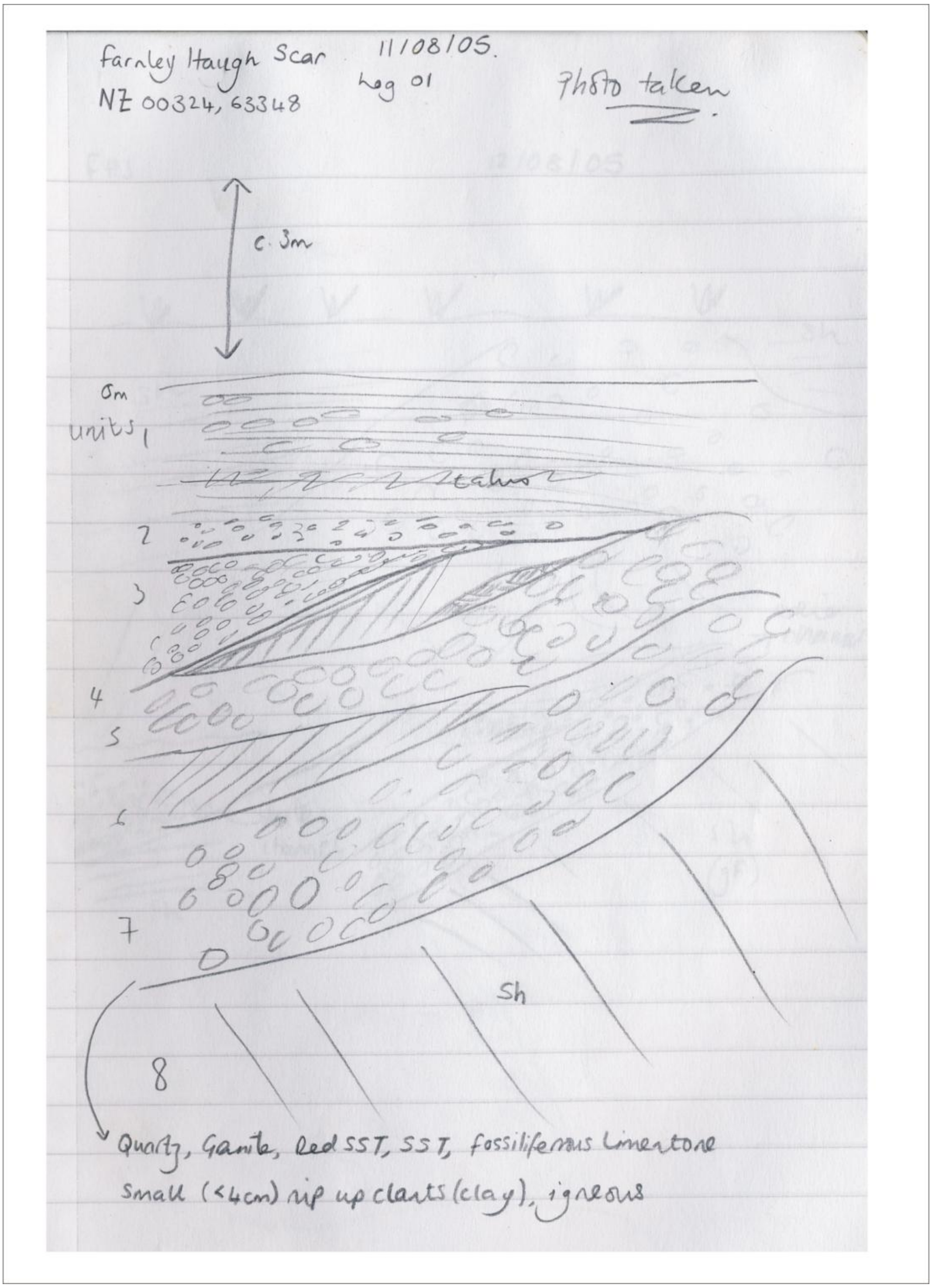


FIGURE 4.1: An example of a field sketch taken at Farnley Haugh Scar. (Scanned copies of all the field sketches drawn during this research can be found in the Appendix).

number of logs described was dependent on local stratigraphic complexities. Logs were terminated where sections were obscured by talus or vegetation. Sections were subdivided into units based mainly on texture and sedimentary structure and then described. Key characteristics recorded include: texture, sorting and modal grain size (particle size and sorting was determined qualitatively by reference to a visual grain size comparison chart), structure, colour (qualitatively assessed by comparison with a Munsell chart) and nature of contacts. From these descriptions, recurring sediment types were grouped into separate facies, each facies representing a broad depositional environment. Some units consisted of a single facies type, whilst others consisted of several facies either intermixed or graded from one to another. Lithofacies are intended to describe the sediment unit and to infer the depositional process and environment (Jones et al., 1999). Interpretation is dependent upon an evaluation of all the lithological (and geomorphological) evidence that is available, and the matching of this information with what are considered the best-analogue facies sequences from contemporary (glacial) contexts (Tucker, 2003). The basis for the lithofacies classification applied here has been the coding developed for braided fluvial deposits (Miall, 1978) (Table 4.1). Whilst these have a specific association with a depositional environment, the codes have been applied by other workers when describing proglacial, glaciofluvial and fluvial deposits (e.g. Dawson 1985; Dawson and Bryant 1987; Aitken 1995; Lewis and Maddy 1999). Vertical logs were drawn up using CorelDraw® v.11 to provide a comprehensive summary of the field sections.

4.2.3 Palaeocurrent direction measurements

Bedding and clast fabrics were measured as an aid to determining transport directions. The orientation and magnitude of the dip of the bedding was recorded. In gravels, the orientation and dip of the *a*-*b* planes (the angle between the plane the clast lies in and the horizontal plane) was measured with a compass-clinometer. The absence of well developed cross-bedding or sizeable clasts resulted in the restriction of some palaeocurrent measurements to a small number of *a*-axis orientations.

Surfaces were cleaned before taking measurements to ensure that *in-situ* clasts were measured and not those representing a recent surface slump. Although for statistical analyses 25 are deemed sufficient (Andrews, 1971), a total of 50 measurements were taken from each gravel unit. Mean orientations were plotted on polar diagrams using Orianna v.2 software (see Figs. 5.4;5.14 for plots).

4.2.4 Particle size

Determinations were made from the <4mm fractions of the matrices of the gravels and bulk samples from other sediment types. All samples were oven dried at 105^0C. Particle size was determined by the dry sieving and hydrometer-sieve method (cf. Syvitski, 1991). Matrix supported gravel was separated at -2ø (4mm), and the gravel sieved manually using standard Endecotts™ test sieves at half-phi intervals between -6ø (64mm) and -2ø (4mm). Grain size characteristics were determined for the <-2ø fraction using standard Endecotts™ test sieves in the range 4ø (63µm) to -2ø (at half phi intervals), nested and agitated on a mechanical shaker. Where the finer than 4ø (pan) fraction did not exceed 2% of the bulk sample it was discarded; those samples where the pan fraction exceed this amount were further analysed using sedimentation method to determine the silt and clay fractions. Results were drawn up using GRADISTAT© v.4.0 (Blott and Pye, 2001), a grain size distribution and statistics package (see Appendix for plotted data).

TABLE 4.1: Classification and coding of facies, lithofacies and sedimentary structures of modern and ancient stream deposits (Miall, 1978).

FACIES CODE	LITHOFACIES	SEDIMENTARY STRUCTURES	INTERPRETATION
Gms	massive, matrix supported gravel	none	debris flow deposit
Gm	massive or crudely bedded gravel, minor sand, silt or clay lenses	ripple marks, horizontal cross-stratification, gravel imbrication	longitudinal bars, channel lag deposits
Gt	gravel, stratified	trough crossbeds	minor channel fills
Gp	gravel, stratified	planar crossbeds	linguoid bars or deltaic growths from older bar remnants
St	sand, medium to v. coarse, may be pebbly	solitary (theta) or grouped (pi) trough crossbeds	dunes (lower flow regime)
Sp	sand, medium to v. coarse, may be pebbly	solitary (alpha) or grouped (omnikron) planar crossbeds	linguoid, transverse bars, sand waves (lower flow regime)
Sr	sand, very fine to coarse	ripple marks of all types	ripples (lower flow regime)
Sh	sand, very fine to very coarse, may be pebbly	horizontal lamination, partial or streaming lineation	planar bed flow (lower and upper flow regime)
Sl	sand, fine	low angle ($<10^0$) crossbeds	scour fills, crevasse splays, antidunes
Se	erosional scours with intraclasts	crude crossbedding	scour fills
Ss	sand, fine to coarse, may be pebbly	broad, shallow scours including eta-cross-stratification	scour fills
Fl	sand, silt, mud	fine laminations, very small ripples	overbank or waning flood deposits
Fm	mud, silt	massive, desiccation cracks	overbank or drape deposits formed in standing water

4.2.5 Clast lithology

Clast lithological analysis was carried out to: (i) determine the broad distribution of lithology; and (ii) to differentiate between deposits of different origins. Bridgland (1986) recommends counts of 250-300 gravel clasts between 16-32mm or 11.2-16mm size fractions. The cut-bank exposures at Fourstones and Farnley were sampled where gravel formations had been identified as a mappable unit (cf. Bridgland, 1986). Sections were cleaned of any debris and weathering before sampling. Bulk samples were collected by excavating an approximate 0.5 x 0.5m section in the face, and separating the samples in the field into the range 16-32mm using standard Endecotts™ test sieves. Although field walking identified significant surface gravel deposits in some of the terraces, the use of trial holes to access the gravels was not possible due to an impenetrable plough layer that could not be accessed without mechanical diggers. This was not permitted by the landowners because the land is continually in use for crop growing.

In the laboratory the samples were washed to remove any weathering debris, and the lithologies of 300 pebbles were identified. Clasts were broken to provide a clean face for identification. Clast identification was at broad level and clasts were categorised as sandstones, limestone, shale/mudstones, granite, quartz/quartzite and igneous/metamorphic (undifferentiated). Statistical analyses (*t*-test) were carried out to identify whether there was any significant difference in percentage of lithologies within and between sites and to look at intercomponent ratios i.e. local to exotic, and change through time.

4.3 Geochronology

Determining the age of glacigenic sediments is fundamental to being able to link timing of events to climatic shifts recorded in ice and marine sediment archives and therein understanding the terrestrial response to palaeoclimatic change during the late Quaternary. Currently there are no independent age controls for the sediments in the Tyne Valley. Historically, Pleistocene megafauna, such as Red Deer (*Cervus elaphus*) and aurochs (*Bos primigenius*), have been found in kettleholes in Northumberland and shell, plant macrofossil and wood fragments have been recorded in fine grained alluvial silts in the lower Tyne Valley (Howse, 1863). However, these sites no longer exist and due to a lack of fossiliferous and organic materials within the currently exposed deposits, radiocarbon dating or amino-acid racemization (AAR) could not be attempted. However, given that boreholes (cf. IMAU reports) and exposures have proved significant thickness of sands infilling the Tyne Valley, direct dating using Optically Stimulated Luminescence (OSL) techniques provided the greatest potential for constraining the sequence and to elucidate details of the deglacial chronology. The OSL technique directly dates sedimentation through determination of the last exposure of the mineral grains to daylight (Huntley *et al*. 1985; Duller 2004). Finally, calcite coatings on some gravels in the upper terrace unit offered the possible application of U-series dating (uranium-thorium disequilibria) techniques. By dating surficial carbonates formed on the gravels it is often possible to elucidate the timing of aggradation (cf. Candy *et al*., 2005).

4.3.1 Optically Stimulated Luminescence (OSL) Dating

Luminescence dating comprises a collection of numerical-age techniques that are amongst the most significant chronological tools used in Quaternary research (Lian and Roberts, p. 2449, 2006). As OSL dating is more widely applied to many sedimentary environments, there is an increasing body of empirical evidence indicating that optical dating provides an accurate index of the period of time elapsed since deposition (Stokes, 1999).

OSL dating can elucidate the timing of deposition of the sediments and sedimentary landforms. This is significant, as sedimentary landforms are direct proxies for environmental (climate) change (Lian and Roberts, p. 2450, 2006). The technique is particularly suited to quartz and feldspar grains (the most abundant common minerals on Earth).

4.3.1.1 Theory

The luminescence dating signal is acquired through exposure to daylight during transport. Following burial, the individual minerals begin to accumulate a trapped-charge population (Lian and Roberts, 2006). OSL dating attempts to determine the time elapsed since burial as the trapped-charge population increases with burial time in a measurable and predictable way (Murray and Olley, 2002, p.1). Electrons may be evicted (reset) from their traps by the addition of energy to the system (bleaching) as is the case when they are exposed to daylight. ‘Fast’ and ‘medium’ components are reset (i.e. quick to bleach) almost immediately following exposure to daylight (Murray and Olley 2002; Wintle and Murray 2006). The luminescence dating signal given by a mineral grain (quartz or feldspar) is the result of the optical release of charge (electron) trapped in the crystal lattice (Smith and Rhodes 1994; Bailey *et al*. 1997). Luminescence results when the addition of an external amount of energy *via* thermal or optical excitement allows the electrons to be released from traps (Aitken 1985; Stokes 1999; Lian and Roberts 2006).

Determination of the luminescence age is calculated by dividing the equivalent dose by dose rate, equation 4.1:

$$Age\,(ka) = \frac{equivalent\ dose\ (Gy)}{dose\ rate\ (GyKyr^{-1})} \qquad \text{[equ. 4.1]}$$

The equivalent dose (D_e), sometimes known as the palaeodose, is a measure of the accumulated luminescence signal in the sample since burial and is calibrated against known laboratory beta doses. The dose rate corresponds to the rate at which the sample was exposed to radiation in the environment (measured in the field or determined in the lab).

4.3.1.2 Optical dating of glacigenic deposits

In glacigenic environments incomplete bleaching of the mineral grains is a potential problem that has limited the application of OSL in the past. This occurs if any previous resident charge population within the crystal lattice is not substantially reduced or removed prior to burial, leaving a residual (latent) component (Stokes 1999; Murray and Olley 2002; Wallinga 2002). Incomplete bleaching results in an overestimation of the optical age from the luminescence signal (Duller, 2004, 2006).

Duller (1994a) identified two types of glacigenic sediments for OSL dating purposes: Type A where the whole population was equally bleached but the process was incomplete; and Type B where the population contained a mix of bleached and insufficiently bleached grains. Where samples contain either of these sediment populations, optical ages will be unreliable (Duller, 2004, 2006). The approach to dating such samples would require careful sampling of glaciofluvial sediments which are most likely to be exposed to daylight and OSL ages derived using multiple grains on each aliquot with supporting single grain measurements to investigate the bleaching history (Duller, 2006). The approached used in this thesis is outlined in section 4.3.2.3.

Gemmell (1999) investigated the bleaching history of the mineral grains of modern glaciofluvial environments in the Italian Alps and demonstrated that contemporary glacigenic sediments do not experience uniform zeroing of the luminescence signal and that sediments are deposited and re-entrained in the outwash environment. It is probable that Quaternary outwash sequences aggraded in the same way, with accumulation being varied both spatially and temporally, and accumulation areas would be regularly subject to erosion. Samples are more likely to reflect a mixed population and Gemmell (1999) advocates the use of the small-aliquot (multi-grain) or single-grain measurements to date glacigenic sediments rather than analysis of bulk samples. However, in a study of glacigenic sediments using modern analogues in southern Norway, Bøe *et al*. (2007) demonstrated that the material was well bleached and that exposure to daylight during transport of the Lateglacial glaciofluvial sediments was sufficient to reset the luminescence signal. Clarke *et al*. (1999) demonstrated the potential for optical dating of glacigenic sediments in Iceland. Their dates showed good agreement (within the errors) with an independently dated organic layer found beneath the sediments.

Good sedimentological analysis that informs landscape genesis models in combination with optical ages can help constrain those models and challenge existing interpretations. A study of outwash sediments in Germany, previously interpreted as MIS 6 deposits, applied optical dating to the sequence (Fehrentz and Radtke, 2001). The OSL dates suggested deposition was during MIS 2. The optical ages were used to constrain the type of depositional environment known to have existed during MIS 2 and the sequence was reinterpreted as aeolian and solifluction deposition. The interpretation was heavily dependent on the accuracy of the optical ages. The authors neglected to consider that the sedimentological interpretation of the site might be in conflict with the dates.

The reliability of optical (or any dating method) ages to inform models of landscape development will always be questioned. Gemmell *et al*. (2007) dated glacigenic sediments in north east Scotland thought to be MIS 4 deposits. The preliminary optical ages indicate deposition between MIS 5d and MIS 5a, suggesting the existing glacial model was inaccurate. Strickertsson *et al*. (2001) encountered the same problem when dating glacigenic samples from Denmark, which highlighted disagreement with the accepted regional history. Both concluded that the optical dates may be inaccurate due to the problems of incomplete bleaching discussed above. Owen *et al*. (2002) and Spencer and Owen (2004) dated glacigenic sediments from the Hindu Kush. They interpreted the depositional environments through geomorphological and sedimentological analysis and were able to demonstrate the reliability of the OSL ages by comparison with independent cosmogenic radionuclide dates derived from morainic boulders. Clearly, a combination of selective grain analysis (see section 4.3.1.3), where OSL ages can be shown to be both accurate and reliable, good supporting sedimentological work to interpret sequences accurately, and independently derived ages based on cosmogenic ^{10}Be or ^{14}C is required before any sequence or landscape can be fully understood.

4.3.1.3 Method of analysis

The methods of analysis adopted for OSL dating are multiple aliquot (sub-samples), single aliquot and single grain measurements. The multiple aliquot method is generally applied to aeolian (well-bleached) sediments and involves grain measurements on each aliquot made up from a sample (cf. Fuller *et al*., 1996). The aliquots are given different laboratory radiation doses and the results from all the grain measurements are combined to produce a growth curve that defines the luminescence response of the material to radiation. From the measurements a single estimate of equivalent dose is made as the method assumes each aliquot has the same equivalent dose. This assumption is true of aeolian sediments where optimum bleaching conditions exist at the time of deposition, thus, regardless of how many grains make up the aliquot or how many replicate measurements are made on different aliquots, the same equivalent dose would be obtained.

Glacigenic sediments consist of a mixed population of bleached and unbleached grains. Therefore, the multiple aliquot approach would result in some aliquots containing a high proportion of bleached grains giving a low equivalent dose, whereas other aliquots might contain a high proportion of unbleached grains that would give a high equivalent dose (Duller, 2004). As Murray and Olley (2002) suggest, if incomplete bleaching is suspected the standard method of analysis should be to examine the dose distribution using a small number of grains, or even single grains. Single aliquot methods ensure all measurements required for equivalent dose determination are made on a single subsample, thereby allowing comparison of equivalent doses obtained on different replicates (Duller, 1991, 1994b, 1995). During the last decade, advances in both the methods (cf. Roberts et al. 1998, 1999; Murray and Mejdahl 1999; Jacobs et al. 2003; Olley et al. 2004; Wintle and Murray, 2006) and in the equipment (cf. Duller et al. 1999; McCoy et al. 2000; Bøtter-Jensen et al. 2002, 2003) used for OSL dating have enabled single grain measurements to be made (Duller, 2006). Applying single aliquot or single grain measurements to glacigenic sediments can distinguish these bleaching histories and samples can be discounted where it is known that an inaccurate optical age will be derived (Duller, 2004, 2006). The sample size for a single aliquot can be varied. Reducing the number of grains used for each equivalent dose increases the chance of detecting a mixed population and it is possible to check reproducibility of the equivalent dose values (Wallinga, 2002). Where samples exhibit large variations in equivalent dose estimates between aliquots it is indicative that the luminescence signal was not reduced to zero for all grains, therefore making it possible to identify samples where luminescence ages may be unreliable (Wallinga 2002; Duller 2004).

Clarke (1996) and Clarke et al. (1999) suggested that the absence of scatter in equivalent doses obtained on single aliquots is a guarantee that the sample is well bleached. However, Wallinga (2002) noted that the scatter observed is largely dependent upon the number of grains that contribute to the luminescence signal of each aliquot. He suggested that to improve the chances of detecting poor bleaching using methods based upon inter-aliquot scatter in equivalent doses, the aliquot size should be taken as small as possible, preferably a single grain. Duller (2006) carried out single grain OSL measurements on glacigenic sediments from Scotland and Chile using the single-aliquot regerative-dose (SAR) protocol to determine the equivalent dose. He demonstrated that single grain measurements offer the possibility of obtaining reliable dates from glacigenic samples when the larger aliquot of the grains combined various bleaching histories and thus optical ages from these grains would have been an overestimate.

Obtaining an optical date is also conditioned by the ‘luminescence sensitivity’ of the quartz grain. Defects within the crystal lattice and chemical impurities (electron traps) store energy from the radiation it is exposed to and these increase with the number of erosional and depositional cycles (re-cycling) the grain has undergone (Stokes, 1999). Sensitivity is how efficient the quartz is at storing and converting the stored energy into luminescence emitted. It is a measure of the amount of luminescence emitted by a sample for a given radiation dose (Duller, 2004). Sensitivity in quartz grains tends to be low where they are freshly derived from the bedrock. Quartz grains that have low luminescence sensitivity are unlikely to produce any dates because there is a low store of electrons which can emit a luminescence signal. The higher the sensitivity (increased by bleaching and dosing cycles), the greater the potential for deriving an OSL date will be (A. Lang, personal communication, 2005).

On the basis of recent research it is encouraging to note that reliable dates are being obtained by OSL from glacigenic sediments and that the advancement in luminescence procedures means sediments that were only poorly bleached at deposition can be more readily detected (Duller, 2004, 2006). When combined with the 'lithofacies approach' (see section 4.3.2) to sampling, the potential for deriving ages from glacigenic environments has been greatly improved.

4.3.1.4 Errors and Uncertainties

Murray and Olley (2002) stress that the determination and subsequent reporting of the uncertainties associated with an optical age estimate is as important as the determination of the age, although invariably most studies do not report such errors. Reported optical age estimates are given with the standard uncertainty associated with the individual age. Single aliquot measurement uncertainties on the average equivalent dose are based on precision estimates i.e. the standard uncertainty on the mean is estimated from several independent, interpolated measurements of equivalent dose and, combined with equivalent dose measurements on single grains to examine dose distribution, this reduces the uncertainty associated with the optical age (Murray and Olley 2002; Duller 2004, 2006).

Errors in optical age estimates include systematic and random errors from beta-source calibration, dose rate determination, sub-sampling for chemical data and optical measurement errors (Colls *et al*. 2001; Rittenour *et al*. 2005). Murray and Olley (2002) concluded that some sources of error are difficult to avoid; these include conversion from concentration data to dose rate (estimated at ~3%), absolute calibration of concentration measurements (~3%), beta-source calibration (~2%), and beta attenuation factor (~2%). Although these estimates are approximate, Murray and Olley (2002) stress that it is difficult to obtain an optical luminescence age with an overall or combined standard uncertainty of much less than 5%. For example, Spencer and Owen (2004) reported the absolute uncertainty of the OSL ages in their study was as great as ~9 – 12%.

4.3.2 Sample strategy and treatment

Eleven samples were collected for OSL dating from the sand and gravel deposits at Crawcrook, Farnley and Fourstones and one from the palaeochannel infill exposed within the gravels at Fourstones. The sampling strategy was simple; lithofacies units were targeted that were most likely to have been exposed to daylight at deposition and therefore expected to show the most comprehensive levels of bleaching. A lithofacies approach to sample selection ensures the problems associated with poor bleaching (zeroing) of glacigenic sediments are minimized and works on the premise that the depositional environment, identified from the lithofacies, affects the zeroing of the luminescence signal and so lithofacies should be the primary factor governing sample selection (A. Lang, personal communication, 2005). The feasibility of a 'facies based' approach to sampling has been established through work on Sandur sequences around the Irish Sea Basin (Thrasher *et al*., 2007) and on glaciofluvial sediments in Sweden (Alexanderson and Murray, 2007). The sample locations were selected based on the lithofacies units. For example, at Crawcrook, sample LV161 was taken from an alternating medium/fine grained sand bed (Sh) where it was inferred that there was good potential for yielding an OSL age because the facies indicated shallow water deposition.

4.3.2.1 Geological setting

Glacigenic sediments in the Tyne Valley reflect the provenance up-ice in the Tyne Basin and in the ice source areas of the Lake District, Pennine and Cheviot Hills (Figure 3.2). However, the dominant bedrock in the Tyne Basin is Westphalian Coal Measure sandstones, and most of the sand is likely to be dominated by quartz with lesser amounts of feldspar. These sandstones have undergone several erosion/deposition cycles and as a consequence, it was anticipated that the quartz grains would have a high sensitivity. However, the depositional environment under which the sediments accumulated was both deltaic and fluvial, and the mineral grains may have not been bleached sufficiently to allow luminescence sensitivity to be built up.

4.3.2.2 Sample collection

OSL samples were obtained by hammering opaque, plastic sampling tubes (5cm diameter x 30cm length) into freshly cleaned sections of the selected lithofacies units (Figure 4.2). Duplicate samples were taken from each sampling point. Once removed the sample tubes were sealed in opaque plastic to avoid exposure to daylight. For determination of moisture content and gamma-radiation (environmental) dose rate, single subsamples (~100g) were collected from within 30cm of the dual sample point in the sediment exposure. Site coordinates (Latitude and Longitude) were measured with a Garmin handheld GPS and altitudes obtained from topographic maps (1:10000) (Table 4.2).

FIGURE 4.2: OSL sampling. A and B: Crawcrook, Location 1 (LV161). C and D: Farnley Haugh (Lv164).

The samples were processed and analysed at the Luminescence Dating Laboratory, University of Liverpool (five samples) and at the Research Laboratory for Archaeology and History of Art (RLAHA), University of Oxford (six samples). The OSL samples were treated to the same routine at each laboratory. Two laboratories carried out the OSL dating because the author was successful in gaining the QRA RLAHA funded Luminescence Award (Jan 2005).

Crawcrook:

Three samples were collected from the main exposure (location 1) in the quarry. The sediments were interpreted to be deposited in a shallow glaciofluvial environment (see section 5.2). Sample LV161 was collected from cm-thick horizontal beds of coarse to medium sands at the top of Unit 1. Sample LV162 was collected from a finely laminated, horizontal beds of medium to coarse sands, with occasional pebbles towards the base of Unit 1. Sample LV163 was collected from finely laminated (mm-thick) medium to fine sands within Unit 2 (Figure 4.2A,B).

Farnley Haugh Scar:

Three samples were collected from the cut-bank exposure of glaciolacustrine silty sands and fluvial sands and silty sands deposited within the channel and as overbank sediments (see section 5.2). Sample LIV164 was collected from cm-thick horizontal beds of medium to coarse sand that are interbedded with coarse gravel towards the base of Unit 2. Sample LV165 was collected from Unit 1, a finely laminated (mm-thick) silty fine sands below a major unconformity. Sample X2734 was collected from thinly laminated (mm-thick) fine sands immediately above gravel towards the top of Unit 2.

TABLE 4.2: Location and context of each dated OSL sample.

UNIQUE SAMPLE IDENTIFIER	LOCATION	LATITUDE	LONGITUDE	ELEVATION (M OD)	DEPTH BELOW SURFACE (M)	NATURE OF SAMPLE	MATERIAL DATED
X2730	Fourstones	54°59'59"	02°11'20"	49.5	0.7	Sh facies	Fluvial
X2731	Fourstones	54°59'59"	02°11'19"	46	5.5	Sh facies	Glacigenic
X2732	Fourstones	55°00'01"	02°11'14"	49	2	Sh facies	Fluvial
X2733	Fourstones	55°00'03"	02°11'05"	44	0.9	Sh facies	Fluvial
X2832	Fourstones	55°00'01"	02°11'12"	50	2	Fl facies	Fluvial
X2734	Farnley Haugh Scar	54°57'54"	01°59'47"	37	2	Sh facies	Fluvial
LV161	Crawcrook	54°58'23"	01°48'11"	47	1	Sh facies	Glacigenic
LV162	Crawcrook	54°58'22"	01°48'09"	45	3	Fl facies	Glacigenic
LV163	Crawcrook	54°58'21"	01°48'08"	41	7	Fl facies	Glacigenic
LV164	Farnley Haugh Scar	54°57'54"	01°59'48"	28	10	Sh facies	Fluvial
LV165	Farnley Haugh Scar	54°57'54"	01°59'48"	23	15	Fl facies	Glacigenic

Fourstones:

Four samples were collected from glaciofluvial silty sands, flood deposit sands and overbank sands that are exposed along the cut-bank (see section 5.2). Sample X2730 was collected from Terrace 1, Unit 2, thinly laminated (cm-thick) horizontal and planar beds of medium sand immediately above a major flood deposit. Sample X2731 was taken from Unit 1 of the Terrace remnant cut in to Terrace 1, a finely laminated (mm-thick) fine sands and silts. Sample X2732 was taken from Terrace 3, Unit 2, horizontal and planar beds of medium sand immediately above a boulder fill deposit. Sample X2733 was taken from Unit 2 of Terrace 4, a finely laminated (cm-thick) horizontal medium to fine sands immediately above gravel. Sample X2832 was taken from Unit 1 of Terrace 3, a horizontally laminated silty clay channel deposit within bedded gravels.

4.3.2.3 Sample preparation and measurement

Sample preparation and measurement were the same in both laboratories. Sample tubes were opened in the laboratory under photographic dark-room conditions (i.e. low sodium light). In the laboratory, the light exposed ends of the sample were removed and set aside and the remaining sample was wet sieved to recover quartz grains at 180-212 and 212-250 μm in diameter. The quartz grains were then treated with 10% and 33% HCL and H_2O_2 to ensure the removal of carbonates and organics. Density separation was carried out using sodium polytungstate to eliminate heavy metals, and subsequently, to separate quartz (2.70-2.62gcm^{-2}) from feldspars. The quartz grains were then etched in 40 % hydrofluoric (HF) acid for 80 min and then subsequently treated in 10% HCL. Finally, aliquots of different sizes of grains (~400 grains and ~100 grains) were mounted on stainless-steel discs using silicone gel. Equivalent dose (D_e) was determined through measurement using a Risø TL/OSL reader. Analyses were carried out on multiple small aliquots using a single aliquot protocol.

4.3.4 Accuracy

The accuracy of OSL ages is verified by using independent age controls. For example, Passmore and Macklin (1994) previously dated the Holocene sediments at Farnley using OSL and were able to compare these to independent ^{14}C chronologies. However, the OSL ages were shown to be unreliable (I. Fuller, 2007, personal communication). Currently, the existing independent ages for the Tyne Valley only constrained the period from the mid Holocene onwards. With this is mind, the opportunity to test the reliability of the OSL dates was limited. Although a small number of gravels (~30) from Farnley Haugh were thinly coated with calcium carbonate it was not possible to obtain a reliable Uranium-series age from such a small sample of calcite (~100mg) to allow comparison with an OSL chronology (D. Hoffman, personal communication, 2007).

4.4 Palaeoecological analysis

Bulk samples (~100g) of silt-rich deposits were examined for plant microfossil (pollen) and insect macrofossil (Coleoptera) remains. Three samples came from channel fill deposits at Fourstones and a single sample came from glacigenic lake sediments at Crawcrook, as they were thought to contain some organics which could augment the sedimentological data with palaeoenvironmental (proxy climatic) data. Subsamples were prepared for pollen analysis in the usual way using HCL, HF and Acetolysis treatment (cf. Moore *et al*., 1991). For Coleoptera analysis the insects were obtained by wet sieving a disaggregated sample over a 300μm mesh and concentrating the retained residue using standard floatation techniques (cf. Coope, 1986). Pollen analysis was carried out by Claire Twiddle (University of Hull) and insect analysis was carried out by Dr Lynda Howard (University of Loughborough).

4.5 Geomorphological mapping

The form and spatial arrangement of the landform-sediment assemblages was determined by comprehensive mapping.

4.5.1 Geomorphological mapping

Geomorphological mapping is one of the most important techniques used in geomorphological research (Gardiner and Dackombe 1983; Cooke and Doornkamp 1990; Anderson *et al*. 1997). Maps produced through field mapping not only contain information on morphology and genesis, but in some cases the age of the landforms (Lowe and Walker, 1997). Extensive field mapping should be carried out to represent accurately the spatial and genetic complexity of the environment (Anderson *et al*., 1997). A considerable amount of time was required in the field to map accurately the areas around each of the four key sites. Field mapping (Figure 4.3A) spanned a few months, and was followed by 1-2 months to reproduce the map in a digital format using ArcGIS® (Figure 4.3B).

Within all former glacial environments, the complexity within the landscape is apparent, and the Tyne Valley is no exception. Thus, it was crucial to represent accurately the geomorphology so that the deglacial story could be disentangled. Mapping was carried out at the small scale i.e. 1:10000, so that a detailed picture could be built up at each study site, where an area of up to ~4km^2 was mapped to capture the pattern of landforms associated with the sediments exposed. In the field, the spatial arrangement of the landform-sediment assemblages (e.g. outwash terraces, eskers, kames, kettleholes, river terraces) was determined by accurately recording the landforms onto base maps. To ensure every landform was accurately recorded on the base map, landforms were identified by systematically walking across the field area and

A.

B.

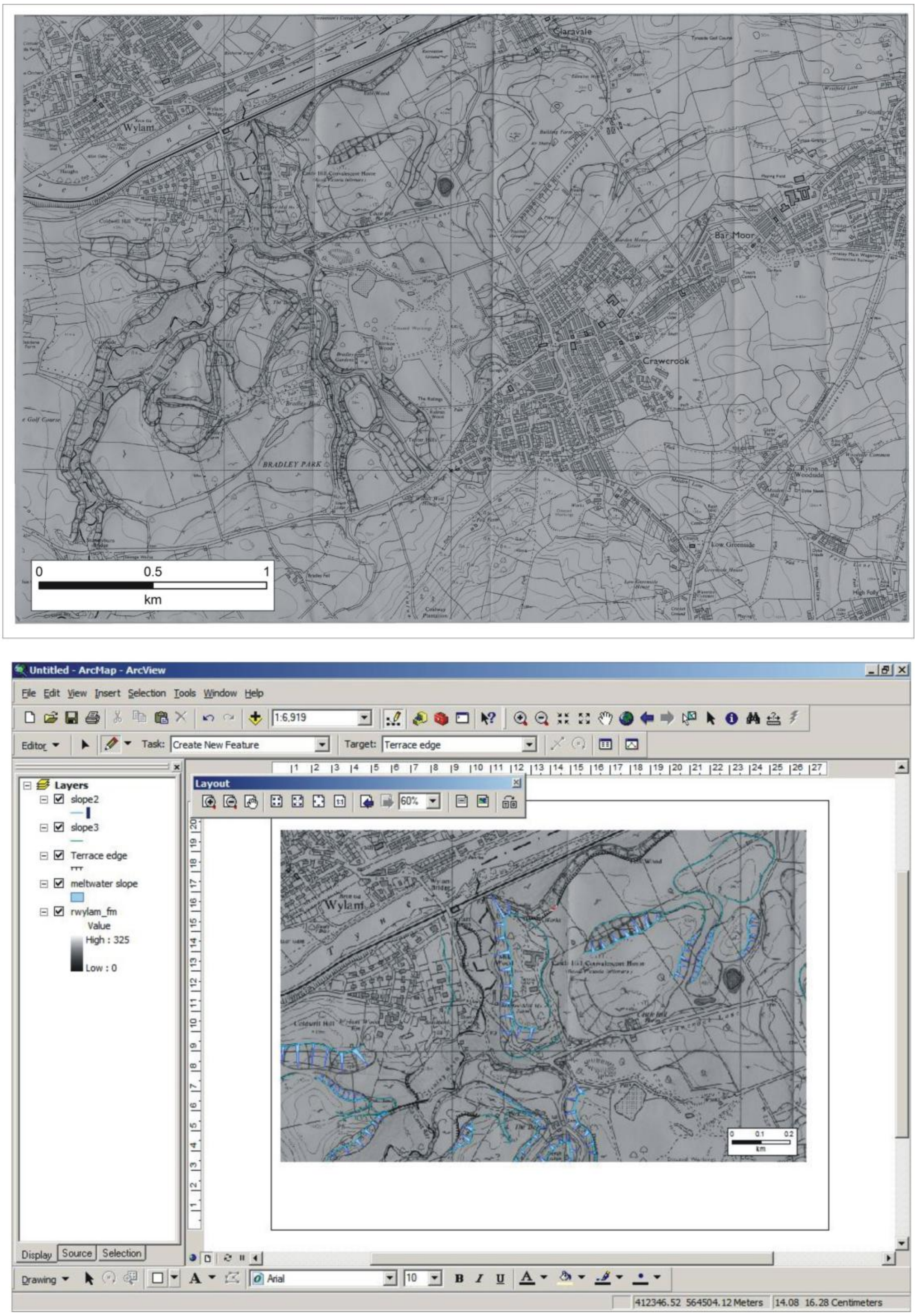


FIGURE 4.3: A. Hand drawn fieldmap of the area around Crawcrook. B. Scanned and geo-referenced fieldmap displayed in ArcMap; lines are digitised as polylines (using editor) to enable a digital map to be produced.

across every landform. By mapping in this way, it ensured that features were not recorded from a distance, which results in distortion of the actual landform shape due to perspective and misrepresentation of the features (Mitchell, 1991). Landform boundaries were delimited by reference to features located on the base OS map. The degree of accuracy of the map and the amount of geomorphological detail is consequent upon the ability of the mapper and the time available. To represent realistically the landforms and features found in former glacial environments, the landforms were recorded on the maps by a set of symbols to represent appropriately the geomorphology of the landscape.

Using ArcGIS, the base maps were redrawn (Fig. 4.3B). The field map was converted to digital format by scanning the field slips and archived to '.jpg' format. The .jpg file was georeferenced to the OSGB36 coordinate system, using the Georectification tool in ArcMap, and a first-order polynomial transformation and nearest neighbour resample, so that the features when digitised would be in the same coordinate system as the geo-rectified aerial photographs and DEMs (see below).

In ArcMap, the geo-rectified base maps were redrawn and digitised using the editor tool. Shape files (.shp) were created in ArcCatalogue, and the landforms were digitised as polylines and polygons, which were represented by a set of ESRI cartographic symbols (Figure 4.4).

4.5.2 GIS visualisation and remote sensing

Field mapping in the study area was complimented by mapping using thematic visualisation models derived from Digital Elevation Model (DEM) data to determine the spatial distribution of the landforms and to understand broad-scale processes and relationships in landform development,. The manipulation of DEMs in *Geographical* Information Systems (GIS) allows the data to be visualised in 3-dimensions. Combining air photographs (APs) with thematic visualisations of the DEM, to construct photo-realistic views of the landscape, aided the identification and mapping of landforms along the river valley, and filled in gaps between sites that were field mapped.

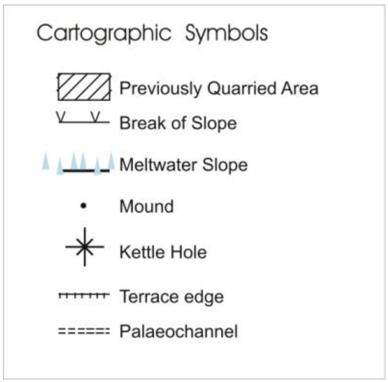


FIGURE 4.4: Key to the cartographic symbols employed in field mapping.

4.5.2.1 Digital Elevation Models

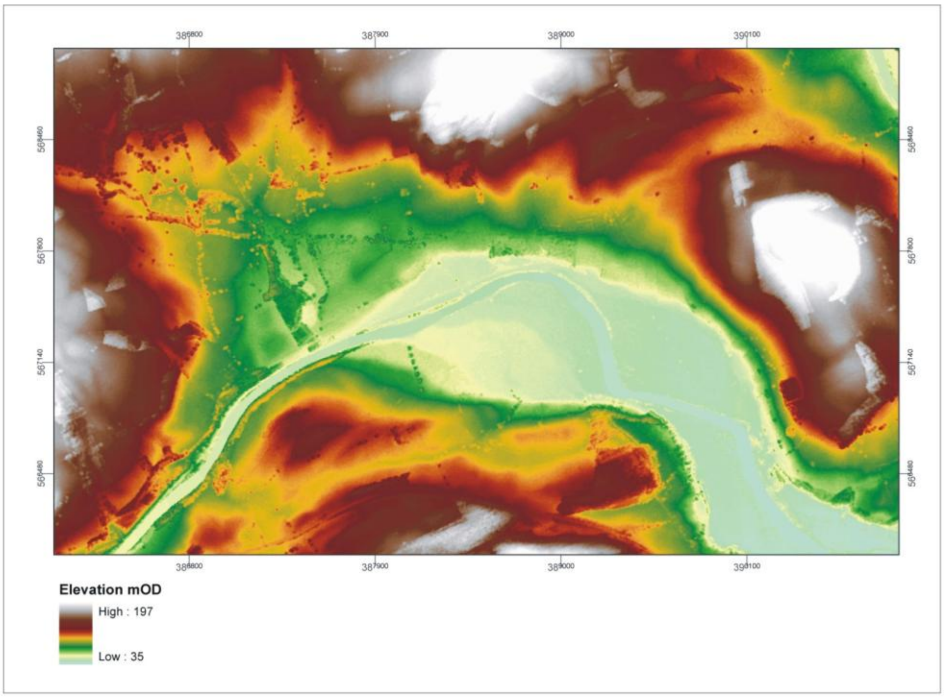
The wealth of remotely sensed data varies in its spatial scale, from the shuttle radar data at one end and the dGPS survey data at the other. The appropriate scale of the remotely sensed dataset used in visualisation must relate to the scale of the features that are to be identified. In this study, features are medium-scale (10^{-1} - $10km^2$) erosional/depositional landforms, therefore the optimum remotely sensed dataset needs a grid resolution of between 5-10m. Digital elevation models (DEMs) at this scale are available from airborne mapping techniques. Light Detection and Ranging (LiDAR), which uses a laser to measure the distance between the aircraft and the ground, provides high resolution DEMs at grids of <1m (cf. Dowman, 2004). Whilst this scale is ideal for mapping recent and small scale valley floor change (cf. Jones *et al*., 2007), in this study, where the features of interest are of the order of 10s of metres, the LiDAR resolution grid would not enhance the identification processes. NEXTMap Great

Britain™ topographic elevation data sets (commercially available DEMs that were commissioned by Norwich Union) provide coverage for the whole of the UK at a scale of 5 and 10 m grid spacings. NEXTMap elevation data is acquired using airborne Interferometric Synthetic Aperture Radar (IFSAR) technology, which compares two radar images taken at slightly different locations to obtain elevation information. IFSAR is configured with a single transmit antenna and dual receiving antennas vertically separated; the separation is referred to as the baseline. The two images are combined interferometrically i.e. the phase difference between antennas 1 and 2 coupled with precise aircraft positional data, to provide the information required to derive the terrain elevation points (Allen 1995; MacKay 2005; Mercer 2004). Downman *et al*. (2003) concluded that on surfaces which are free of trees and other obstructions NEXTMap data gives an excellent representation of the terrain surface at both grid spacings. Smith *et al.* (2006) evaluated the available data sources by comparing field mapping with landform identification based on thematic visualisation models derived from NEXTMap Great Britain™, OS Panorama® and LiDAR for a glaciated landscape in Scotland. Whilst they concluded that LiDAR was the superior product in terms of ground resolution and provided the greatest potential for landform mapping, they also determined that detailed mapping can be satisfactorily generated from the NEXTMap DEM, giving a meaningful approximation to field mapping.

4.5.2.2 NEXTMap DEM datasets

In this study, two digital elevation models were obtained from NEXTMap; a surface model (Digital Surface Model, DSM) and a bald-earth model (Digital Terrain Model, DTM) (Figure 4.5). The DEMs have a horizontal (spatial) resolution of 5m and a 0.5-1m vertical accuracy. The DSM is a first return image and provides a similar surface

A.



B.

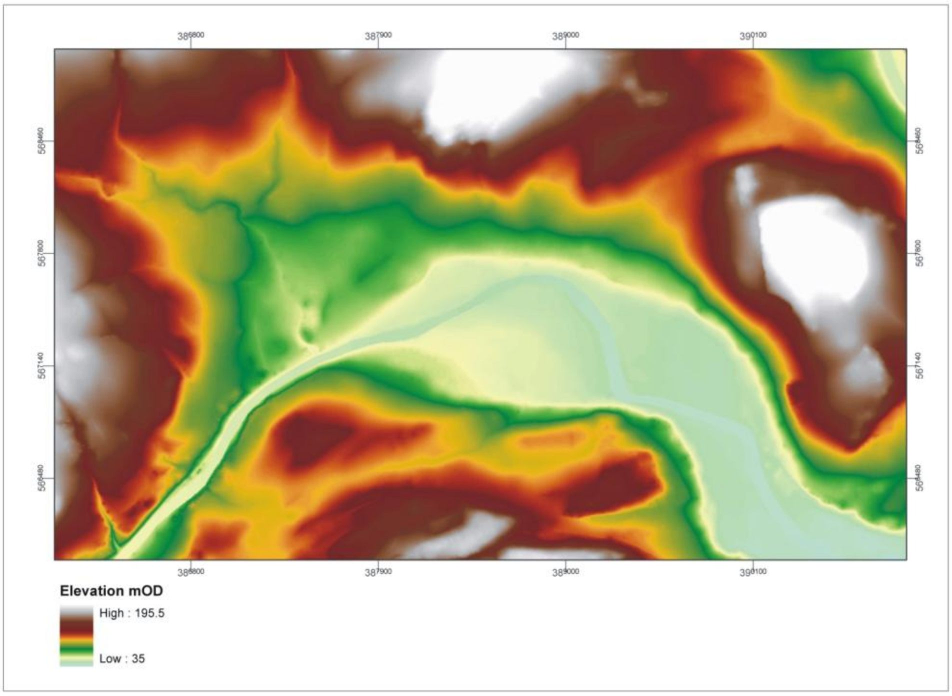


FIGURE. 4.5: NEXTMap tiles for the area around Fourstones for comparison. A. DSM tile (colour-shaded). B. DTM tile (colour-shaded).

representation to LiDAR. However, IFSAR is unable to penetrate dense forest cover, thus there is no returned image from the ground surface in these areas (cf. Downman *et al*., 2003). Thus, the DTM is a processed image where the surface cover has been removed using an algorithm rather than representing the actual elevation points on the ground that would be derived from second return. This gives the DTM image a smoothed appearance. There is some residual 'noise' (artefacts) in the DTM, with artefacts that appear as 'pin-pricks', 'puck-marks' or 'grooves' and relate to general vegetation, forest plantations and road and rail line removal. Landforms are easy to distinguish on the DSM, although in some locations the surface cover masks feature boundaries. Therefore, both the DSM and DTM models were used for the visualisations, as they provide slightly different views of the landscape, which aids landform identification.

4.5.2.3 Air Photographs (APs)

To provide an additional layer of spatial landform information that complemented the DEM visualisations, air photos covering the study area were acquired. Black and white vertical APs of the Tyne Basin were available at the 1:10560 scale (1969, from Fairey surveys, ©Simmons Aerofilms) and at the 1:10000 scale (1978, University of Cambridge Aerial Photographic unit). There were 4 flight lines covering the Tyne Valley between Hexham and Blaydon, giving good spatial coverage of the valley bottom and sides but only a single flight line along the South Tyne Valley between Haltwhistle and Hexham which covered the valley floor only. The APs, which consist of high-quality mapping camera images, were scanned using an Epson A3 flat-bed scanner at a resolution of 400 dpi (dots per inch), and archived to '.tif' format. This gives a reasonable visual resolution for photo-interpretation, where 1 pixel is equivalent to 0.635m on the ground. Scanned APs were loaded into Jasc® Paint Shop Pro, the border removed using the crop tool (dragging from a fixed start to a fixed end point at diagonally opposite corners of the photo) and images were rotated to their correct orientation using the rotate tool.

4.5.2.4 Georectification of the APs

To use the scanned APs in a form that could be viewed in conjunction with the DEM required their conversion to a common projection and coordinate system, i.e. georegistration. Polynomial georectification is readily applied to large sets of APs (e.g. flight line along a river valley) and is widely used for georegistration of APs (Hughes *et al.*, 2006). The scanned photos were georectified with reference to the Ordnance Survey of Great Britain 1936 (OSGB36) National Grid by using a set of Ground Control Points (GCPs). GCPs on the target (unregistered) layer are given coordinates from the corresponding point on the base (georeferenced) layer. The translation between the two coordinate systems is called an 'affine transformation', and incorporates translation, scaling and rotation. The predicted locations X',Y' are compared with the real values of locations of the control points X,Y and the image warped along the point locations (Hughes *et al*., 2006).

Images were loaded into ArcMap so they could be georegistered. A minimum of 25 pairs of GCPs were selected from a digital OS 1:10000 raster tile base layer (downloaded from www.edina.ac.uk/digimap/). These were well distributed over the image because the key features of interest are spatially distributed across the photograph and were a mix of hard (sharp edge or corner) and soft (irregular or fuzzy edges) GCPs (cf. Hughes *et al*., 2006). Transformations using curvilinear (quadratic or higher) polynomial functions are the most popular for aerial photos because they can correct for some of the effects of both radial error (related to curvature of the Earth) and geometric error (related to topography and camera lenses distortion) and because they lend map-like qualities to a georectified photo without orthorectification (Hughes *et al*., 2006). Using the Georectification tool in ArcMap, images were georectified to the OSGB36 coordinate system using second- and third-order (quadratic, cubic) polynomial transformations, though some warping at the edge of the photo was associated with the cubic transformations as the image was stretched (i.e. rubber-sheeting) to match the control points. To provide geometric correction, the georegistered images were resampled **(**modified) by applying a cubic convolution interpolator (nearest neighbour). Finally, to create a single (digital) image for each key study area, georectified APs were combined (mosaic[*]). Air photos overlap along the same (~60%) and adjacent (~20%) flight lines which allows them to be combined to form a single image (National Air Photo Library, 2004). The mosaic tool (an add-in for ArcMap downloaded from the ERSI® site) was used to mosaic the APs.

4.5.3 New approaches to landform geomorphology: visualisation of medium-scale landscapes

To identify ice contact or stagnation landscapes, traditional methods have relied on interpreting printed topographic maps or classical field geology, which although very effective, are time consuming (Brown *et al*., 1998). However, in recent years, the emergence of easily available remotely sensed datasets where large areas of the UK are covered means it has been possible to map large areas in less time (cf. Clark, 1997) and to pose questions about regional landscape evolution (cf. Baker, 2007). Visualisation in GIS is about displaying information (DEM, Air Photo, geology, etc.) in such a way that it becomes possible to extract/reveal information and develop hypotheses. With the availability of high quality, high resolution DEMs, perhaps the most promising data

[*] A mosaic is simply a digital reproduction of a series of APs put together in such a way that the detail of one photograph matches the detail of all adjacent photographs (National Air Photo Library, 2004).

source for future research as they record absolute elevation and can therefore be used to visualise landscapes (cf. Smith *et al*., 2006), opportunities abound to map landforms across large areas with relative ease. GIS software enables the creation of thematic visualisation models such as shaded-relief, or digital cross-sections and 3-demensional surface modelling which aid rapid and accurate landform mapping and are increasingly used in geomorphology in preference to contour maps (Burrough and McDonnell 1998; Jordan *et al*. 2005).

Information can be generated either by quantitative (objective) assessment of the data to develop a supervised or unsupervised landform classifications or through a qualitative (subjective) visual assessment of thematic visualisation models. The former method offers an un-biased way of assigning landform classes to a DEM. Supervised classification involves generating a training set of data describing the multiple morphological terrain descriptors derived from thematic maps (slope, elevation and aspect) that are then used to parameterise and define simplistic morphometric classes using an algorithm which can then be applied to another area of the DEM to automatically classify the landforms (Brown *et al*. 1998; Adediran *et al*. 2004; Bolongaro-Crevenna *et al*. 2005; Prima *et al*. 2006; Demoulin *et al*. 2007). This method has been successfully applied to mountainous areas, areas of tectonic relief and even river valleys. However, automated classification of landforms involves generating a training set from which the algorithm is parameterised to identify the landforms and in this study there was not an existing classification output by which to train the classification algorithm.

Qualitative methods, based on recognition of landform elements, involve the visual inspection of thematic visualisation models (grey-scale, contours, shaded-relief and 3D surface views) derived from the DEM and is similar to the interpretation of remote sensing (e.g. APs) images. Shaded-relief models are the most commonly used for identification and mapping features such as landslips, tectonic faults and glacial landforms (Oguchi *et al*. 2003; Jordan *et al*. 2005; Smith and Clark 2005; Van Den Eeckhaut *et al*. 2005). Smith and Clark (2005) examined shaded-relief, slope and surface curvature visualisation models to aid the identification of glacial landforms in Ireland. They concluded that no single thematic visualisation model provided complete or unbiased mapping, and recommended using profile curvature and shaded-relief models to start mapping an area. Given that visual inspection of thematic visualisation models has been successfully employed to map glacial landforms in Ireland and Scotland, this was decided as the preferred approach to landform identification.

4.5.3.1 Producing the geomorphological map from the thematic visualisation models

In order to test further the use of thematic visualisation models, a pilot study was carried out using the DEM of the Crawcrook study area. The reliability of feature identification from the visualisations for mapping purposes was determined by comparing the visualisation-derived map to the independently-produced geomorphological map, which is taken to be a true and accurate representation of all the landforms in the study area. To evaluate the thematic visualisation models for landform identification, the second phase of the assessment was to apply the methods for visualisation of the landscape to a DEM of the un-mapped area of the River South Tyne Valley between Bardon Mill and Haydon Bridge where Lunn (1995) suggested some very prominent landforms could be seen in the field, and then validate the identified landforms through field observation. Based on field mapping in the Tyne Valley, the prevalent landforms can be divided into three broad morphological units. These were:

(1) terrace (break of slope and lateral persistent),

(2) mound (irregular outline and steep slopes) and

(3) valley (sinuous or linear depressions).

To identify and record these landforms, the DEM was reduced to simple relief elements by manipulating the raw data using visualisation techniques to produce three thematic visualisation models: (1) relief-shading, (2) slope, and (3) profile curvature. Prior to visualisation, the DTM image underwent some image processing i.e. spatial filters were applied to improve visual interpretation in the final image. The thematic visualisation models and image processing was carried out using the spatial analyst extension and ArcToolbox in ArcMap and the features were digitised on-screen using the editor function in ArcMap.

4.5.3.2 Image processing using spatial filters

Spatial filtering involves (mathematically) breaking the image into spatial frequencies, emphasising certain frequencies relative to others to enhance or suppress spatial detail and then recombining the frequencies into an enhanced image. Convolution filtering is the common mathematical method for implementing spatial filters. Each pixel value is replaced by the average over a square area centred on that pixel. Square sizes typically are 3x3, 5x5, or 9x9 pixels, and the square array is often called a moving window. Using the spatial analyst functionality in Arctoolbox, filters were applied to the DTM. To remove the some of the artefacts in the DTM a 3x3 low-pass filter (i.e. separates the low frequency, slowly altering component from the remainder of the image) was applied three times, which tends to reduce deviations from the local average, thus smoothing the image and removing or reducing the ‘noise’. To enhance feature detection in the DTM, a 3x3 high-pass filter (i.e. separates the high frequency, rapidly altering component from the remainder of the image) was applied three times, which emphasises fine detail, abrupt discontinuities and edges. Even with some image processing, artefacts (e.g. rail line) could still be seen in the image, thus to ensure true features were identified and not artefacts, features were cross-checked with the APs and digital OS maps.

4.5.3.3 Thematic visualisation models: Relief-Shading

Relief-shading has been shown to be a valuable tool for landform identification at the medium-scale (cf. Jordan *et al*. 2005; Smith and Clark, 2005; Van Den Eeckhaut *et al.*, 2005). Relief-shading highlights subtle variations in the topographic surface by illuminating the landscape from a particular direction (azimuth) and elevation (zenith), allowing a realistic depiction and interpretation it is increasingly the default visualisation technique for landform mapping (Onorati *et al.* 1992; Ouguchi *et al.* 2003). Although the interpretation of relief-shaded maps is based upon the recognition or identification of morphological elements, it is important to be aware of its limitations, such as azimuth biasing. This is when a single azimuth illumination is used but whilst this may highlight some features others are made invisible because of their relationship to the illumination direction (Graham and Grant 1991; Onorati *et al*. 1992; Smith and Clark, 2005).

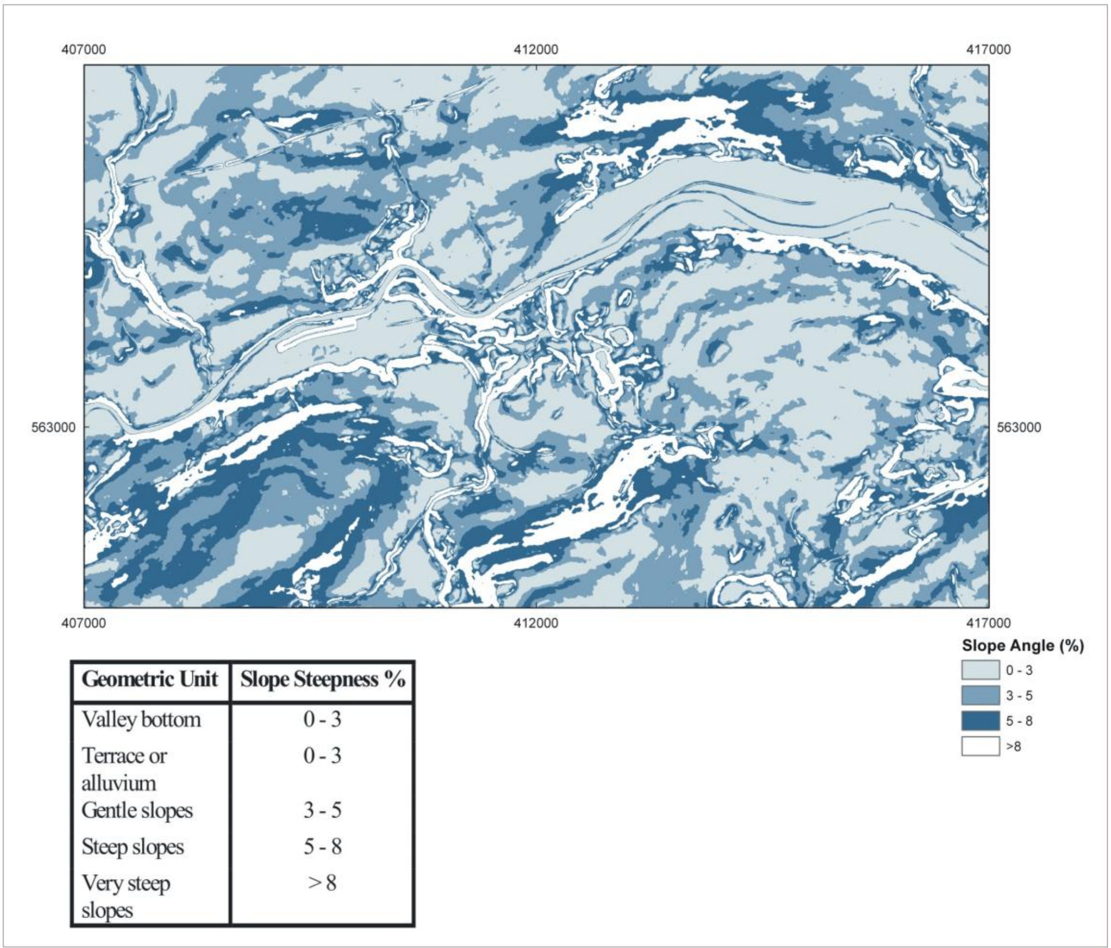
Rather than using trial and error to determine which combination of illumination azimuth and zenith angle contributed the most information to the relief-shaded model, a systematic approach was taken using a Principal Components Analysis (PCA). Using the spatial analyst tool and hillshade extension in ArcMap, relief-shaded models of the Crawcrook DEM were created at 8 different azimuth angles and 6 different zenith angles and a standard PCA was applied using the spatial analyst tools in ArcToolbox. The azimuths were north (0^0), north-east (45^0), east (90^0), south-east (145^0), south

(180^0), south-west (245^0), west (270^0) and north-west (315^0) and zeniths were applied every 15 degrees between 15 and 90 degrees (15^0, 30^0, 45^0, 60^0, 75^0, 90^0).

4.5.3.4 First and second surface derivatives: Slope (gradient) and Curvature (change in gradient)

In traditional field mapping, break of slope is used to mark boundaries between geomorphological units (cf. Gardiner and Dackombe, 1983), thus slope is a good indicator of the type of features in an area i.e. delimiting feature outlines, lateral persistence of features and break of slope. Both the slope gradient and local relative relief all measure the shape of the surface. Using the Slope function in ArcMap, a slope map was calculated from the DEM. A colour-shaded (stretched but restricted to a narrow elevation range to encompass the heights of landforms under examination) model highlighted clearly topographic expressions such as break of slope and prominent feature outlines. To enhance further the features of interest, the slope map was re-classified in to 5 geometric units (Figure 4.6). This classification was based upon the slope breaks in the cumulative slope histogram.

Profile curvature is the rate of change of slope per unit distance and is based upon the recognition of slope variation and change of curvature across the DEM (Demoulin *et al*., 2006). Curvature models highlight rapid changes at the base and top of a slope. Where there is no curvature a flat surface can be inferred, ridges are signified by convex curvature (positive sign) and slopes by concave curvature (negative sign). Using the spatial analyst tool and curvature function in ArcToolbox, two curvature models (profile and planiform; Figure 4.7) were generated of the DEM. The profile map delimits features such as slope-breaks, valleys and ridges very clearly, whereas the planiform map indicates the drainage network, clearly highlighting valleys and channel routing (drainage network).



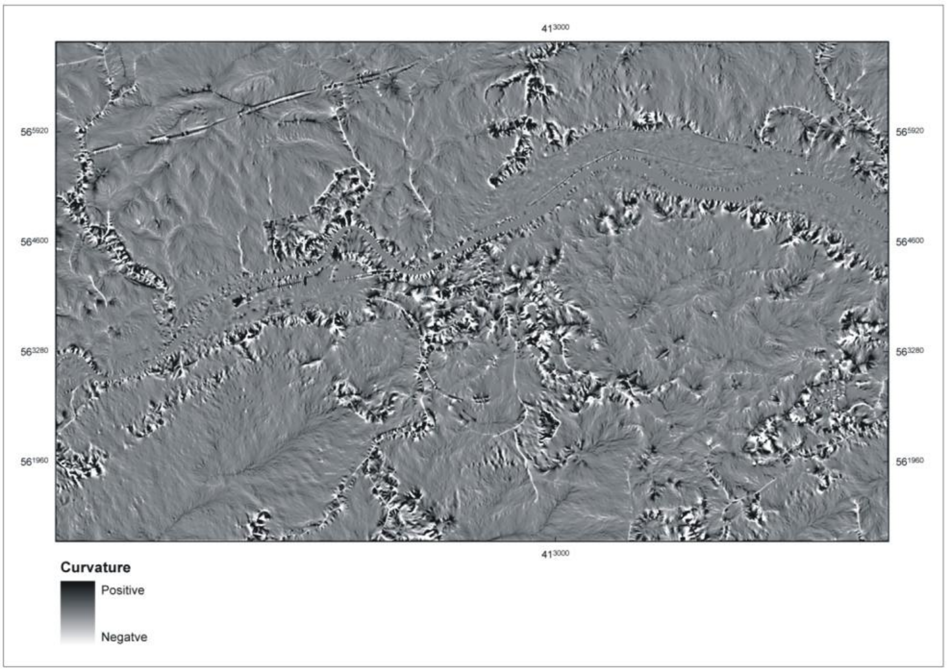
Geometric Unit	Slope Steepness %
Valley bottom	0 - 3
Terrace or alluvium	0 - 3
Gentle slopes	3 - 5
Steep slopes	5 - 8
Very steep slopes	> 8

FIGURE. 4.6: Slope map of Crawcrook area classified into geometric units. Inset table gives the classification of geometric units by slope steepness (in per cent).

4.2.4 Differential Global Positioning System (dGPS) survey

The accuracy of the NEXTMap DEM was assessed by comparison with field survey elevation data. Intensive mapping of the surface topography above the River South Tyne at Fourstones was carried out using a Leica GPS 1200 series (with a precision of 10mm horizontal and 20mm vertical) and resulted in the collection of 17,621 *x,y,z* data points. GPS phase data observations were collected by a roving receiver in the field and also by a static receiver recording base station data, located in a field nearby (i.e. within 1km). The phase data was used to compute a high precision (cm accuracy) GPS solution, which was assumed to be the "true" position of the rover at each point. Data collection was accomplished using Leica GPS 1200 series receivers. Phase data

A.



B.

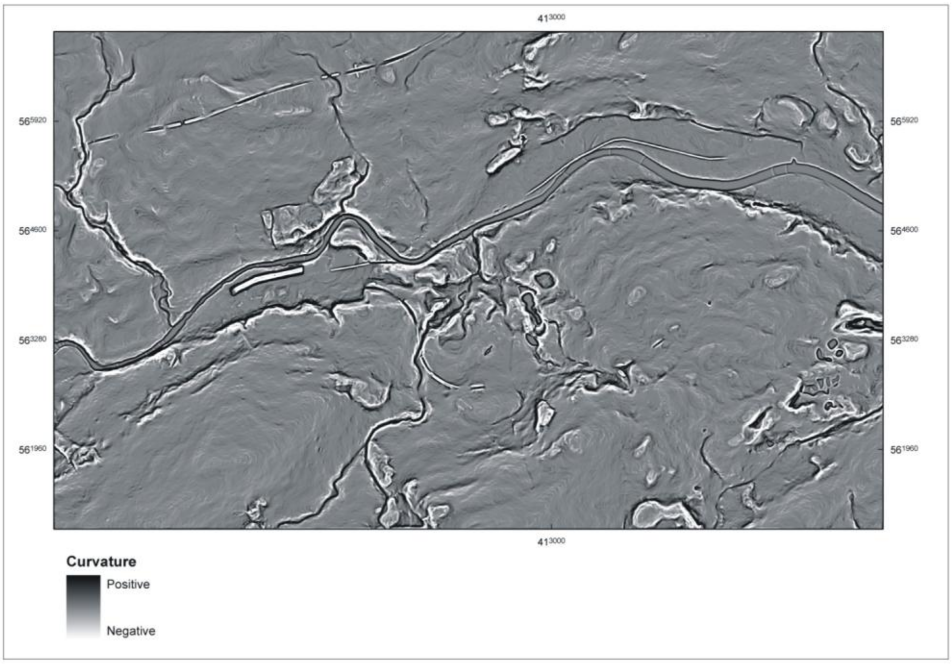


FIGURE 4.7: Curvature models (generated in ArcMap) of the NEXTMap DTM tile for Crawcrook. A. Planimetric curvature. B. Profile curvature.

processing was achieved with Leica Geo Office software. All coordinates were transformed from World Geodetic System 1984 (WGS84) to Ordnance Survey of Great Britain 1936 (OSGB36) National Grid. The data points were spatially interpolated to create a DEM surface from the set of *x,y,z* measurements using Geostatistical Analyses (see section 4.6.1.4).

4.5.5 River Terraces

To identify the lateral extent and spatial distribution of the terraces, all flat surfaces between Crawcrook in the lower Tyne Valley and Melkridge in the lower South Tyne Valley that might be terraces or remnants of terraces were mapped from a combination of field survey, DEM visualisation, air photo interpretation and OS (1:25000) map analysis. Terraces were distinguished from Holocene cut and fill units and the modern floodplain by their elevation relative to the present river level. Holocene units lie between 8 and 1m above the present river (Passmore and Macklin, 1994, 2000, 2001), therefore, all flat surfaces that lie above 8m above the present river level were assumed to be Lateglacial terrace surfaces, with no assumption of their origin which may have been as river, outwash, kame or strath terraces or bedrock benches (e.g. those that crop out above the River South Tyne between Haltwhislte and Fourstones).

4.5.5.1 Morphological Analysis

The good resolution of the DEM meant it was possible to discern terraces less than 3m apart in height. Terrace scarps were clearly delimited on air photos through the identification of physical boundaries such as field boundaries, hedgerow or fence lines which are characterised by some degree of sinuosity indicating former channel banks. Contour intervals on OS 1:25000 maps were examined to identify terraces. Widely spaced contours generally indicated a relatively flat surface and comparison of this data with the air photo helped to clarify the landform. The integration of field mapped data allowed terrace edges to be traced along the valley by extending the mapped boundary in combination with the methods above. Finally, cross-sections of the valley floor were generated from transects positioned across the DEM (NEXTMap DSM) image using the spatial analyst extension in ArcMap. Cross-sections were located at the upper, middle and lower sections of the study area. Prominent benches in the cross profile that were continuous throughout the study area were defined as terraces. Terraces that had the same elevation on opposite sides of the valley were assumed to be paired.

4.5.5.2 Longitudinal Profiles

The gradient of the river terraces was established to examine the former profile of the river and to determine the controls on base profile change. In general, the preferred method of terrace correlation is through detailed surveying from reach to reach, following surfaces rather than matching relative heights. However, it was not practical to spend several months in the field purely on producing highly detailed long profile maps. Therefore, as with most other workers (e.g. Bridgland 1994; Maddy *et al*. 1995; Maddy and Bridgland 2000), correlation of terraces along the Tyne valley was based upon matching the heights of the 'terrace' above present river level. Rather than carrying out isolated field surveys, the DEMs obtained for the visualisations were employed as they provided almost complete coverage of the Tyne Valley. Altitudinal spot heights on each terrace were obtained from the DSM. These points were selected from the centre line of each terrace, thus avoiding colluvial slumps or degradation at the terrace bluff and edge. A relative terrace chronology was established on the basis of height. The terrace longitudinal profiles were reconstructed from the altitudinal spot heights and plotted, along with the present day river long profile, as elevation versus distance along the valley (see Fig. 5.7 for data). Profiles were examined for change since the LGM i.e. knick-points and regression analysis was used to predict former base levels.

4.6 Quantification of valley infill (distribution and volume) using a geomorphometric approach

During deglaciation large masses of sediment were available in river systems. Combined with low vegetation cover and high transport capacity the River Tyne was likely to have eroded, reworked and deposited vast amounts of glacigenic sediment during the Lateglacial period. It is widely acknowledged that a paraglacial sediment cycle (cf. Church and Ryder 1972; Church and Slaymaker 1989) continues long after glaciation. However, quantification of the volume of Late Devensian sediment aggraded in river valleys, constrained by reliable dates, has not been obtained in the UK and little is known of the paraglacial cycles in upland glaciated valleys. Furthermore, the magnitude of sediment delivery and reworking during the deglacial period, and subsequent denudation rates are also unknown (see section 4.6.2). Therefore, to determine volumes of sediment stored and reworked, a morphometric approach was taken to model the valley profile, in order to calculate sediment volumes and budgets.

4.6.1 Mathematical modelling: cross valley profile prediction

To quantify the amount of sediment storage within the present day river valley, a bedrock model was first produced. This was achieved by modelling the valley sides and predicting a valley bottom profile. Glacial and/or glaciated trough valley cross-sections are generally referred to as being U-shaped (parabolic), while valleys that underwent little glacial modification tend to be V-shaped, characterised by more or less linear slopes (Hirano and Aniya 1988, 1989; Harbor 1990; Harbor and Wheeler 1992;

Pattyn and Van Huele 1998). Most glaciated cross valley profiles consist of both erosional and depositional elements, as the erosional profile is usually partly covered by glacial and postglacial fill (Aniya and Welch, 1981). Cross valley profiles can be adequately described by a simple power curve (parabolic; cf. Graf, 1970). Thus, curve fitting using parabolas can predict the rockhead profile, and consequently, be extrapolated beneath the glacial and postglacial deposits for an estimation of valley fill thickness (Aniya and Welch, 1981). However, although the power equation is valuable as it provides a steady description of glacial valley cross-sections (Li *et al*., 2001), this approach has the *a priori* assumption that the cross valley profile form is parabolic and symmetrical (Harbor and Wheeler 1992; James 1996). Schrott *et al*. (2003) employed an alternative polynomial function to estimate the shape of preglacial valley floors in the Bavarian Alps, now infilled with glacigenic deposits. The advantage of using polynomials is that they allow valley shape modelling with only minimal data requirements. To avoid fitting curves where the valley form has been partly obscured by glacial and postglacial deposits, Aniya and Welch (1981) recommend that a best fit curve be generated for the erosional surface, thus excluding data which represent the depositional elements of the cross section. Although the application of polynomials for valley shape modelling is speculative in its character and actual surveys are highly recommended (Schrott *et al*., 2003), it provides a reasonable and cheap approximation (Greenwood and Humphrey, 2002).

4.6.1.1 GIS modelling approach

The mid South Tyne Valley and the mid Tyne Valley are characterised by steep valley side slopes and narrow valley bottoms, whereas the lower Tyne Valley has low valley slopes and wide valley bottoms. Initially, where there was good valley symmetry, polynomial functions were applied to cross valley transects taken across the surface

DEM. However, it was apparent that the polynomial function was not forecasting for valley bottom/rockhead elevations but predicting the present day valley profile. Therefore, in order to calculate the depth of valley infill, the valley bottom data was removed so that polynomials were generated based on the valley side profiles only. This was achieved by dividing the valley into two parts: the glacial and postglacial valley infill, and the adjacent valley sides, which although they were entirely under ice during the last glaciation, have relatively thin surficial sediments. The elevation to which the present day valley is infilled was removed from the surface DEM so that pre-infill cross sections could be predicted by mathematical modelling. The elevation at which glacial sand and gravel (valley fill) was reasonably determined by obtaining the highest elevation at which surficial deposits (i.e. sand and gravel) have been recorded by the IMAU survey: approximately 150m above OD. Till was not included as it is ubiquitous due to the complete inundation of ice over this area. The underlying assumption was that the valley sides above this level represent the pre-glacial valley profile with relatively little modification (cf. Mills and Holliday, 1998). The NEXTMap DEM tiles obtained for this study did not provide complete valley coverage i.e. ridge to ridge cross-sections could not be obtained. Therefore, the surface DEM used for the generation of valley cross profiles was provided by Landmap (www.landmap.mimas.ac.uk). The Landmap DEM is a bald-earth model derived from LANDSAT satellite data, and has a grid resolution of 25m. Generation of the bedrock DEM followed the steps outlined in Figure 4.8.

4.6.1.2 Geomorphometric analysis: cross valley transects

Cross valley transects were obtained for the area between Haltwhistle and Blaydon. The surface profiling tool EZ Profiler 9.03, an add in for ArcMap downloaded from the ERSI® site, was used to generate *x,y,z* values for the transects. The *x,y* coordinates were extracted every 10 map units; this distance was dependent on the length of the relative transect, but on average points were taken every 30m. Given the base DEM from which the values were extracted is a 25m grid this is an appropriate sampling resolution. Due to the irregular placement of transects, which were located by eye along

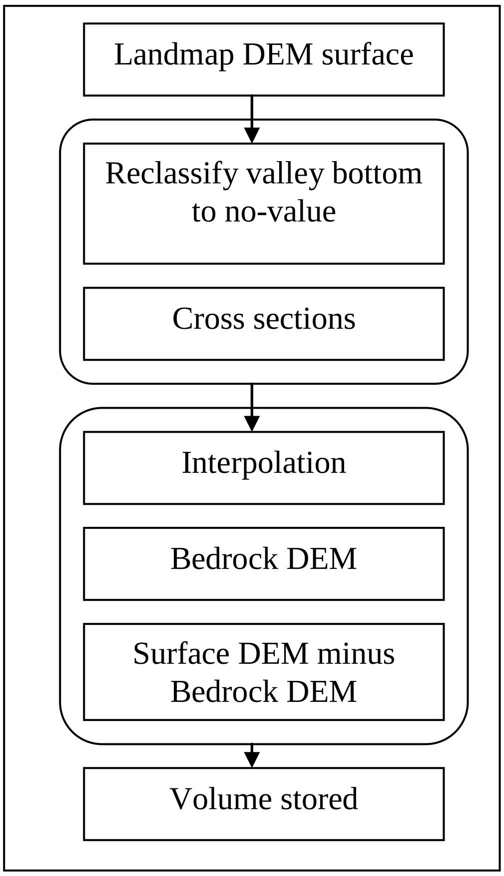


FIGURE 4.8: Flowchart showing the methodology for calculating the sediment volume.

the DEM to avoid sections where tributary valleys enter the main valley, the data points did not form a grid of evenly distributed points. However, subsampling the data to create an even grid only reduced the quality of the interpolation, thus analyses were performed on the full dataset. The attribute (*x*,*y*,*z*) data was exported to an Excel spreadsheet, where the valley bottom profiles were predicted using Schrott *et al*'s (2003) approach. For each cross valley transect, a polynomial function (2nd to 6th order) was used for modelling the profile. Although this model assumes the valley profile is both parabolic and symmetrical, which is not true throughout the valley, it offered the best approach to generating a useful model. To further condition the polynomial model, known *z* values at the rock head were obtained from those borehole logs that encountered bedrock adjacent to transect locations and the data included so the polynomial curve would pass through that point. Empirical data was fitted with a quadratic equation (2^{nd} order polynomial) in the form of equation 4.2:

$$y(x) = A + Bx + Cx^2$$

[equ. 4.2]

where elevation (z) data was then generated based on the polynomial equation for given x,y locations.

At some cross valley locations, the polynomial function was an inappropriate model to predict valley depth, because the slopes above the valley infill were poorly defined (i.e. not smooth or straight). For these cases, using Hinderer's (2001) approach a simple linear regression model was applied to each valley side. Assuming a V-shaped valley profile, the intersection of the constructed regression lines yielded the maximum valley depth. Although this was a rather crude method, the resulting regression line fitted the valley sides more closely. Minimisation of the actual valley bottom width was inevitable, unless known bedrock depths from borehole logs in the vicinity of the transect were included to qualify the model. Empirical data was fitted for each valley side slope to a linear regression equation of the form of equation 4.3:

$$y(x) = ax + b$$

[equ. 4.3]

Generation of *z* data for the bedrock DEM was based on predictions from the two models, polynomial and linear. Spatial interpolation was carried out on a full set of data points. The linear model included *x,y,z* data from the available borehole logs. Analysis was carried out on a training data set (80%), leaving a validation data set (20%) to check the accuracy of the predictions. The creation of a second bedrock DEM from the borehole logs which encountered bedrock, to test and validate the geomorphometric results, was not possible because the resolution of the data points (only 58) was too low. Spatial analysis (see 4.4.1.4) was performed on the data and DEMs generated.

4.6.1.3 Spatial distribution of sediment within the valley: Borehole records

Development of a three-dimensional surface map (cf. Challis and Howard, 2003) of Quaternary sediment (glacial drift, river terrace and alluvium) thickness was not possible due to the limited amount of information available from IMAU borehole log data and too few points (<20) to give a reliable interpolation. To calculate absolute sediment thickness infilling the valley floor, interpolation was performed on the data provided by 91 borehole logs (IMAU; Giles 1981; Lovell 1981). Although, 33 of the borehole logs do not encounter bedrock, the data they provide give 'at least' sediment thickness values.

4.6.1.4 Spatial analysis and GIS modelling approach

To generate a DEM surface from the *x,y,z* data, the data needed to be interpolated. Spatial interpolation is a set of procedures for estimating the values of the attributes at unsampled locations based on the set of observed values at known locations (Burrough, 1986). The justification underlying spatial interpolation is the assumption that points closer together in space are more likely to have similar values than points more distant (U.S. EPA, 2004).

There are a number of *ad hoc* and geostatistical methods available for interpolating data; these include Nearest Neighbour, Inverse Distance Weighting (IDW), Splines and Kriging (Simple, Ordinary and Universal). Deterministic methods (Nearest Neighbour, IDW, Spline) utilise no information about the system being analysed other than the measured data. Nearest Neighbour techniques are polygonal or proximal techniques (e.g. Theissen polygons) and predictions of attributes at unsampled locations are provided by the nearest single point. IDW should be an exact interpolator i.e. predicts a value identical to the measured value at the sample location, and is commonly used for simplicity to produce a DEM (cf. Moore *et al.* 1991; Chappell *et al.* 2003). However, the weights do not sum to 1 and the surface is a locally varying mean so it will not pass through the sample points (M. Charlton, personal communication, 2007). It predicts values by assigning more weight to the points closer to the measured data points in the averaging formula, though it does not make assumptions about spatial relationships. Splines (radial basis functions, RBF) are an exact interpolation model but also make no assumptions about the data. Splines work like a rubber-sheet as the methods seek to minimise the overall curvature of the estimated surface while passing through the measured data points. Splines do not perform as well when there are large changes in the surface within short distances, such as terrace edges, and are not constrained to the range of measured values.

Geostatistical methods for interpolation, known as kriging, recognise that the spatial variation[2] of any continuous attribute is often too irregular to be modelled by a simple, smooth mathematical function. Instead the variation can be better described by a stochastic surface. The attribute is then called a “regionalised variable”. The variability in the space of the data is characterised by the semivariance, which is a

[2] Spatial variation is recognised in the theory of Regionalised Variables developed by Matheron (1963, 1971) which has developed into geostatistics; the application of geostatistical methods enables the spatial variability of the interpolated data to be estimated.

measure of the deviation between pairs of z-values at a certain distance and direction (Zuuring, 2004). Geostatistical techniques create surfaces incorporating the statistical properties of the measured data. Kriging, like IDW, is a weighted average method of interpolation. However, in kriging the weights are not only determined based on the location of the data but also on the degree of spatial continuity (or autocorrelation) present in the data as expressed by a semivariogram. The weights are determined so that the average error of estimation is zero and the variance of estimation minimized. Predicted values are not constrained to the range of measured values. To predict a surface using kriging, the spatial dependency must be determined from variograms and covariance and a model fitted. With simple kriging the mean is known. Ordinary kriging is where the expected value of the underlying trend of z is assumed to be constant over the study area and assumes a constant but unknown mean (equation 4.4). The weight, λ_α, depends upon a model fitted to the measured points, the distance to the prediction location and the spatial relationships to the measured values around the prediction location. With kriging the weights sum to unity so the surface passes through the sample points (M. Charlton, personal communication, 2007).

$$P(z_{\mathrm{I}}) = \sum_{\alpha=1}^{N} \lambda_\alpha P(z_\alpha)$$

[equ. 4.4]

where:

P : main variable (elevation)

Z_{I} : point to interpolate

Z_α : points at observed main variable

N : number of points to include in the linear combination

λ_α : weights of the linear combination of the observed main variable

Universal kriging (UK) belongs to non-stationary geostatistical methods (Wackernagle, 2003) and is used when there is known external drift in the data as it allows the specification of a drift polynomial to describe the underlying trend and assumes a varying mean over space (ESRI 2004; Krivoruchko and Gotway 2004; U.S. EPA 2004; Zuuring 2004; Tatalovich 2005). The predictions for UK are made by equation 4.5:

$$P(z) = m(z) + \mu(z)$$

[equ. 4.5]

where:

z = point (x, y), where (x, y) are coordinates

$P(z)$: the stochastic process

$m(z)$: mean of the generating stochastic process

$\mu(z)$: random component

The semivariogram is displayed in a plot of semivariance against distance (lag) where semivariance is calculated for all pair of locations at distance *h* as:

$$h = 0.5 \times \text{average}\left[(\text{empircal covariance})^2\right]$$

[equ 4.6]

The plotted semivariogram is then a measure of the spatial correlation or covariance between the *x*,*y* data points for the *z* values (Figure 4.9). Usually semivariance (g(h), gamma) will increase with *h* (lag, a distance vector), indicating greater deviation and lower correlation between z-values with increasing distance. Often after some lag(h) distance, the Range (a), g(h) will level out at a level called the sill (C_0+C_1). The range is the distance beyond which the deviation in z-values does not depend on distance and hence z-values are no longer correlated. To compute the experimental semivariogram, data are divided into distance classes, lags and an estimation of gamma is made for each lag (Zuuring, 2002, 2004). Ripley (1981) and Stein (1988) both show that, for efficient production, it is often the behaviour of the variogram nearer the origin that needs to be captured. Thus it is really only necessary to model the small-scale part of the covariance structure. The prediction procedure essentially copies this local behaviour elsewhere.

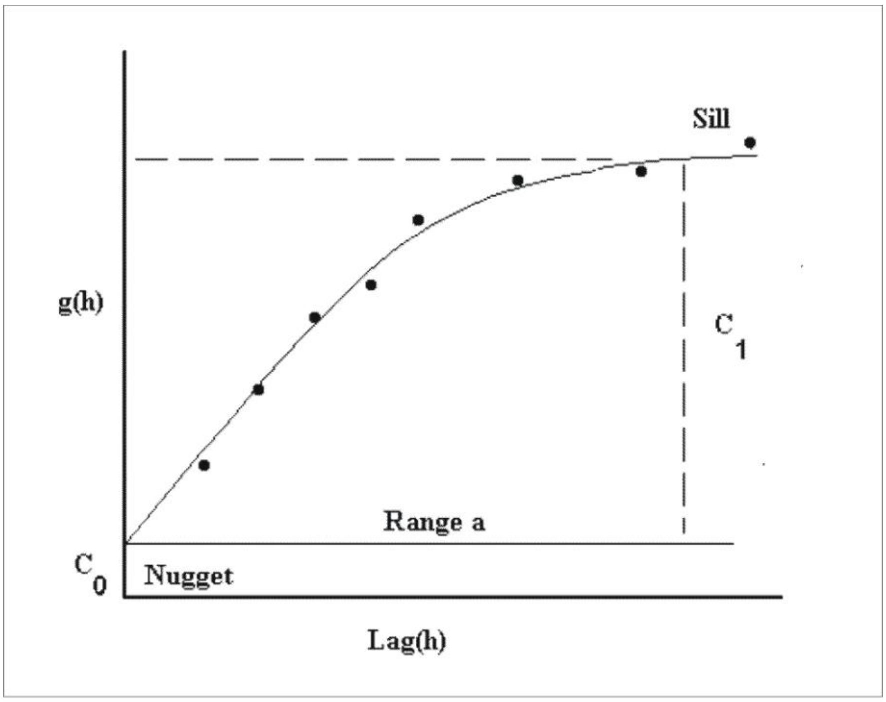


FIGURE 4.9: Theoretical semivariogram (taken from ArcView Help © ESRI).

A number of theoretical models can be fitted to the experimental semivariogram (e.g. Spherical, Circular, Exponential, and Gaussian) (Figure 4.10). Discontinuity at the origin, the nugget effect (C_0)**,** is caused by microscale variations in the data. Mathematically, this cannot happen, thus the only reason for C_0 is measurement error (Cme). Therefore, the fitted model does not pass through the origin, but intersects the y-axis at a positive value of gamma, labelled C_0. The nugget is that part of the variance

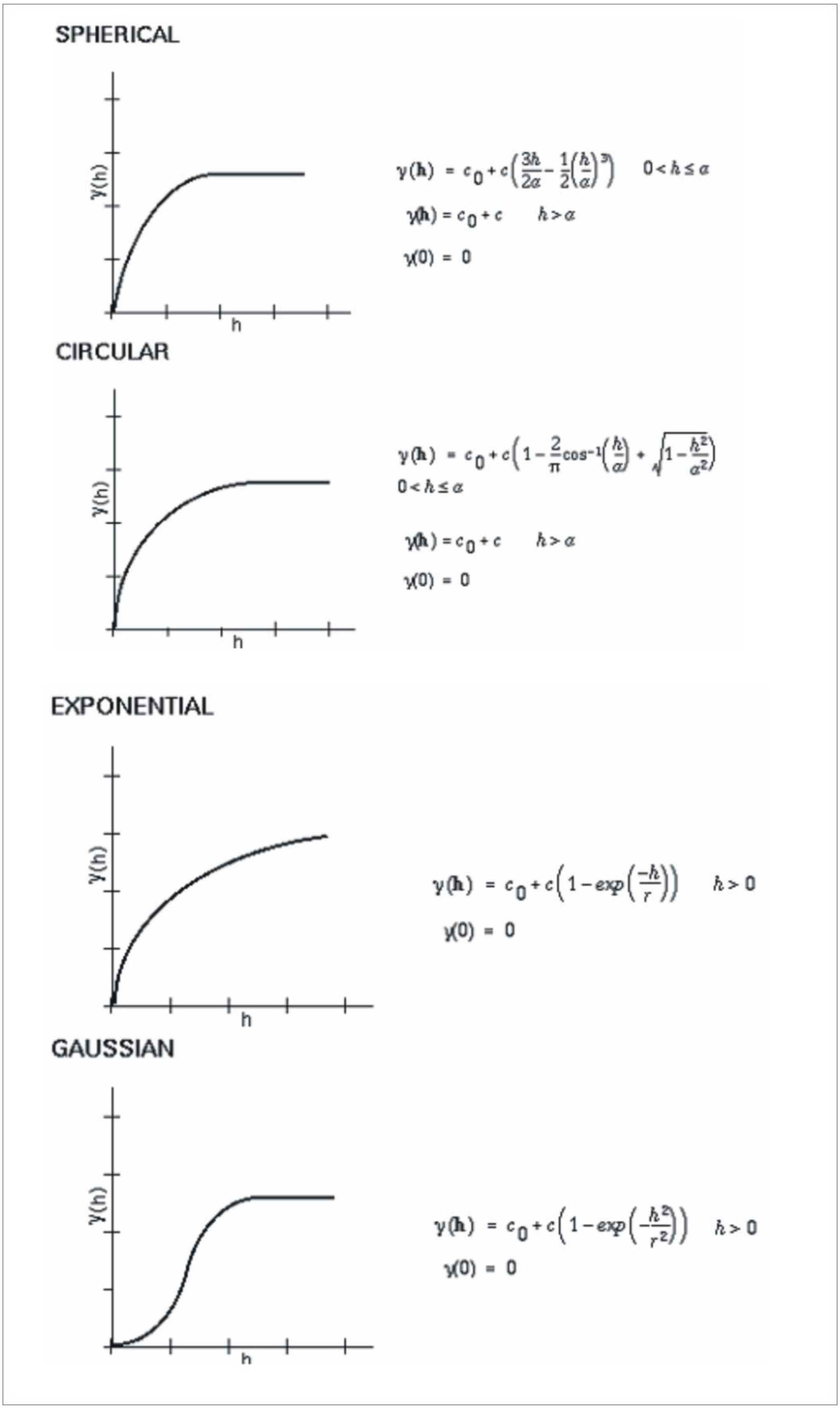


FIGURE 4.10: Theoretical models (taken from ArcGIS Help ©ESRI).

of a regionalised variable that has no spatial component (variation due to measurement errors and short-range spatial variation at distances within the smallest inter-sample spacing). Kriging has been shown to be one of the most reliable two-dimensional spatial estimators (cf. Laslett *et al.*, 1987; Laslett and McBratney, 1990; Laslett, 1994) and it is expected to produce more reliable estimates of elevation data than *ad hoc* methods. Kriging was chosen as the preferred method of spatial interpolation, firstly, because it assigns weights based not only on the distance between the surrounding points but also on the spatial autocorrelation among the measured points, which is determined by modelling the variability between points as a function of separation distance (EPA, 2004). Secondly, it provides an estimation variance[3] which gives an indication of the reliability of the predictions associated with the model (Chappell *et al*., 2003).

4.6.1.5 DEM generation

Using Gstat (Pebsema and Wesseling, 1998), the data was explored spatially and an empirical model fitted to the semivariogram to determine which type of spatial prediction was required to produce the best representation of the landscape. Given the anisotrophic nature of the data, there was a trend to the data set: drift is common in elevation data (cf. Chappell 1995; Chappell *et al.* 1996, 2003). The spatial variation in elevation of the Tyne Valley is distinctly anisotrophic; the degree of elevation variation across the valley is very different from that along the valley, therefore the data required detrending. The form of the drift (trend) was estimated by fitting low-order (linear and quadratic) polynomials on the spatial co-ordinates of the elevation data using least squares regression. A semivariogram model was fitted to the empirical semivariogram,

[3] Estimation variance is the sum of the cross-products of the weights assigned to the sample points and each cross-point is weighted by the covariance between the corresponding points (Wackernagle, 2003, p.30).

using the nugget effect, spherical component and range determined by eye from the empirical semivariogram. Universal kriging (UK) and weighted least squares were used to fit the model and generate the bedrock surfaces.

The dGPS DEM was also generated using Gstat. There was bias in the sample data, which resulted in an uneven grid. Therefore, the grid interval for the interpolation was generated from a systematic 10 per cent subsample of the original data set. A semivariogram was estimated for the 10 per cent (systematic) subsample a second order trend (quadratic) was applied to the data set to account for the anisotropy and a nugget effect and spherical component were fitted to the semivariogram by eye. The universal kriging (UK) functionality of Gstat was used and weighted least squares were used to fit the model and the grid was exported to ArcGIS to view the DEM.

4.6.2 Sediment budgeting

Sediment budgeting determined the approximate volumes of sediment stored, eroded and transported within the Tyne Valley since deglaciation. Using a morphometric approach (after Huisink, 1999 and Passmore and Macklin, 2001), sediment budgets were calculated for each respective phase of terrace development. Volume of stored sediment for each terrace was calculated using equation 4.7:

$$\textit{Surface area } (m^2) \times \textit{Thickness of deposit } (m) = \textit{Volume } (m^3) \qquad \text{[equ. 4.7]}$$

To estimate the total surface area for each terrace, terrace fragment areas were estimated (areas lost through erosion cannot be accounted for) and summed with all younger terrace areas and the modern channel area[5] (as per Passmore and Macklin, 2001). Terrace fragment areas were identified and delimited by field mapping and

[5] Present channel area was calculated by multiplying average channel width by its length

DEM visualisation. These were then digitised by hand as polygons in ArcMap, and the surface area (m^2) calculated using the 'calculate area' functionality in Arc Editor, exported as a .txt file into an Excel spreadsheet.

The thickness of the sediment body associated with each terrace (i.e. glacial sand and gravel, till and fluvial gravels) was determined from the IMAU borehole log data[6] (Giles 1981; Lovell 1981). The logs give the elevation of each sedimentological unit (e.g. alluvium or glacigenic sand and gravel). At each borehole, the thickness of a sedimentological unit was inferred by subtracting the upper and lower elevation of their characteristic sediments. For this work, two separate sediment thickness were measured: glacigenic sand and gravel, and alluvium. Where till was interbedded with glacigenic sand and gravel, it was not differentiated as a separate lithological unit but included in the glacigenic unit thickness. Without exposure to assess if the till represented a re-advance of ice in the valley, and with little evidence for such an interpretation, it was classed as part of the outwash sequence. Only where till directly overlay bedrock was it interpreted as basal till.

A degree of uncertainty arises from the natural undulation of the interface between sediment types. Huisink (1999) suggests this variation is best managed by using averaged sediment thickness in calculations, and this method was applied here. However, there was a second source of uncertainty in the boreholes from the Tyne Valley. In most cases, the lower elevation of the sedimentological unit was clearly demarcated within the log. However, in a few cases, the borehole log was terminated "due to obstruction or slow progress" before the lower elevation was reached (Giles

[6] There are fewer boreholes than might be hoped for statistical robustness.

1981; Lovell 1981). For this work, the lower elevation of the sedimentological unit from these borehole logs was assumed to be the elevation at termination of the boring. Finally, for ease of calculation, it was assumed that the maximal aggradation of the terrace across the entire valley floor was uniform. This height is assumed to be now represented by the current terrace surface elevation above present river level.

Incision was calculated as the height difference between the terraces plus the maximum depth of the infill of the lower terrace. As described earlier, the area of each successive floodplain is calculated by summing the area of the terrace fragments, all the lower terrace fragments and the area of the present channel. Area lost between each phase of terrace development was calculated as the area of the earlier floodplain minus the subsequent floodplain area.

The following equations (equ. 4.8-4.12) were used to estimate sediment lost (denudation rate) during phases of valley floor reworking. The size of the catchment area used in these calculations was assumed to correspond to the present day drainage basin area (~2.9km^2). This was assumed because sediment units (glacigenic and alluvium) outcrop throughout the study area thus implying the catchment was ice-free at the time of their deposition. Volume reworked between each terrace phase was calculated using equation 4.8:

$$\textit{Area } (m^2) \times (\textit{Thickness of deposit } (m) + \textit{Height difference between terraces } (m)) = \textit{Volume reworked } (m^3)$$

[equ. 4.8]

Volume of sediment lost from the terrace was calculated using equation 4.9:

$$\textit{Area } (m^2) \times \textit{Thickness of deposit } (m) = \textit{Volume lost } (m^3)$$

[equ. 4.9]

Volume lost was converted to a volume loss per square metre of catchment using equation 4.10:

$$\textit{Volume lost } (m^3) \textit{ / Catchment area}^{7} \ (m^2) = \textit{Loss } (m^3/\,m^2)$$

[equ. 4.10]

The volume lost per square metre was converted to metric tonnes using equation 4.11:

$$\textit{Loss } (m^3/\,m^2) \textit{ x Bulk density}^{8} \ (t/\,m^3) = \textit{Loss } (t/\,m^3)$$

[equ. 4.11]

Annual sediment export was calculated using equation 4.12:

$$\textit{Loss } (t/\,m^3) \textit{ / Time (yr) = Annual sediment export } (t/\,m^3/yr)$$

[equ. 4.12]

4.6.2.1 Estimate of potential errors

There are a large number of potential errors associated with calculating the sediment budgets. Spatial variability in thickness could not be accounted for because of a deficit of boreholes. As discussed, estimates of sediment unit thickness at terminated boreholes are likely to be an underestimation of the actual thickness at these points, an underestimation compounded in calculation of total terrace volume. Estimates of bulk density were calculated for various combinations of sand and gravel proportions; they are generic values and are not based on sediment packing in the field. Sediment area was calculated based on polygons delimited in ArcMap. There is a subjective element in identifying and delimiting terraces in the field and DEM visualisation, and limits to accuracy in digitising these outlines as polygons in ArcMap. This error cannot be quantified. Finally, there were uncertainties in the terrace ages and therefore the time elapsed between each development phase could not be absolutely determined. Estimates of time used in the calculations were based on a poorly defined chronology.

[7] The catchment area for which the sediment volumes have been calculated was assumed to correspond to the present day drainage basin
[8] Bulk density varied dependant on type of sediment infill and packing situation assumed

Chapter Five

Quaternary Landforms and Sediments

5.1 Introduction

In order to understand the development of the Tyne Valley during and following deglaciation, one of the aims of the thesis was to map the landform-sediment assemblages along the Tyne Valley using traditional field mapping and new, innovative DEM visualisation techniques. The field mapping provided detailed information about the type of features (landforms) at each of the key study areas, whereas the DEM visualisations provided feature identification between, and beyond, the study areas allowing the full suite of depositional features to be mapped and allowing ideas about landscape development formulated. Visualisations, field maps and original sketches are stored on CD located at the back of the thesis in the map pocket.

Thomas (1989) defined a sediment-landform assemblage as a mappable unit, on a scale of 1:50000, in which relatively homogeneous morphological, stratigraphical and lithological characteristics occur. The types of sediment- landform assemblages, which were identified, represent glacial (ice proximal), deglacial and postglacial processes/environments/conditions. On the basis of detailed field survey and mapping, landforms in the study area (Figure 3.5) were categorised into glacial, proglacial and postglacial features. They comprise mound and terrace landforms underlain by till, glaciofluvial and glaciolacustrine sands, and alluvial sand and gravel deposits.

With a view to better understanding the behaviour and development of the Tyne Valley since deglaciation, this chapter comprises a detailed explanation of the sedimentology, the chronological framework, a review of the distribution of landforms, and a calculation of sediment storage and reworking.

5.2 Sedimentology

The following section provides a detailed description of sedimentary sequences at each of the study locations, and includes a review of the borehole log data where available, and a reconstruction and interpretation of the depositional environment.

5.2.1 Crawcrook

The sedimentary sequence: Location 1 (working quarry)

A single, major, north–south-trending face was studied, comprising a sequence of sediments up to 8m thick, divided into two major, spatially extensive units (1 and 2) (Figures 5.1; 5.2). In addition to this major face, a number of smaller exposures were observed and a more localised diamicton unit (3) was recorded (see Appendix 2). The basal unit (1) comprised a tripartite division of sediments between 1.8m and 2.3m thick. At the base, the sediment generally comprised horizontally stratified, well-sorted medium sand (1.4% fine gravel, 97.2% medium sand and 1.4% silt) and exhibited some evidence of climbing ripple lamination (Sh, Sr) and occasional pebble stringers. This graded upwards, conformably, into a sandy silt (48.1% fine sand and 51.9% silt) with wavy laminations (0.05m thick), which in turn was replaced by a (0.15–0.25m thick) blanket of clayey sand (Fl). This tripartite sequence was repeated at least three times in the recorded exposures and suggests some degree of cyclicity to sedimentation of this particular unit. No erosional contacts were observed or any incorporation of rip-up clasts in the basal layers of sands in each successive bed-set, suggesting that the units were not deposited under high-energy flow conditions. In the central part

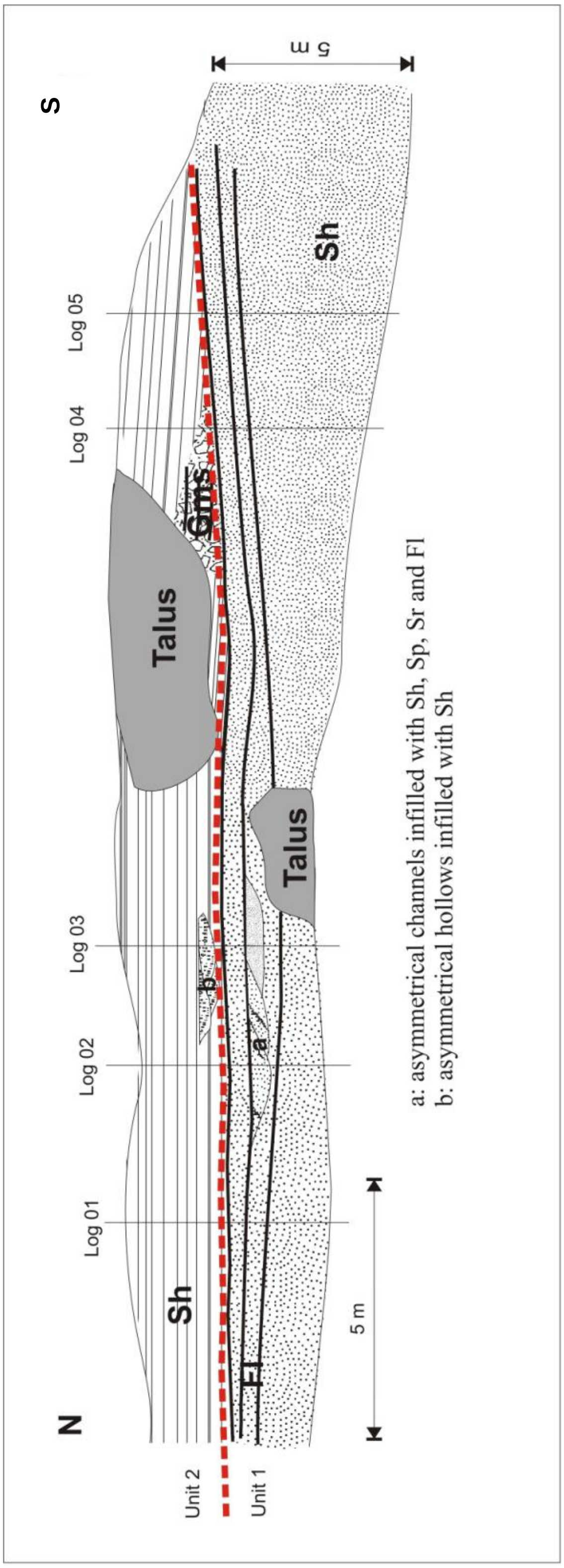


FIGURE 5.1: Generalised sketch and photograph of the face at Crawcrook, location 1 indicating the two lithofacies units. The locations of the vertical logs are denoted. Unit 1 comprises Sh and Fl facies, and unit 2 comprises Sh and Gms facies.

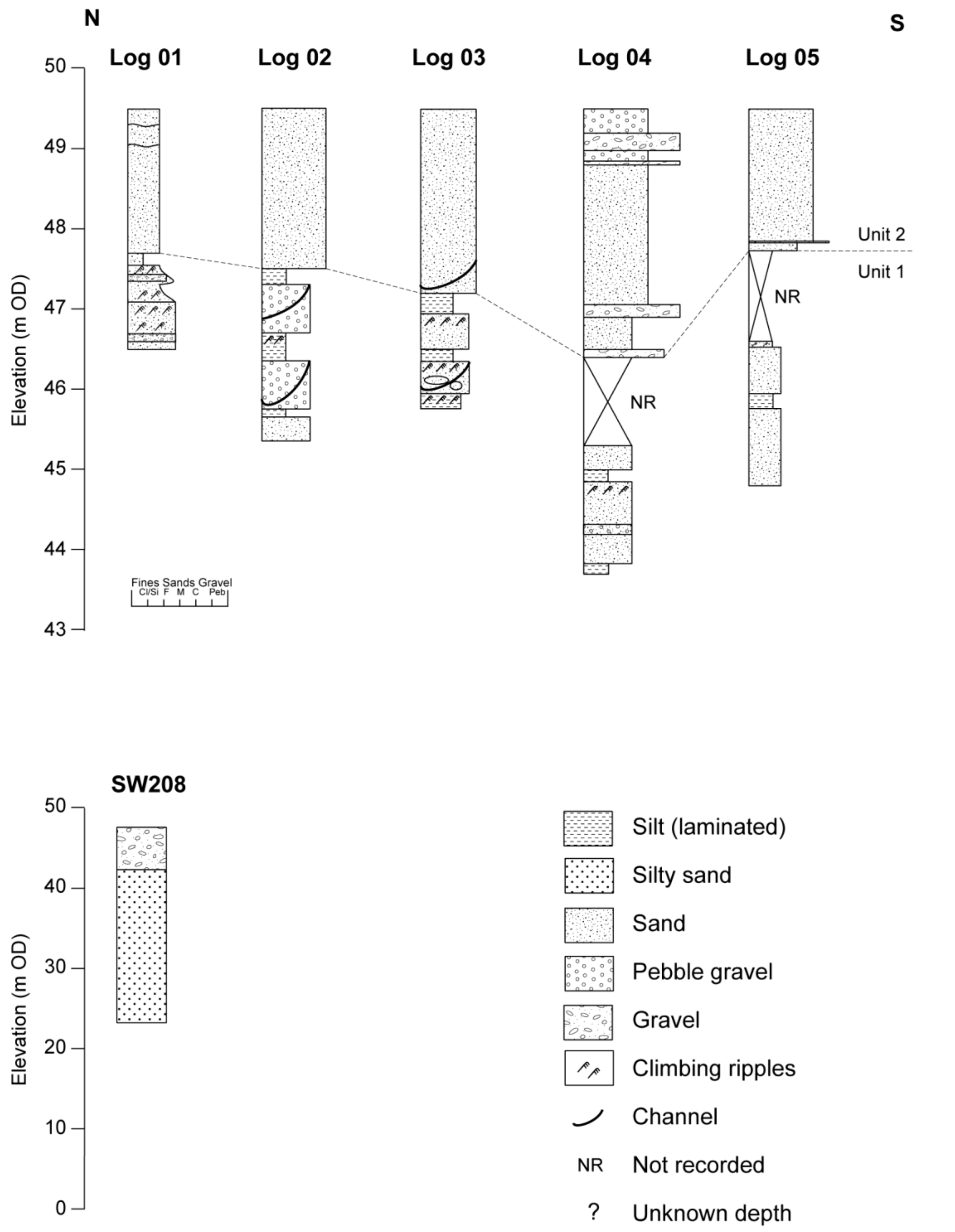


FIGURE 5.2: Stratigraphic logs recorded at Crawcrook, location 1 and the IMAU borehole log (SW208, see Figure 5.29 for location), which was recorded nearby. The sequence is dominated by cyclic units.

of the quarry exposures, unit 1 contains a large, asymmetrical channel feature (*c*.3.5m wide by 0.5m deep), which was infilled with a combination of planar cross-bedded and horizontally bedded sands (0.8% fine gravel, 98% medium sand and 1.3% silt) (Sp, Sh), climbing ripple laminations (Sr) and silty sand drapes (Fl). Elsewhere in the quarry, other channels within unit 1 were infilled with both clast and matrix-supported, poorly sorted, horizontally stratified, medium gravels and sands (55.3% medium gravel, 44.4% medium sand and 0.3% silt) (Gms, Sh). A single V-shaped structure similar to an ice-wedge cast (cf. Worsley, 1996), less than 0.5 m in length and width, was observed in unit 1 (Figure 5.3A). No other periglacial evidence, such as cryoturbation or frost shattering of pebbles was observed. Post-depositional structures recorded in unit 1 included both small-scale, high-angle normal faults with throws between a few millimetres and as much as 5cm and large-scale normal faults with throws up to 0.5m (Figure 5.3B). Palaeocurrent analysis from unit 1 suggests that the sediments were deposited by water issuing from the southeast.

The overlying unit (2) comprised a horizontally stratified, medium-grained, moderately sorted sand (0.2% fine gravel, 98.9% medium sand and 0.9% silt) (unit 2, Sh). This unit varied between 1.8m and 2.2m thick and rested conformably on unit 1. The sands formed couplets of alternating medium- and coarse-grained material (50mm thick). Some couplets were indurated, and occasional silt drapes and abundant small flecks of coal were observed along individual lamination planes, the latter tending to increase in abundance towards the unit base. Smaller, asymmetrical hollows (*sensu* Marren, 2001) infilled with horizontally bedded, medium-grained sands were recorded incised into the base of this unit. Elsewhere in the quarry, a discrete unit of diamicton (4.1% fine gravel, 50.2% coarse sand and 45.7% silt) (dm) between 0.10m and 0.15m thick, but

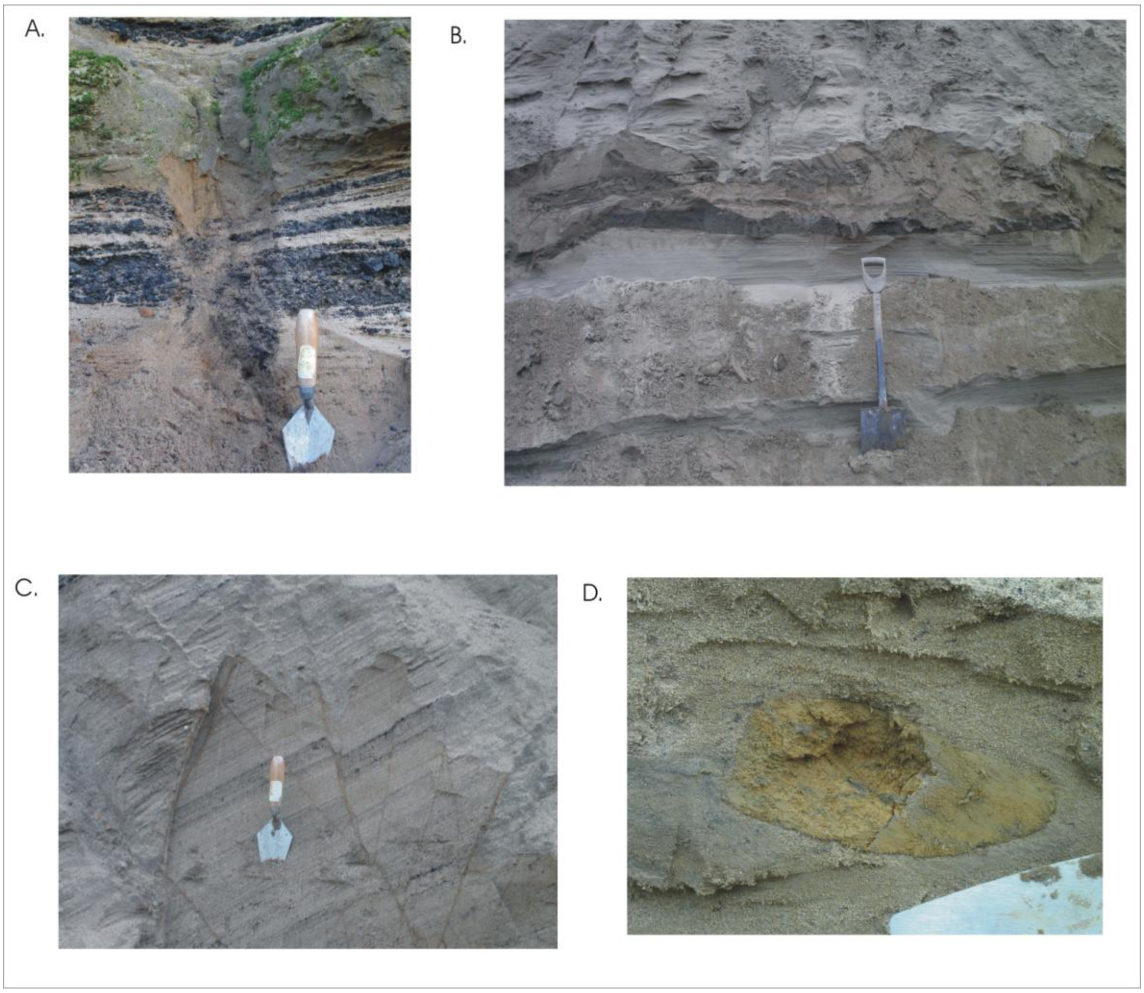


FIGURE 5.3: Interesting features identified at Crawcrook location 1. A. V-shaped feature. B. Diamicton raft. C. Small-scale faults. D. Soft sediment clast. For scale, trowel is 28cm in length and spade is c.1m in length.

less than 5m in lateral extent, was observed (unit 3; Figure 5.3C). It was recorded within horizontally bedded medium- to fine grained, moderately well-sorted sands, similar to those described for unit 1. The contact between the diamicton and the surrounding sediment was sharp. Thin (20–100 mm thick) laterally discontinuous stringers (dm) of the unit interdigitated with the overlying sands, suggesting erosion and reworking of the diamicton. Within the surrounding sands, an isolated clast of silty sand was observed, with no deformation of the surrounding sediment (Figure 5.3D).

Interpretation

The cyclical repetition of fining upward sequences comprising unit 1 (Sh, Sr, F1) suggests repeated periods of increasing and declining flow conditions. Given the thickness of these deposits with cm-scale laminations, this transition may be indicative of seasonal or fluctuating diurnal discharges. The relatively well-sorted, medium-grained, horizontally bedded sands (Sh) imply high-energy currents, while the presence of pebble stringers suggests the rolling of clasts along the channel bed. The transition from climbing ripples (Sr) to silty sand drapes (Fl) indicates decelerating flow conditions. The low silt and clay content (<2%) of these sediments suggests rapid settling and accumulation, a characteristic response to variations in meltwater discharge (Jopling and Walker 1968; Allen 1982; Ashley *et al*. 1982), or possibly low content of silt-clay sized material. The Fl sediments represent deposition under the lowest energy conditions at the end of the cycle and accumulation by suspension fall out (Fraser and Cobb 1982; Marren 2002). Within unit 1, channel deposits comprising clast- and matrix-supported gravels and sands suggest rapid aggradation under high-energy conditions followed by low-stage winnowing of fine sediments (Marren 2002; Tucker 2003). Marren (2001) described similar depositional units in proglacial fluvial sequences in Scotland, which he attributed to shallow, high velocity, variable flows. Locally, pebbles (Gms) overlie the horizontally bedded sands (Sh) and are conformably overlain by the silty fine sand drapes (Fl), giving the appearance of silts interbedded with dispersed pebbles. This suggests very rapid changes in energy conditions, combined with access to coarser sediments. These silty pebble units are similar to glaciolacustrine deposits recorded by Marren (2002) in a modern proglacial deltaic environment. However, bottomsets, foresets or topsets associated with deltaic sequences (cf. Brown, 1994; Aitken 1998) have not been recorded at this site. Furthermore, although glaciolacustrine sequences (cf. Ward and Rutter, 2000) are characterized by units of sand with thick silt and clay drapes, there are no obvious lateral ice contact structures or ice-rafted debris (i.e. dropstones) to suggest glaciolacustrine deposition. Therefore, unit 1 is interpreted as a glaciofluvial rather than a glaciolacustrine deposit. The alternating medium- and coarse-grained sand couplets (Sh) of unit 2 and the presence of localized scours suggest rapid deposition under high energy, sheet-flood conditions (Marren, 2002). The textural variations can be related to water-stage changes, whilst the scours are interpreted as small channels infilled with migrating bedforms, suggesting localized fluctuations in stream power (Shaw, 1972).

The sequence exposed at location 1, Crawcrook is interpreted as a glaciofluvial sequence and represents deposition as part of a distal, sandy ephemeral stream. The sequence is dominated by horizontally bedded sands (Sh), characteristic of sheet flow, and finer-grained ripples and laminated silts, characteristic of waning flood deposits. Miall's (1978) Bijou Creek facies assemblage, which recognized repeated flood events and the cyclic nature of sedimentation, is the most appropriate model for the sequence. The diamicton of unit 3 (dm) is suggested to represent a large raft of eroded sediment incorporated within the outwash plain. The sharp contact between the diamicton block and the horizontally bedded sands (Sh), and the undeformed nature of the underlying sands suggests both were frozen when emplaced (cf. Krainer and Poscher, 1990). Inclusion of diamicton blocks suggests proximity to ice, and they are possibly ice-rafted. However, there is no other evidence for lacustrine deposition within this sequence. Thus, the frozen sediments probably represent undercutting of channel banks formed in glacigenic sediments during peak flows; the laminated sands probably represent upper flow beds (Tucker, 2003). The stringers of diamicton and soft sediment clast within the sands clearly indicate subsequent erosion and reworking of glacigenic deposits. The tentative identification of a single ice-wedge cast (might indicate subaerial exposure under periglacial conditions with the development of permafrost indicating mean annual air temperature *c*. -6 to -8˚C or colder during parts of the Devensian (cf. Murton and Kolstrup, 2003). However, as Worsley (1996) cautions, care should be taken when identifying and interpreting features as ice-wedges. As the suggested ice-wedge cast does not truncate the sedimentary sequence, it is clear the structure is not a post-depositional feature, suggesting the presence of syn-depositional permafrost. Seddon and Holyoak's (1985) criteria for inferring syn-depositional permafrost cannot be identified at this site, and the criteria outlined by Worsley (1996) for identification of ice-wedge casts cannot be met either. This suggests the feature is not an ice-wedge cast, and was interpreted as a fault displacement. The faults in unit 1 are similar to those described by a number of workers from glaciofluvial deposits in the UK (e.g. Thomas and Montague 1997; Huddart 1999; McCarroll and Rijsdijk 2003). They attributed faulting to the slow and gradual melting or collapse of stagnant and dead ice beneath the sediments (cf. McDonald and Shilts, 1975). It is notable that the deposits at Crawcrook are composed almost entirely of sands. The absence of a rudaceous component might suggest that the sediments were deposited distally to an ice-margin.

The sedimentary sequence: Location 2 (former quarry)

Figure 5.4 shows the type of sedimentary sequences investigated at location 2 (the former quarry) at Crawcrook. Location 2 is located in a mound parallel to location 1 within the Crawcrook complex, *c*. 50m in length and a single major face was examined. The face comprised a north-south and west-east exposure, although this was degraded as the quarry is no longer worked for sediment. The sequence comprises a single unit incised by three subunits with up to 20m depth of sediment exposed (see Appendix 2). Unit 1 consists of sand and is the most laterally extensive unit exposed. In general, this fines upwards from coarse sand to silty fine sand. The sand is thinly laminated (sub

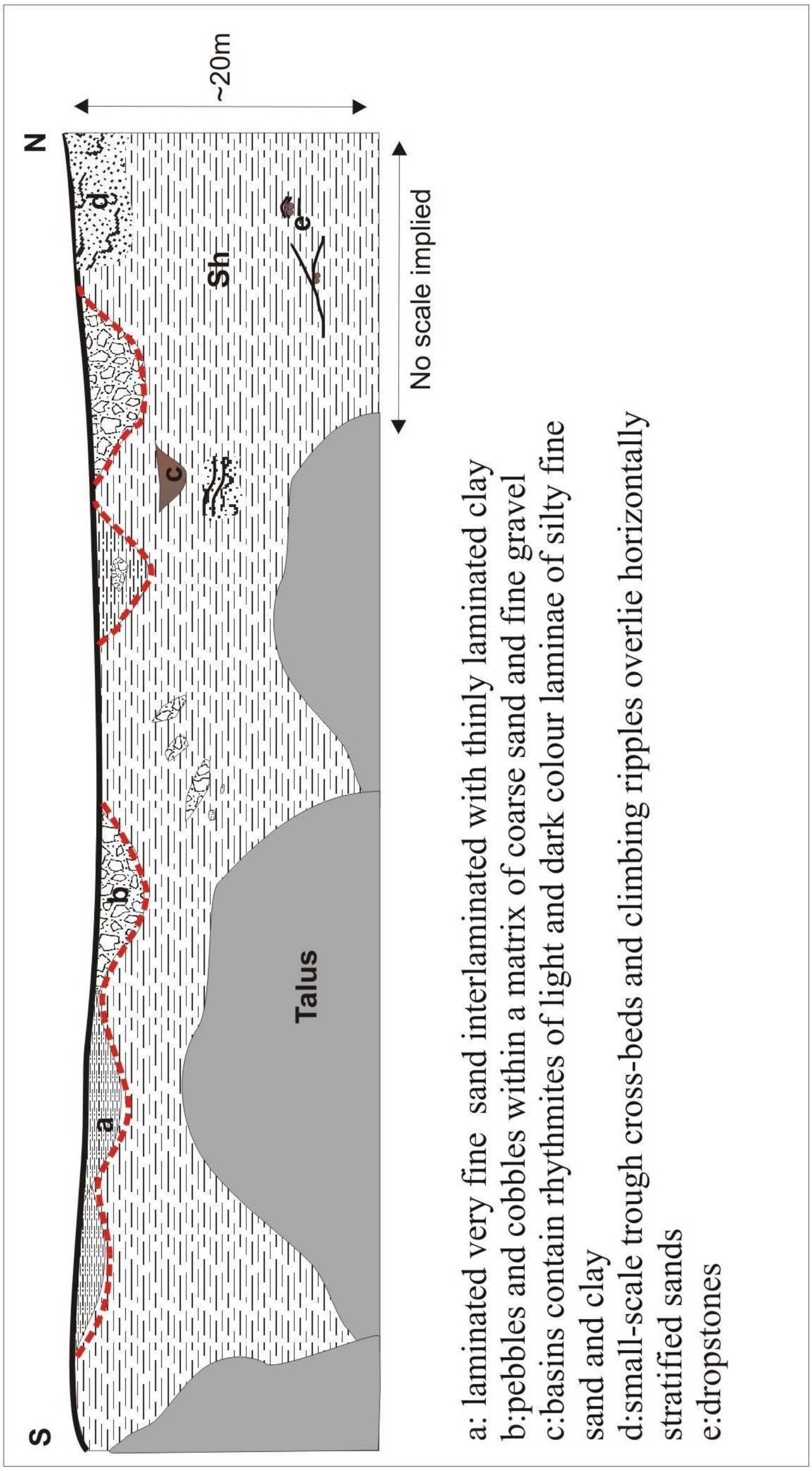


FIGURE 5. 4: Generalised sketch of location 2 illustrating the location of channels (Gms facies) and basins (Fl, Sh facies) at the top of the sequence.

cm) (Sh) and is draped with beds (20-30cm thick) of silty fine sands (Fl). Towards the base of the exposure the sand is draped with beds of very thinly laminated red clays between 20-40cm thick (Fl). These subunits are disturbed, locally discrete and associated with large boulder dropstones. Towards the top of the sequence sets of small-scale trough cross-beds and climbing ripples (Sr, St, Sp) overlie the horizontally stratified sands (Sh) and are in turn draped by laminae of silty fine sand or clay-rich silts (Fl). The uppermost sets of ripples have a more pronounced wavy form, dipping at 10˚, and are probably clinoforms. Within the sequence a range of coarse gravel sediments occur. These are laterally discrete and highly localised units, which show abrupt vertical and lateral variations in texture. They comprise massive cobble and pebble gravel supported in a matrix of coarse sand and granules (Gms) and are intercalated with horizontally stratified silty fine sands. Francis (1975) reported a similar sequence when he investigated an earlier exposure in the mounded complex at Crawcrook, which he interpreted to be eskers.

Incised into Unit 1 are a series of small basins and channels. The basins are *c*. 4m in depth and *c*. 12m wide (Figure 5.4). The basins comprises thin beds (sub cm - 4cm thick) of laminated (single grain - 1mm thick), normally graded very fine to fine sand interlaminated with thinly laminated clay (red) beds. Some of the beds are intricately folded and faulted and there is evidence of soft-sediment deformation (flame structures) (Figure 5.5A, B). The sequence is disturbed by dropstones in the form of gravel pods. The gravel is moderately stratified, pebble gravel with a matrix of coarse sand and granules (Figure 5.5C, D). Other basins contain rhythmites (sub cm - 4cm thick) of light and dark colour laminae of silty fine sand and clay. The basins have a sharp contact with the sands and the basal contact is a clay-rich bed *c*. 10cm thick. The channels are *c*. 2m in depth and *c*. 6m wide. The channel sequence comprises a single pebble gravel unit. The unit sharply overlies a coarsening upward sequence of fine to medium sand with occasional outsized pebble lags. The gravel is stratified, consisting of pebbles and cobbles with a matrix of coarse sand and fine gravel (Gms). The cobbles show evidence of frost shattering.

Finally, there is a locally discrete sequence, which truncates Unit 1. The sequence comprises two units: the upper unit comprises cobble and boulder gravel with a matrix of fine gravel and silty fine sand (Gms). The pebbles show evidence of frost shattering. The lower unit comprises lenses of pebbles arranged in clusters, they show weak imbrication with smaller pebbles trapped behind larger ones, with a matrix of fine gravel (Gms), within stratified, very fine to silty fine sands (Sh).

Interpretation

Unit 1 consists of fine-grained, horizontally laminated sands with interlaminated layers of silt and clay that become ripple laminated, and are arranged as climbing ripples towards the top of the sequence that could be associated with both fluvial or lacustrine depositional environments. However, the presence of occasional dropstones set in the fine-grained sediments is the clearest evidence of a lacustrine setting, where ice-rafting and subsequent dumping of material occurs as the ice-bergs float across the water surface (cf. Bennett and Glasser, 1996). The concentration of bergs may have been low, given that no evidence of diamicton deposition was found. Unit 1 could be interpreted as a distal lake sequence, and deposition certainly took place from tractive bedload and suspension settling from low-concentration turbidity currents flowing as quasi-continuous currents within the lake (cf. Jopling and Walker 1968; Lowe 1982; Allen 1985). The presence of climbing ripples towards the top of the sequence implies decelerating flow (cf. Ashley, 1988) resulting from turbidity currents, and the sandy

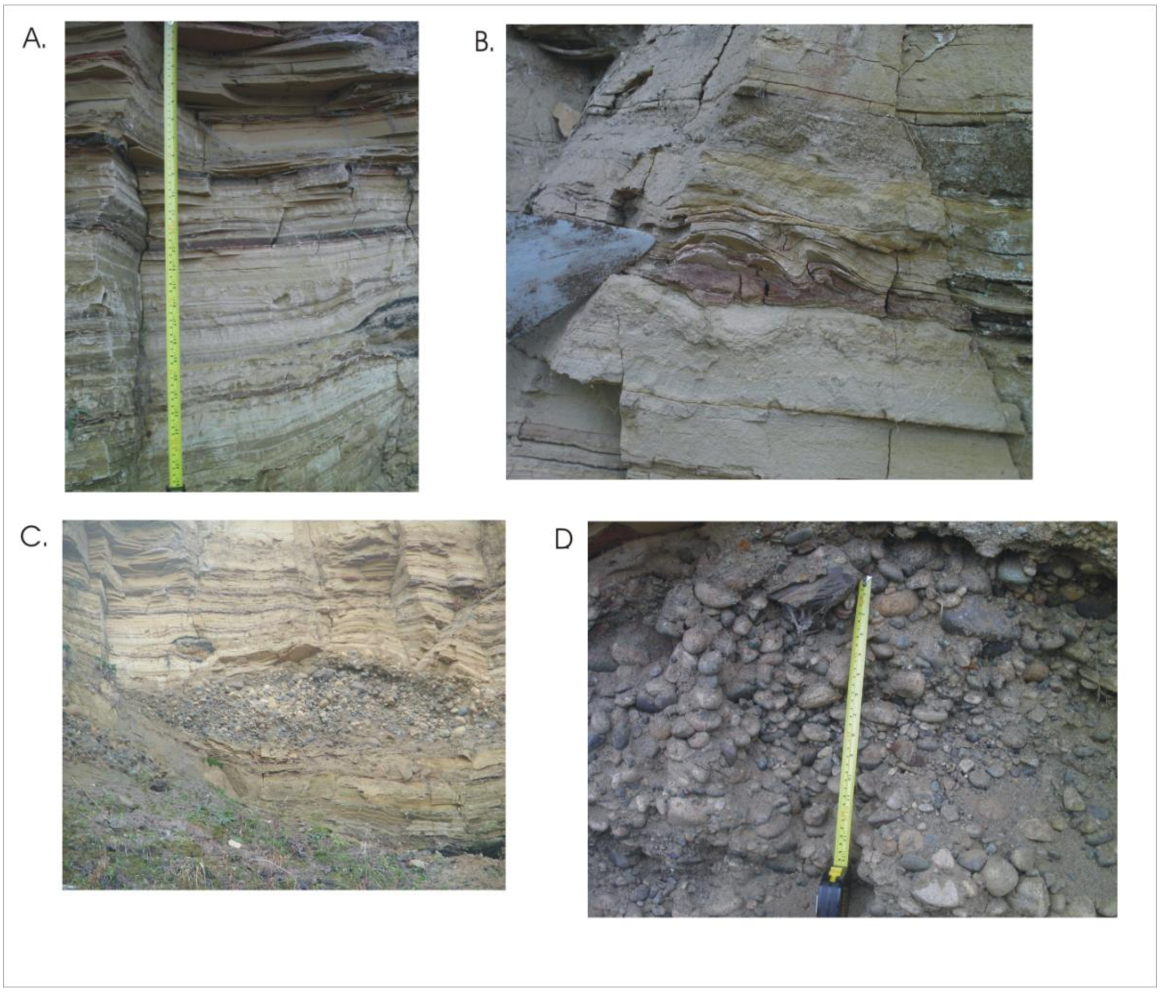


FIGURE 5.5: A. Laminated sediments of the pond, rule is c.70cm in length. B. Soft sediment deformation within the laminated layers of the pond. C. Gravel dropstone or underflow deposit within the pond; laminations are conformable above the gravel. D. Rounded gravels comprising the gravel deposit, rule is c.35cm in length.

clinoforms at the very top of the sequence have been interpreted as foreset beds deposited on a gentle prograding slope. The borehole logs do not indicate laterally extensive or thick gravel deposits in the vicinity of this sequence, which would be associated with deltaic foresets (cf. Gilbert, 1890), therefore suggesting that the development of a delta was inhibited through the continued deposition from underflows (Bennett and Glasser, 1996).

There is evidence of sediment slumping within the lake, possibly related to re-distribution of sediments by gravity deformation. The sequence contained a discrete layer of silty clay, which was clearly disturbed (i.e. thrust upwards and bifurcated) although the laminae were not distorted. This anomalous occurrence was the only exposed evidence of subaqueous slumping within the sequence, and is interpreted as the result of upwards thrust due to the accumulating weight of sediments above the unit (cf. Fulton and Pullen, 1969) rather than a down slope re-distribution of sediment.

The discrete coarse gravel units reported here have been interpreted as sediment-laden surges into the lake as coarse material become available. They occur in close proximity to the channel fills which have been incised at the top of the sequence and probably represent subaqueous channel deposits. Francis (1975) reported extensive gravel cored units within the Crawcrook sequence. These may represent lake floor esker ridges (cf. Gustavson, 1975) or bottom flow surge current deposits that became flanked by lake bottom sediments.

The localised sequence which truncates Unit 1 is interpreted as a high-concentration debris flow deposit. The two units that constitute the sequence may represent the flow separation that occurred as the debris flow entered the water body (cf. Middleton and Hampton, 1973). Benn (1996) reported a similar facies from a subaqueous sequence in a former lake in Scotland. The stratified laminated sands at the base probably represent deposition of the tractive bedload, where the concentration was low at the base of the flow (cf. Lowe, 1982). The outsized clasts, which appear floating in the sequence, the absence of grading and the discrete flow structures, are all indicative of cohesionless flow where particle movement is strongly influence by the surrounding sediments (cf. Hein 1982; Nemec and Steel 1984; Postma *et al*. 1988).

As the sediment infill of the lake was reaching the water surface level, a series of channels and small lake basins were incised into unit 1. One of the infilled basins is associated with a lens-shaped accumulation of pebble gravel, interpreted as a bottom flow deposit or possibly a large dropstone. However, there are small dropstones within the sequence that indicate that the basin was still ice-proximal and subaqueous. These infilled basins have been interpreted as ponds, given they comprise thinly stratified layers of alternating sequences of silt-rich and sand-rich sediments which posses the rudiments of rhythmites, occasional dropstones, small-scale deformation structures and contorted layers of clay-rich sediments. Rhythmites are typical of rapid deposition and can accumulate in a matter of hours or days (Smith and Ashley, 1985) due to settling from suspension.

Unit 1 is interpreted as a subaqueous fan deposit, formed as a subglacial tunnel entered the water body. Deposition took place from underflow turbidity currents. Sediment delivery was continuous and discharges were sufficiently competent to transport the sediments away from the ice margin, thus preventing the development of a large delta. As the fan built up towards the surface water level, channels were cut across the surface of the deposit. These are interpreted as a system of feeder channels delivering sediment to the fan. Frost shattering occurred after the sediments were subaerially exposed. Deposition from debris flows suggests local slumps from the ice margin. The IMAU boreholes taken from the area around the Crawcrook quarry delimit the extent of the lacustrine sequence (Figure 5.6). Where till is encountered, it is indicative of ice cover, serving as a proxy for delimiting the boundaries of the lake. A palaeocurrent fabric could not be obtained from the lacustrine sequence because the face was degraded and access was difficult to the only cross-bedded structures at the top of the sequence. Thus it is unclear whether the lake was draining the main Tyne Valley (west-east), or

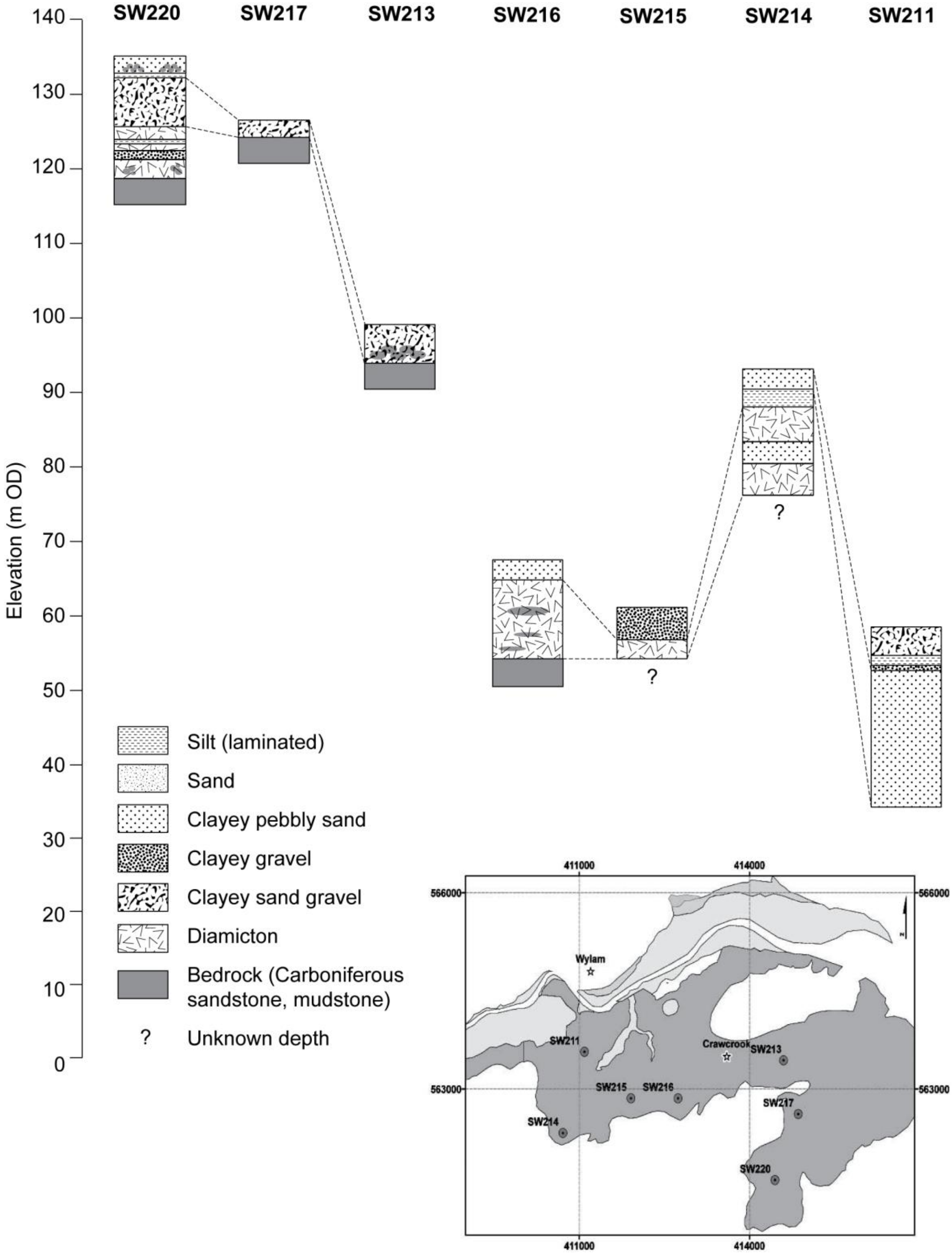


FIGURE 5.6: The stratigraphic relationship of the IMAU borehole logs in the vicinity of the Crawcrook exposures is illustrated. The sedimentary units recorded in the boreholes can be correlated over a wide area. The sequence comprises clayey gravel, clayey pebbly sand and silt which thicken towards the valley centre (up to 25m thick), only forming a veneer on the valley sides. The cover is not ubiquitous and the upper valley sides are mantled in diamicton. Borehole location map inset.

was a localised lake draining ice issuing from the North Pennines (north-east). However, given its limited extent, the lake (hereafter named Glacial Lake Bradley) probably developed through localised ponding of meltwaters in the lower valley.

The sedimentary sequence: Location 3

Locally, in a cutting through the mounds, a discrete unit of diamicton was recorded overlying sands, which are adjacent to location 1 (Figures 5.7). The diamicton exposed was, approximately, <5m in length and *c*. 1-1.5m in height. The lower contact with the sands was sharp but gently undulating, and in the lower 20-30cm it was clear that sand lenses had been incorporated and there was evidence of soft-sediment deformation. The diamicton was massive and stiff, with concentration of clasts, consisting mostly of small pebbles, around 20-25%. The underlying sands were sedimentologically the same as those recorded at location 1, and are considered to be part of the same sandur sequence.

Interpretation

The diamicton could represent a number of depositional environments, such as sub-glacial, pro-glacial, and glaciolacustrine. Smith (1994) recorded a diamicton overlying outwash deposits in the Wear Valley, which he attributed to periglacial reworking through solifluction and slope wash. Thus, the presence of diamicton within the sediment sequence is not necessarily indicative of englacial deposition as previously suggested by Mills and Holliday (1998). However, the diamicton here is massive, dense, and has a consolidated structure, suggesting it was not a supraglacial melt-out or debris flow till. Levson and Rutter (1988) reported gradational lower contacts in basal meltout tills and it is possible this diamict unit formed from both lodgement and meltout processes. No fabric or lithological data was obtained from the diamict unit,

FIGURE 5.7: A. Location 3 indicating diamicton overlying sands. For scale, figure is 1.80m in height. B. Detailed view of boundary between diamicton and sands, showing incorporation of sands into the basal diamicton (blue ellipse). Spade is *c*. 1m in length.

which would have provided an indication of its origin, due to poor exposure and time Mills, 1991) and striated clasts are indicative of extensive glacial transport. Lithological analyses could have helped differentiate whether this diamicton is related to movement of the coastal ice sheet, therefore indicating its possible landward extent, or if it simply reflects a local movement of ice.

The origin of the diamicton may remain undetermined, but it is clear that following aggradation of the extensive sandur plain, ice was present and retained forward movement. The presence of ice cover within the locality has already been established from the borehole logs which record till overlying bedrock. However, this deposit represents a possible late stage incursion of ice, where the margin was probably thin and oscillating and gently bumping and pushing along, overriding the sandur sequence. Thus, the diamicton has been interpreted as a subglacial (lodgement) till.

5.2.2 Stocksfield

The sedimentary sequence: Location 1

Figure 5.8 shows the stratigraphic logs of the Bullion Hills section at Stocksfield. The section comprises a single unit divided into layers between 0.2 and 1m thick of sand, gravel and sandy diamicton (Figure 5.8B). Generally, the gravels, up to 0.75m thick, are moderately to weakly stratified, consisting mostly of well-rounded pebbles and cobbles (Gms). Clasts are predominantly Coal Measure sandstones with subsidiary amounts of Lake District igneous rocks; some are striated and have typical 'flat-iron' shapes. The matrix texture is dominantly moderately to poorly sorted fine sand to gravel (6.7% gravel, 92.9% sand and 0.4% silt/clay). Within the gravels are thick layers of moderately sorted, coarse sand and granules up to 0.4 m thick (Sh). The sands are interspersed within the sequence, >1m thick, and are thinly stratified. They contain

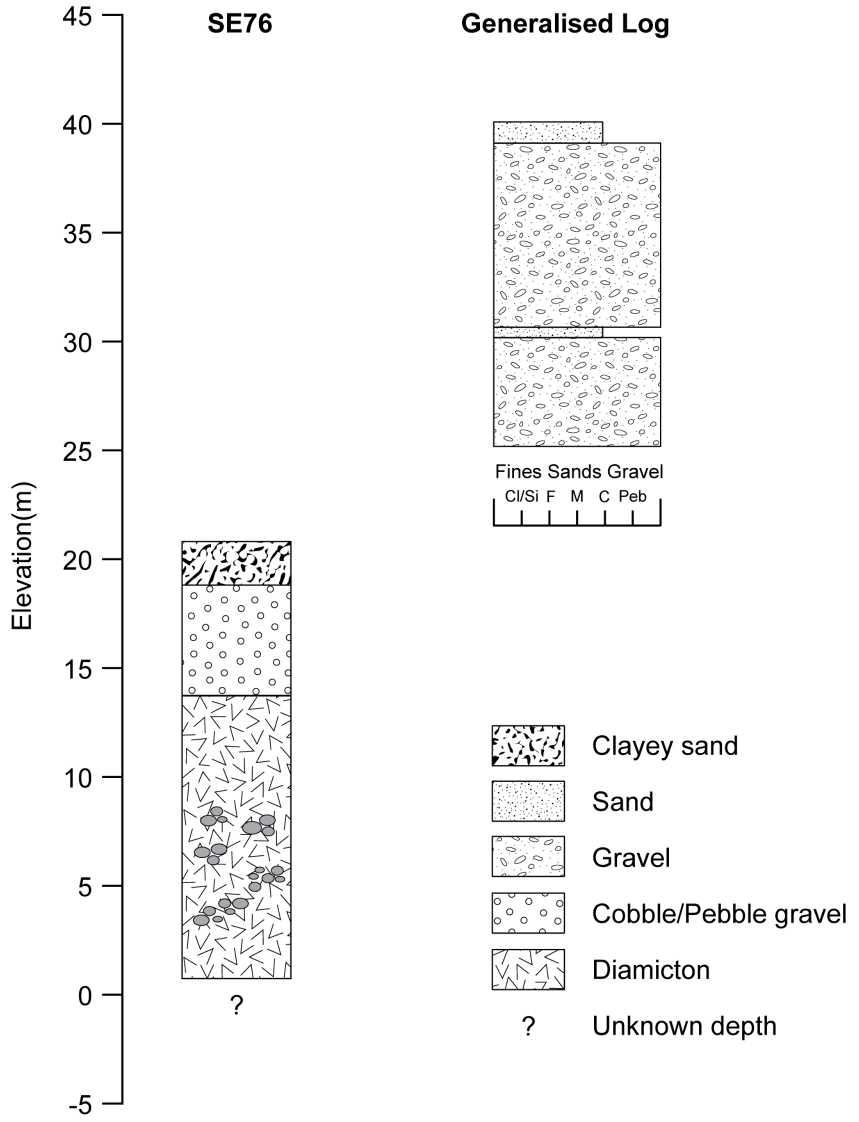


FIGURE 5.8: Generalised stratigraphic log for the exposed section at Stocksfield (Bullion Hills) and the IMAU borehole log (SE76), which was recovered nearby (see Figure 5.31 for location).

fining up sequences of medium sand to very fine sand/silt (62% sand and 38% silt). Contacts between layers vary, some are sharp and some gradational. The sandy diamicton is exposed at the base of the sequence; it contains facetted clasts and is a medium to fine sand. The sequence is chaotic and appears to reflect varying depositional conditions, with evidence of some sorting by water but is, essentially, unstratified. There is some localised evidence of soft-sediment deformation in the form of load casts, suggesting deposition over saturated sediments.

The sedimentary sequence: Location 2

A second location at Stocksfield was recorded, *c*. 0.7km to the east of location 1. The sediments are exposed in a quarry at Merryshields (no longer working), providing access to the sediments underlying the degraded and ice moulded landforms. The former working face comprised a sequence of cobble gravel up to 3m thick (although exposure was limited), with some boulders (up to 0.5m *a*-axis) (Figure 5.9A). Overall, the gravel is weakly stratified, although some boulders show possible imbrication with small cobbles trapped behind the larger clasts. The unit is matrix supported, with the matrix texture dominantly pebbles and coarse sand/granules (Figure 5.9B). There is clear evidence of frost shattering of boulders and cobbles within this unit. The sequence is poorly exposed but deposition appears to have been in channels which can been traced at the top of the present day exposure.

Sediment exposure in the vicinity of Stocksfield is extremely poor; both in the quarry and the Bullion Hills exposure and it is difficult to appreciate the full extent of the face. However, notwithstanding the paucity of exposure there was no clear evidence of syn- or post-depositional faulting which could suggest deposition over stagnant ice or any evidence that the sediments were exposed to permafrost i.e. ice-wedge casts. However,

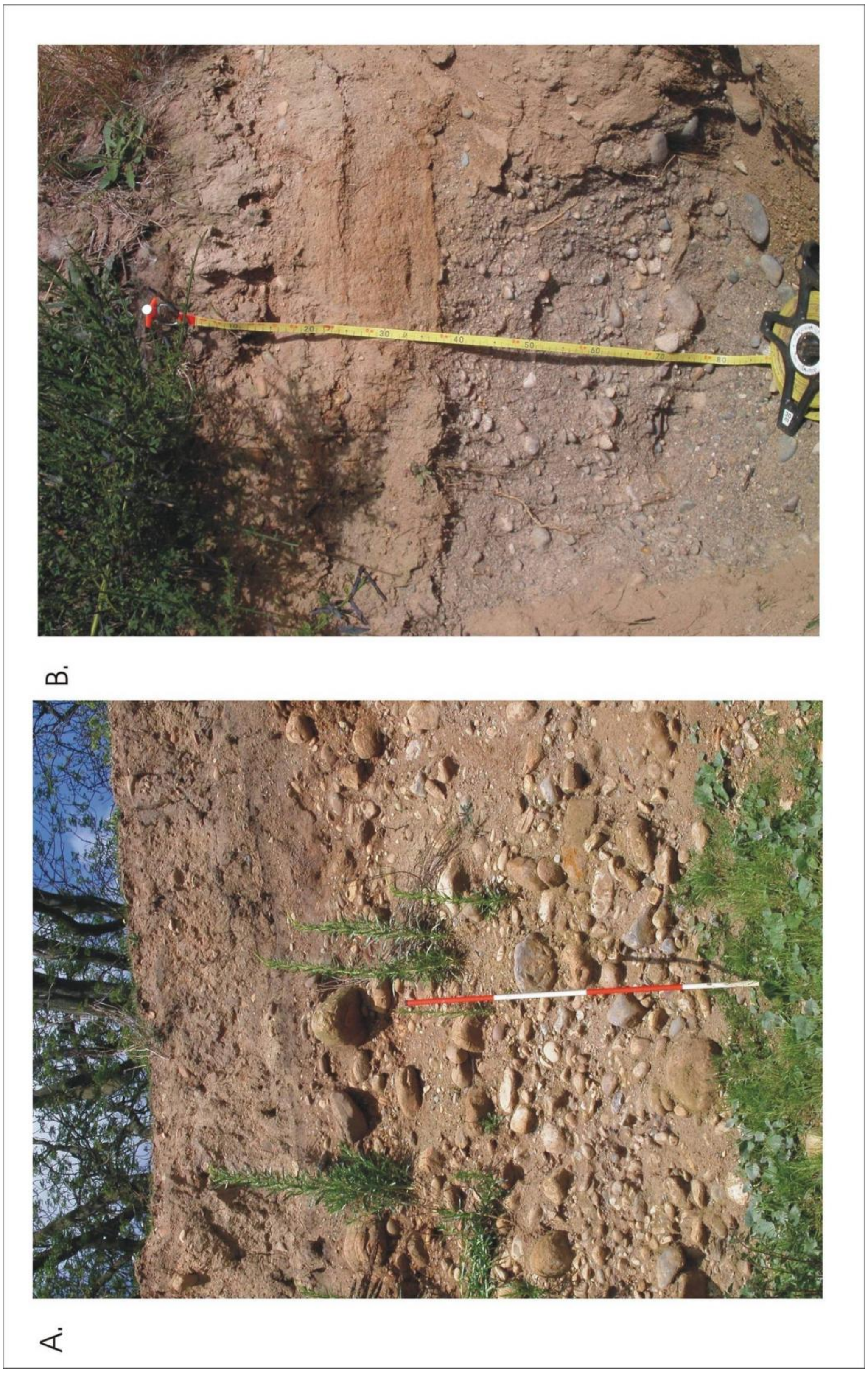


FIGURE 5.9: A. Detailed view of matrix supported boulder/cobble unit exposed at location 2, Merryshields quarry. Ranging pole has 0.5m markings. B. Detailed view of stratified gravel overlain by sands, at location 1, Bullion Hills. Tape rule is c. 1m in length.

the IMAU borehole log data does provide some indication of the likely depositional environments. Figure 5.10 shows the IMAU borehole logs for the area around the Bullion Hills and Merryshields Quarry, Stocksfield. The logs illustrate a degree of local sedimentological variation, which is not unexpected in a former glaciated river valley. Borehole SE78, taken through the undulating and glacially moulded topography, reveal

a tripartite sequence comprising >7m thick of pebble and cobble gravel and the sequence is capped by ~3m of clayey sand. Borehole SE75, located immediately to the east of the Bullion Hills exposure records a bipartite sequence comprising diamicton, intercalated with clay (sandy silt). The boreholes taken through the valley floor reveal sequences up to 25m thick (possibly even greater but the logs only record 25m) of clayey sands, silts and clay. Borehole SE77, located immediately to the west of Merryshields quarry (Figure 5.10) records a sequence comprising diamicton, sand, and up to 14m of laminated silts that alternate with bands of silt and clayey sand. The sequence underlies a landform-sediment assemblage that is arranged as series of gently sloping mounds, now dissected and lying within the much larger former channel of the Stocksfield Burn. Boreholes SE74 and SE76, which lie in the flat, River Tyne Valley floor zone, beyond the landform-sediment assemblages of Merryshields and Bullion Hills, record >10m of clay and silt (diamict) underlying the an upper ~7m comprising typical channel bed and overbank sediments. The sequences around the Bullion Hills and Merryshields quarry are locally variable, comprising large matrix supported boulders, sands and gravels, intercalated diamicton, and rhythmites.

Interpretation

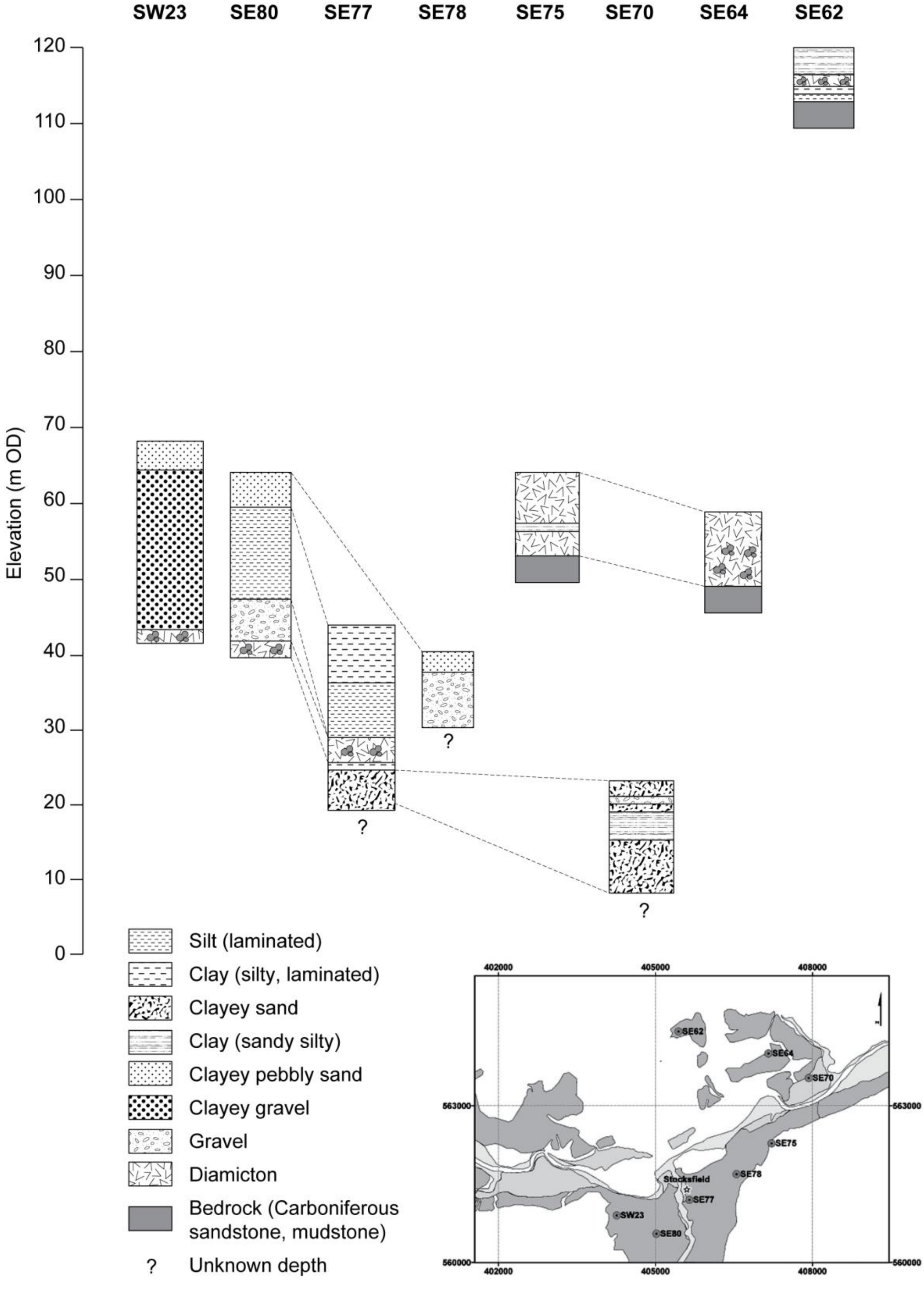


FIGURE 5.10: Stratigraphic relationship of the sediments in the vicinity of Stocksfield from IMAU boreholes. Correlations between sedimentary units are tentative. Borehole location map inset.

The interpretation of the sequences exposed at Stocksfield is based on the integration of borehole data with field observations. The sediments exposed have a rather chaotic depositional pattern and it is not possible to assign Miall's (1978) lithofacies codes to the exposed sections. The sequence exposed in Bullion Hills is dominated by a contrast between sorted and unsorted material, from coarse boulders to medium sands, combined with intercalated diamicton recorded in the borehole logs; it suggests an ice-proximal or pro-glacial environment. The general lack of sorting suggests limited transport and points away from an alluvial or glaciofluvial depositional setting. The sediments show some evidence of sorting by water, but this is very localised. Chaotic sequences like this are usually interpreted as indicative of deposition by gravity flows (Lowe 1982). The exposure at Merryshields, which comprises an eclectic range of poorly sorted gravel from cobble to boulder supported in a granule matrix, could represent debris flows at the ice margin, forming a series of overlapping channel deposits. The sequence shows some imbrication where open-work gravels occur, clast clusters, an absence of grading and common outsize clasts. Similar ice proximal/marginal deposits were observed in late Pleistocene progading outwash sequences recorded by Benn (1996) in the Cairngorms, and Aitken (1998) in the Don Valley. They suggested emplacement as a high-concentration, cohesionless debris flow (cf. Eyles 1987; Nemec and Steel 1984; Postma *et al*. 1988).

Based on the morphology of the landforms, particularly the Bullion Hills, it would be easy to interpret these irregular, hummocky features as glaciofluvial outwash deposits aggraded over stagnating ice; however, the lack of sorting, chaotic nature of the sequence and the inclusion of stratified diamicton suggests proximity to ice (cf. Bennet and Glasser 1996; Fard *et al*. 1997; McCabe *et al*. 1987; Lønne 1993; Ward and Rutter 2000). On the basis of the borehole data, Mills and Holliday (1998) interpreted the landform-sediment assemblages at Stocksfield as the result of englacial deposition, thus explaining both the sedimentology and the surface morphology. However, there is no evidence, such as major faulting or deformation, to suggest that the sediments accumulated either within or above ice.

The lack of sorting would preclude glaciofluvial transport beyond the ice-margin; therefore the sequence must have developed at the ice-margin. The sediments were most likely being dumped and were slumping off the ice in no particular order as it disintegrated. The intercalation of diamicton recorded in some of the borehole logs would further suggest that deposition was ice-marginal but, although the sediments were probably aggraded as debris flows, this cannot be confirmed. The occurrence of intercalated sands and diamict, and laminated silts in a zone adjacent to the chaotic assemblages of Merryshields and the Bullion Hills, suggests a proglacial sedimentary sequence developed as a subaqueous fan, probably a distal fan given the presence of rhythmites in the borehole log (cf. Delaney 2002; Thomas *et al*. 2004). The load casts suggest a saturated environment (cf. Delaney 2002), not unexpected at a disintegrating ice-margin with continued drainage of water and ponded water bodies develop between the margin and moraine. The silty sand diamicton that underlies the zone of deposition beyond the mounded topography is similar to the sediments exposed in Bullion Hills, providing support for the notion that ice sat in the valley at this location (cf. Ward and Thomson, 2004), a mechanism for localised ponding of the area.

Thus, the landform-sediment assemblages described are interpreted as the zones of deposition around a disintegrating ice margin, possibly a tongue of ice sitting between the main valley and the Stocksfield tributary valley. The surface morphology is, however, the result of subsequent erosional processes, which dissected the sediments as either meltwater flows or when the local drainage network was re-established. There is no evidence to suggest deposition over ice as an explanation for the morphology. Thus, although the morphological features at Crawcrook and Stocksfield share the same genetic development model, they represent different landform-sediment assemblages.

5.2.3 Farnley Haugh Scar

The sedimentary sequence

An exposure of the sediments underlying terrace landforms occurred in a cut-bank, where a sequence has been exposed by contemporary river erosion, undermining the slope and revealing up to 10m of sediments (Figure 5.11). The sequence consists of two units (see Appendix). The basal unit (Unit 1) comprised of up to 5m thick of horizontally stratified, silty sand, dipping at 22˚. The sand is poorly sorted and its texture is dominantly very fine sand (64.5%) and silts (35.5%). The lower contact of unit 1 could not be determined due to the amount of talus at the base of the scar. The upper contact is sharp and erosive. Figure 5.11 illustrates the stratigraphic relationship of the logs recorded from the Farnley Haugh Scar exposure.

Unit 1 was uncomformably overlain by Unit 2, a series of coarse gravel and sand layers up to 10m thick that pinch out to the east. The gravel represents cross-section through a large channel (Figure 5.12A). Overall, the most extensive gravel layer (*c*. 1.5m thick) consists of coarse gravel. The lower 50cm is a horizontally stratified, clast-supported, cobble and large pebble gravel with occasional boulders (up to 0.37m *a*-axis), with a matrix of moderately sorted gravel (29.5% coarse, 49.4% medium and 21% fine) (Gms). This grades upwards from cobbles to pebbles and granules in a repeated sequence of three normally graded layers up to 60cm thick. The sub-units, present as layers (10-30cm thick) within the gravels, consist of small pebbles in a matrix of

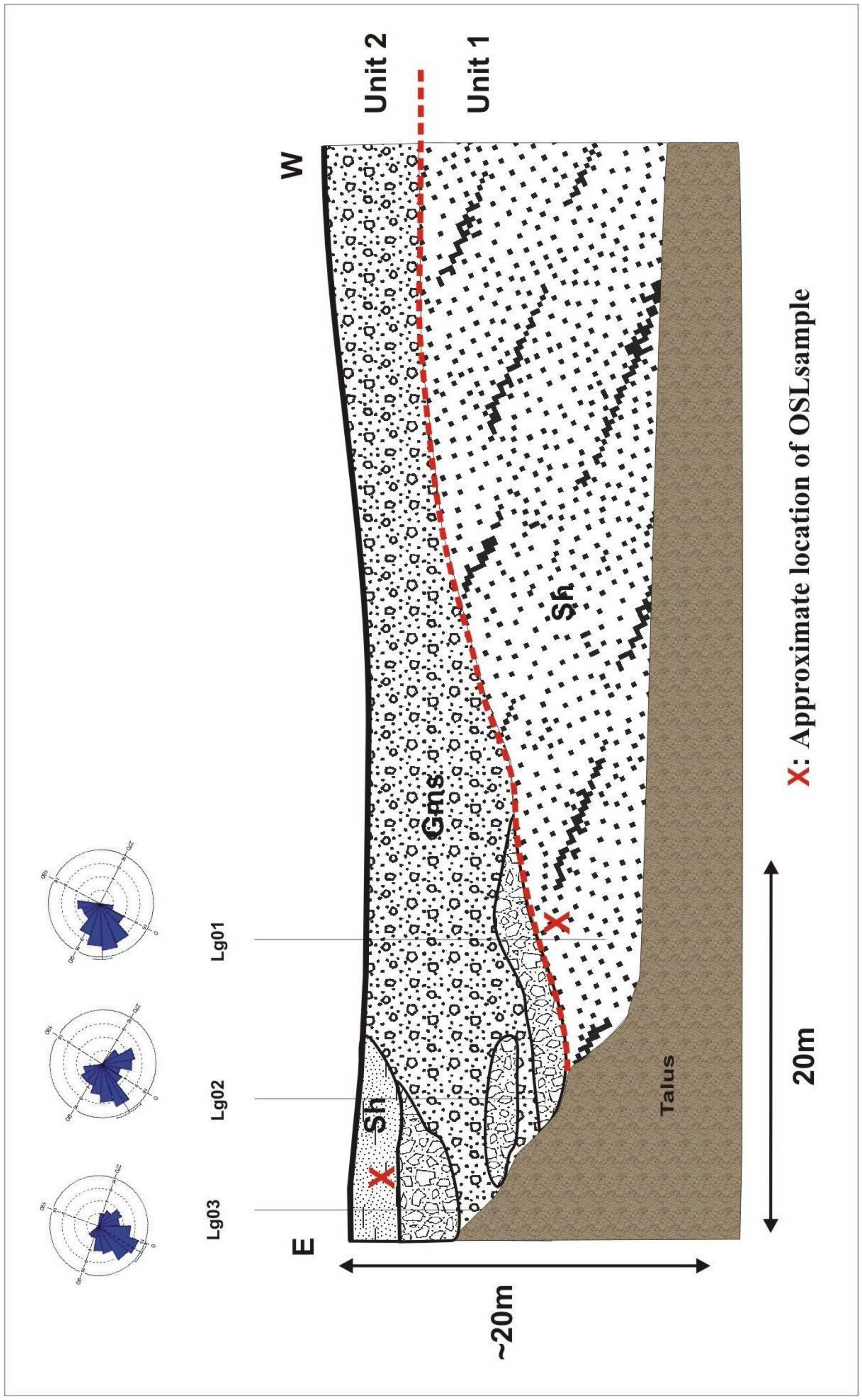


FIGURE 5.11: Generalised sketch of the main face at Farnley Haugh Scar. The location of the vertical logs, OSL sample locations are marked, and the palaeocurrent fabrics given (0 representing north).

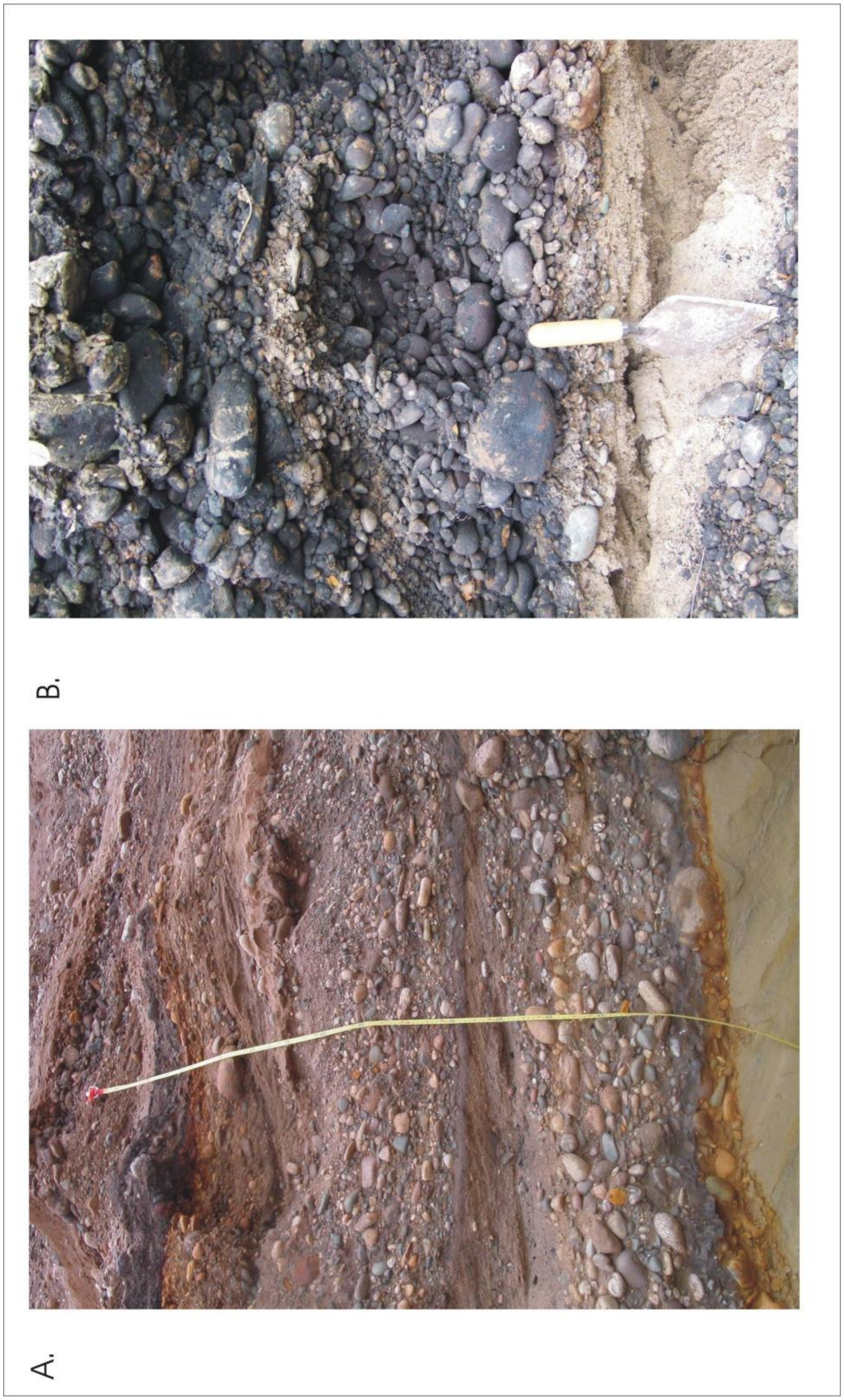


FIGURE 5.12: A. Detailed view of channel infill of unit 2. Note the iron staining, and the sharp contact between the gravels of unit 2 and the glacigenic sands of unit 1. Tape rule is *c*. 1.5m in length. B. Clast supported pebbles, overlying horizontally bedded sands. Trowel is 28cm in length.

moderately sorted, fine gravels and coarse sands (8.7% gravel and 91.2% sand) and cross-stratified (dip 026^0), very fine sands (Sp). The basal gravel unit is indurated and the cobbles are stained black (with manganese), although most of the pebbles within the unit have either black or orange-buff (hydroxide) colouration. The two further gravel layers that constitute this unit are much smaller in their spatial extent (~3m wide). The intermediate gravel is a massively to crudely stratified, clast-supported pebble gravel, that shows vertical and lateral grain size variations. The gravels are imbricate, highly indurated and covered with a black staining (Figure 5.12B). The upper gravels are clast- to matrix-supported, the matrix texture is a moderately to well sorted, fine gravel and coarse sand. The gravels are weakly cemented with calcium carbonate ($CaCO_3$), suggesting the terrace may be quite old, assuming a period of time required $CaCO_3$ to precipitate from aqueous solution and form a surface cover on the gravels under the temperate Holocene climate (Koutsoukos and Kontoyannis, 1984). Interspersed with the gravel wedges are subunits of three types: (1) thinly laminated, moderately sorted coarse sands and fine gravel (98.5% sand, 1.4% silt and 0.1% gravel) up to 25cm thick; (2) thinly laminated, poorly sorted gravely sands (5.1% medium/fine gravel, 91.7% sand and 3.2% silt) and (3) thinly laminated, poorly sorted silty fine sand (55% fine sand and 45% silt). Contacts between the gravels and the sands are sharp and distinct. Palaeocurrent fabrics obtained from the three gravel wedges indicate flow down valley (Figure 5.13), in a south-south-west direction. The lithological composition of Unit 2 suggests a high input of local material. Local sandstone predominates with subordinate amounts of Permo-Triassic and Greywacke sandstones and igneous material from the Lake District and Southern Scotland (95.1% local material and 4.8% exotic material).

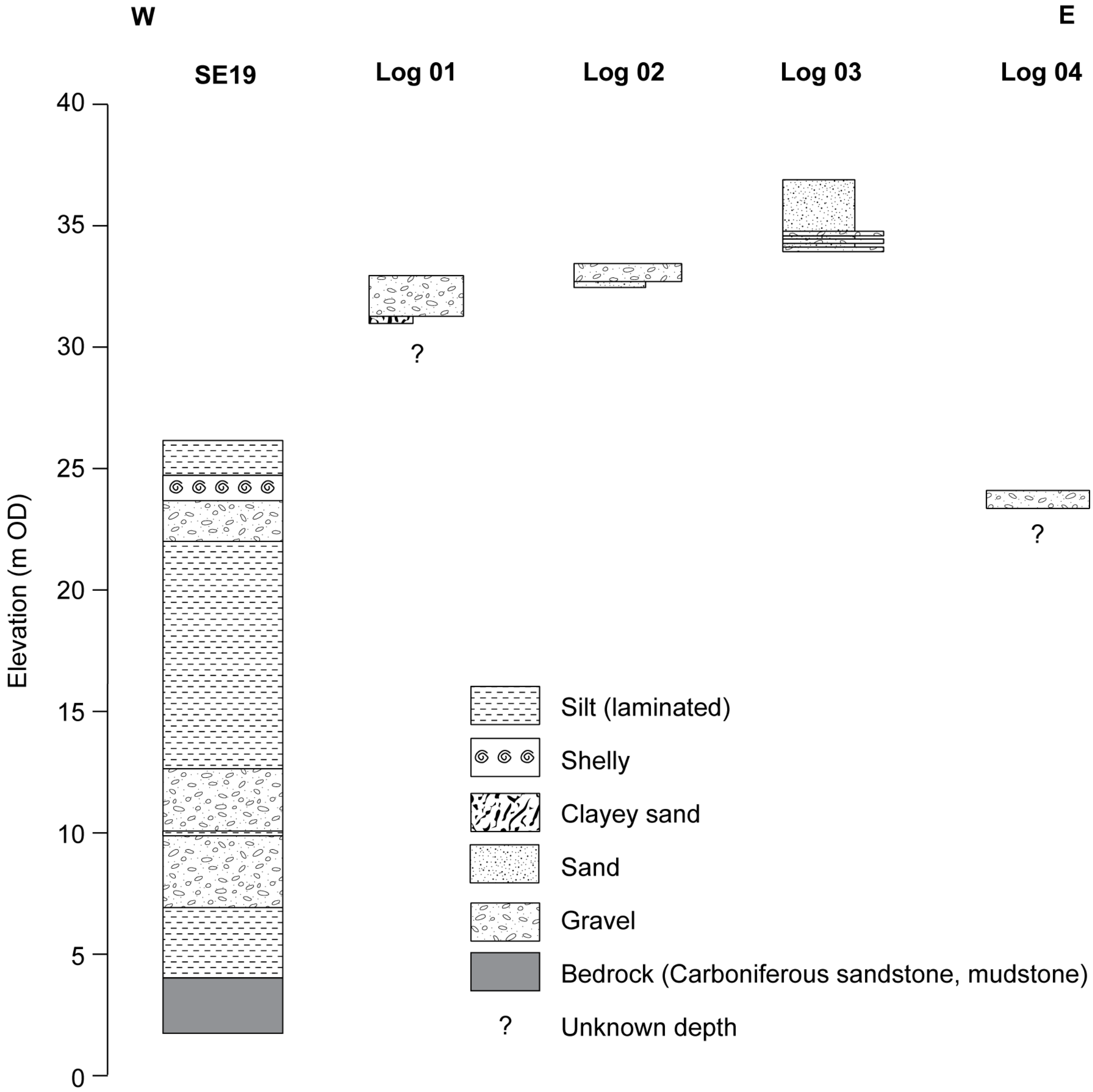


FIGURE 5.13: Stratigraphic relationship of the vertical logs recorded from the exposed face at Farnley Haugh Scar. IMAU Borehole SE19 is also shown to illustrate sedimentology underlying the valley floor at this locality.

The unit is capped by up to 2.3m of horizontally stratified, thinly laminated, and poorly sorted sands. The texture is dominantly silty fine sand (64.5% sand and 35.5% silt) although there are very thin layers of granules interspersed within the sands (Sh). The contact is gradational between the upper gravel and the sands. Locally, although it could not be examined due to access issues, a weakly stratified, matrix-supported coarse gravel unit comprising cobbles and boulders, with some imbricated clasts (Gms), can be seen at the top of the sequence.

Interpretation

The basal unit (Unit 1) is sedimentologically similar to Crawcrook Unit 1 (location 1 and 2). The horizontally stratified, silty sands suggest a low energy depositional environment, probably from suspension settling or distal underflow currents in a glaciolacustrine environment. Given the dip of the beds, the deposit is interpreted as a subaqueous fan, rather than lake bottom sediments. The lack of sorting, structures and truncations that are usually associated with glaciofluvial or fluvial environments lends support to the interpretation. Given this is a limited exposure, the fact that this layer has been truncated and eroded by the River Tyne as it cut down through the valley fill sequence, and that there are limited borehole records in the immediate area from which to gather more data, the interpretation must remain tentative. Unit 2 is interpreted as a fluvial deposit. The imbricate cobble-pebble facies is interpreted as a channel-floor and bar deposit (Ferguson and Werrity 1983; Forbes 1983). Aggradation occurred in low water as tractional processes transported the gravels as part of the bedload and during the falling flow stage allowing the bar to form and bar top sediments to be deposited (Bluck, 1979). The matrix, where present, subsequently infilled the intersticies between the clasts. The intercalated sets of horizontally stratified and cross-stratified sands and small pebbles are interpreted as bar-top/edge accretions (Miall 1978; Rust

1972). The small-scale stratification is the result of shallow flows over the top of the gravel bar (Blacknell 1981, 1982; Nemec and Steel 1984). Smith (1990) suggests the iron-manganese coatings form at the sediment water interface, therefore the coated gravels represent accretion in the lower portions of the channel or bar. The local occurrence of matrix-supported coarse gravels at the top of the sequence (which could not be properly examined) probably represents a debris-flow or flood deposit as the poor stratification, outsize clasts and extensive matrix suggests sediments moving in a high density, sediment laden flow (Nemec and Steel, 1984). The horizontally stratified silty sands and fine gravel that cap the sequence are interpreted as channel fill or overbank sediments deposited from upper regime flow during a flood (Deslodges and Church, 1987). The sequence is typical of a wandering gravel bed river (*sensu* Church *et al*., 1981) with migration across the valley floor and deposition as both channel floor and bars.

5.2.4 Fourstones

The sedimentary sequence

Four major surfaces have been identified on the south side of the River South Tyne at Fourstones, inset below a glacially smooth rock outcrop and glacially eroded channel. A single major exposure (~1.5km long) has been created by fluvial erosion and the cut-bank section revealed sediment exposure, up to 8m thick, in two terrace units (upper and lower) (Figure 5.14). In part, some of the sediments exposed at Fourstones directly overlie diamicton, and it is clear to see that the contemporary river partly runs out on a diamicton channel bed. This indicates that a significant level of incision has been achieved by the river since deglaciation.

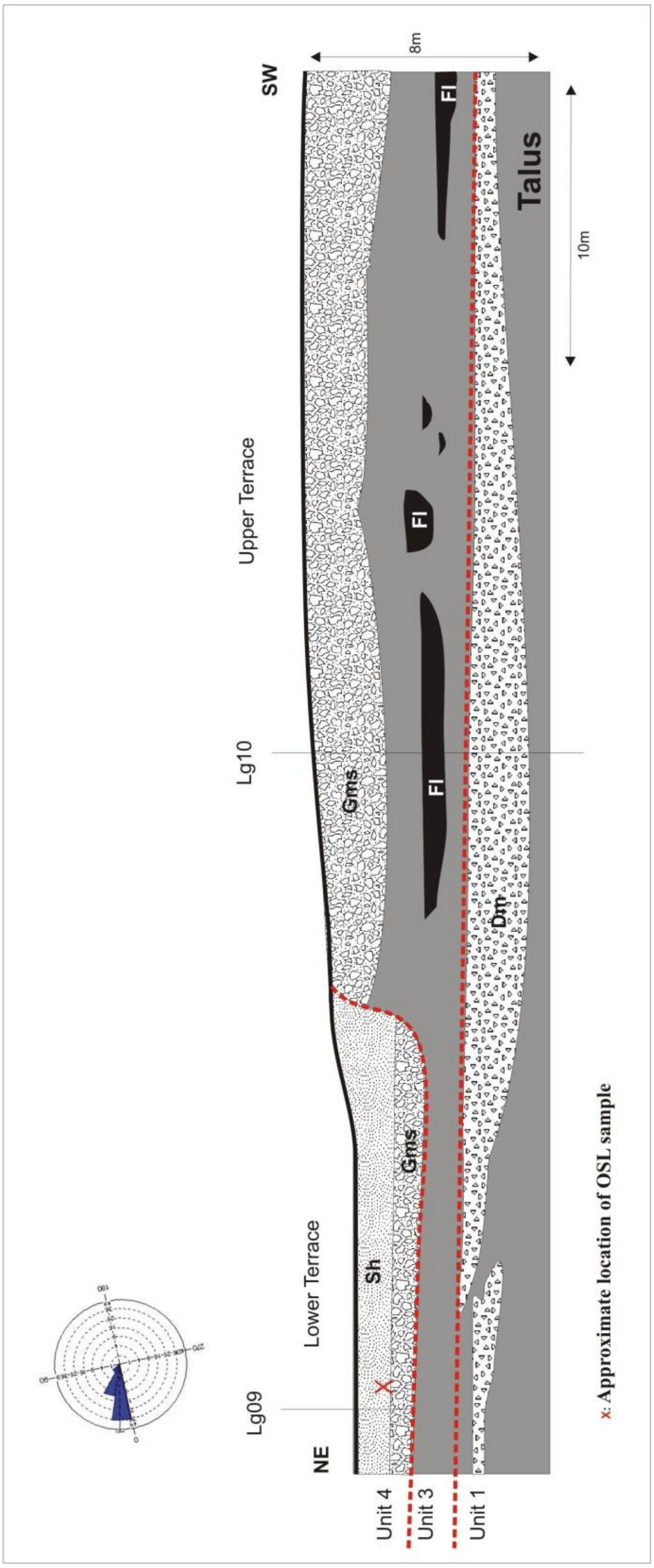


FIGURE 5.14: Generalised sketch of part of the Fourstones face, revealing the wandering gravel bed sediments of the upper terrace, and the contact between the upper and lower terraces. Log locations, OSL sampling locations and palaeocurrent directions all marked.

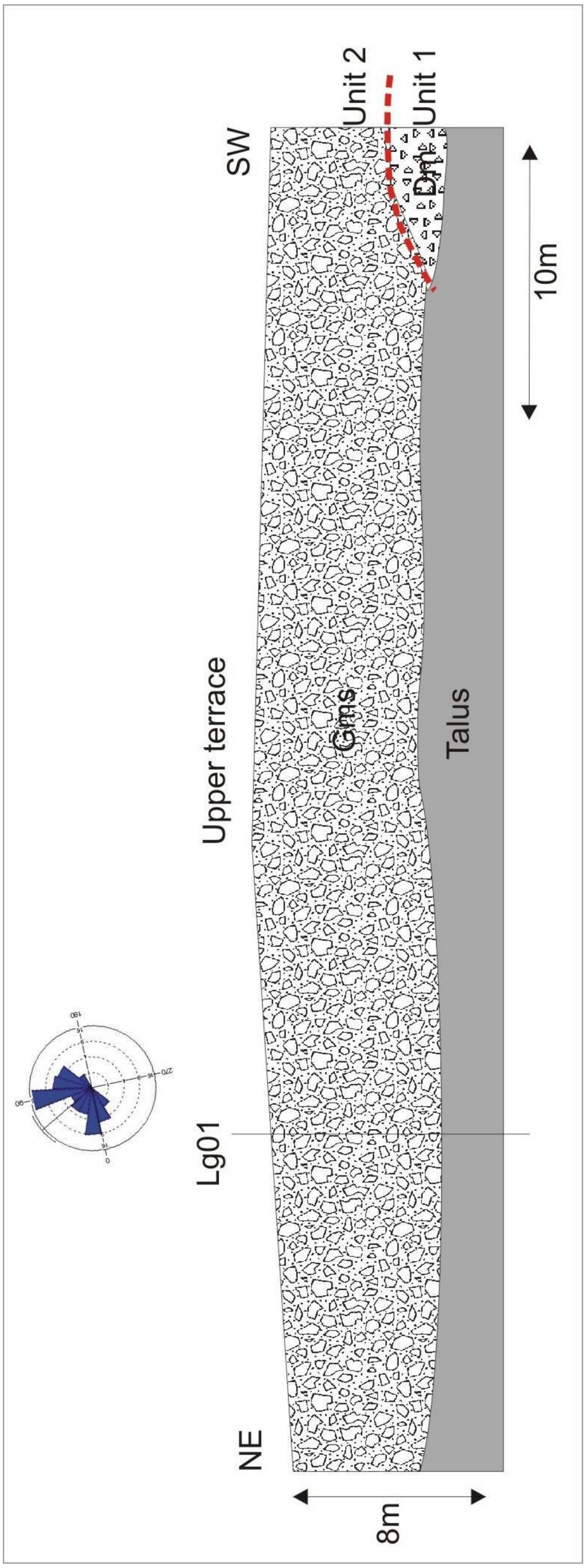


FIGURE 5.15: Generalised sketch of the of the Fourstones face, revealing the cobble gravels of the upper terrace, showing log locations, OSL sampling locations and palaeocurrent directions.

The surface of the upper terrace examined (see section 5.6 for classification) is at 52 m OD and is ~8m above the present river-bed. Unit 1 (Figure 5.14; Figure 5.15; Figure 5.16 Log01, 02, 03) consists of a massive diamicton. The upper contact is sharp but undulating; the lower contact was not determined. Concentration of clasts is ~25%, and consists of mostly pebbles with some boulders. The clasts are striated, and comprise Lake District igneous material and local sandstones. The matrix is poorly sorted, consisting of 11.6% fine gravel, 46.9% sand and 41.6% silt. The diamicton appears thinly laminated in the upper 30cm, at its contact with the overlying unit. Unit 2 consists of a thick sequence of coarse gravel, with an erosional lower contact. The lower 2m is massive to horizontally stratified, iron-indurated gravel with clasts up to 10cm, supported in a matrix of gravely sand (20.9% fine gravel, 78.6% coarse sand and 0.5% silt) (Gms) with a well developed imbrication. A palaeocurrent fabric obtained from this layer indicates flow in a north-north-east direction, approximately downstream (Fig. 5.15). Above this are ~3m of cobble- and boulder-rich gravels, supported in a matrix of gravely sand (20.1% fine gravel, 78.7% coarse sand and 1.2% silt). The gravels are structureless or weakly stratified in horizontal beds, with normal grading in some of the layers (Gms) but more often in a non-uniform distribution. Clasts are mostly cobble size with boulders up to 0.68m *a*-axis, and are well-rounded. Clasts comprise local sandstones, with subsidiary amounts of mudstone, shale and fossiliferous limestones, subordinate amounts of Lake District igneous rock and more rarely, diamicton soft-sediment clasts. Gravels are indurated with black-orange staining and cementation, most likely iron manganese coatings (Smith, 1990), and imbrication is well developed. Two palaeocurrent fabrics obtained from this layer indicate a flow direction of north-north-east and east-south-east, predominantly down valley. The unit is truncated towards the top of the sequence by a scour feature infilled with planar cross-stratified cobble and pebble gravel (Gms, Gp) and massive gravels

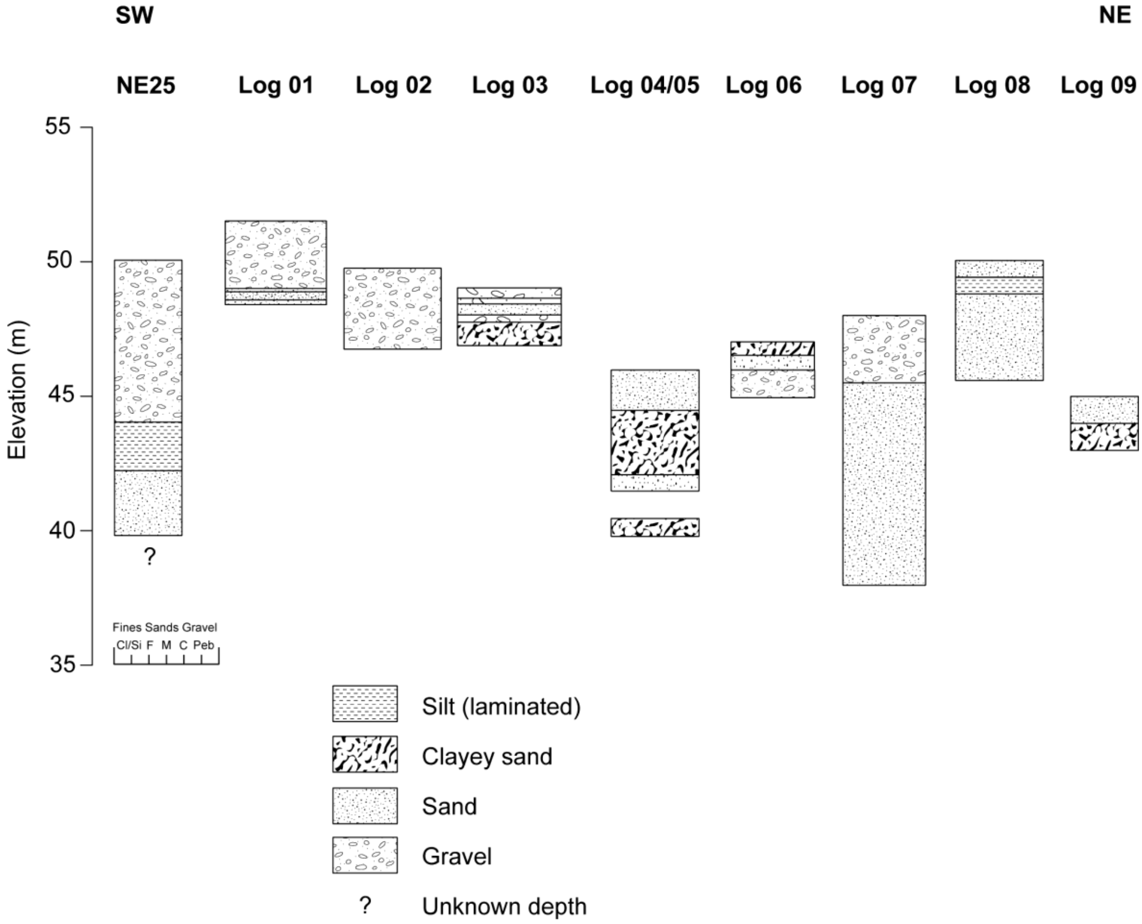


Figure 5.16. Stratigraphic relationship of the vertical logs recorded from the cut-bank exposure at Fourstones. NE25 represents the IMAU borehole taken through the lowest terrace. Location of OSL sample locations are indicated by the red X.

supported in a matrix of granules and coarse sand. Interbedded with the gravels are subunits of thinly laminated, planar cross-stratified (dip 020˚-028˚) pebbles and granules (Gp) and horizontally stratified coarse sands (Sh) up to 10cm thick and only 1-2m long. Foresets are rare in this sequence and the absence is notable.

There is a marked change in the sedimentology of the upper terrace downstream (Figure 5.14), and the sequence here comprises a single unit (Unit 3) erosively overlying the basal unit (Unit 1). It consists of coarse gravels and sands. The gravels are truncated by asymmetrical channels (*c*. 1m deep, 1-2m long) infilled with horizontally stratified, poorly sorted fine sandy silt (42.8% fine sand, 57.2% silt) (Sh). The gravels consist of weakly to horizontally stratified, cobble and pebble gravels, especially towards the top of the unit, supported in a matrix of granules and coarse sand (1.9% fine gravel, 97.4% medium sand and 0.7% silt) (Gms). Horizontally stratified sands (texture undetermined) overlie and interfinger with the gravels in the upper part of the sequence, but these are laterally discrete (Sh). Much of the sequence underlying this terrace was inaccessible due to the sheer face of the exposure and, therefore, some detail is missing in the description.

The lower terrace is the laterally most extensive and well-developed unit at Fourstones, lying ~5m above the present river-bed (see section 5.6 for classification). The sequence comprises a single unit, Unit 4 (Figure 5.14; Figure 5.16, Log09) which consists of coarse gravel overlain by ~2m of sands. The lower contact is obscured by talus and <0.5m are revealed of the gravels. The gravels consist of stratified cobbles, with occasional boulders up to 20cm, supported in a matrix of poorly sorted gravelly sand (21.6% fine gravel, 78% coarse sand and 0.4% silt) (Gms). Clasts are rounded, comprising local sandstones with subsidiary amounts of Lake District igneous rocks.

Clasts are iron-indurated (unconsolidated) and imbrication is well-developed; a palaeocurrent fabric obtained from this layer indicates a flow direction of north-north-east, approximately downstream (Fig. 5.14). The gravels grade upwards into the sands and the contact is gradational. Overall, the sands consist of moderately sorted gravely sands (Sh) (0.2% fine gravel, 96.1% medium sand and 3.7% silt) and are horizontally stratified. Towards the top of the unit, occasional pebbles up to 7cm are present.

Inset into the upper terrace is a laterally discrete unit that is, sedimentologically, very different from the terrace and represents a major change in the calibre of sediment supplied from the catchment (Figure 5.16, Log06, Log07). This sequence probably represents a remnant of a terrace (see section 5.6 for classification). The surface of the unit is at 52 m OD and is situated ~8m above the present river bed. It comprises three units (Appendix Table A.4). The basal unit, Unit 1, consists of diamicton and underlies the terrace sequence. Up to 1.5m of diamicton is exposed and the contact is both erosive and irregular. Unit 2 consists of <1m of sands, the upper and lower contacts are both obscured by talus (Figure 5.15, Lg04/05). The sands are horizontally stratified, thinly laminated, moderately well sorted fine-grained sand with drapes of silty clay and clay. Lithic grains pick out individual lamina planes. The texture is dominantly fine sand (97.7%) but it does contain very fine gravel (0.7%) and silt (1.6%).

This is overlain, although the contact could not be determined due to talus, by Unit 3, which comprises ~2m of coarse gravel and sands. The lower ~25cm consists of weakly stratified, imbricate cobbles, pebbles and rare boulders supported in a matrix of pebbles, granules and sand (Gms). This grades upwards into ~1.5m of horizontally stratified and massive, well sorted silty fine sands (Sh, Sm, Fl). Lithic grains pick out individual laminae planes. The upper 1m comprises soil, although cobbles and pebbles are common towards the base, and the contact is gradational.

The unit is truncated by two large scour features (up to 3.5m deep and 1-2m long) consisting of coarse gravel (Figure 5.16, Log06). The first scour feature is infilled by massive boulders and cobbles, up to 16cm, and in parts is clast-supported but is mostly supported in a matrix of granules and coarse sands (Gms). Interbedded with the gravels is ~40cm of horizontally stratified and planar cross-stratified (dip 24˚), well sorted medium to coarse sands (Sp, Sh). The upper ~1.5m of gravel consists of horizontally stratified pebbles supported in a matrix of granules and sands. The subunits of gravel exhibit both normal and inverse grading. The second scour infill contains a lower ~20cm of imbricated (well developed) pebbles that are indurated with orange-black (iron-manganese) staining. This is overlain by ~30cm of clast supported (pebble) gravel with occasional cobbles and capped by ~10cm of planar cross-stratified sands (Sp). The clasts are mostly local sandstones, with subordinate amounts of igneous material. The first scour feature includes some rip-up soft-sediment clasts. A palaeocurrent fabric obtained from these layers indicated a flow direction of south-south-east, approximately perpendicular to the contemporary downstream flow direction.

Figure 5.17 shows the IMAU borehole logs for the terrace sequence that have been identified along the South Tyne Valley. The borehole logs illustrate that the sequence underlying the upper terrace comprises thick accumulations of gravel, similar to those recorded in the cut-bank exposure, with gravel at the top of the sequence underlain by sands and intercalated laminated silts repeated up the sequence. Borehole SW33, taken through the lower terrace, suggests a veneer of more recent alluvial gravels and

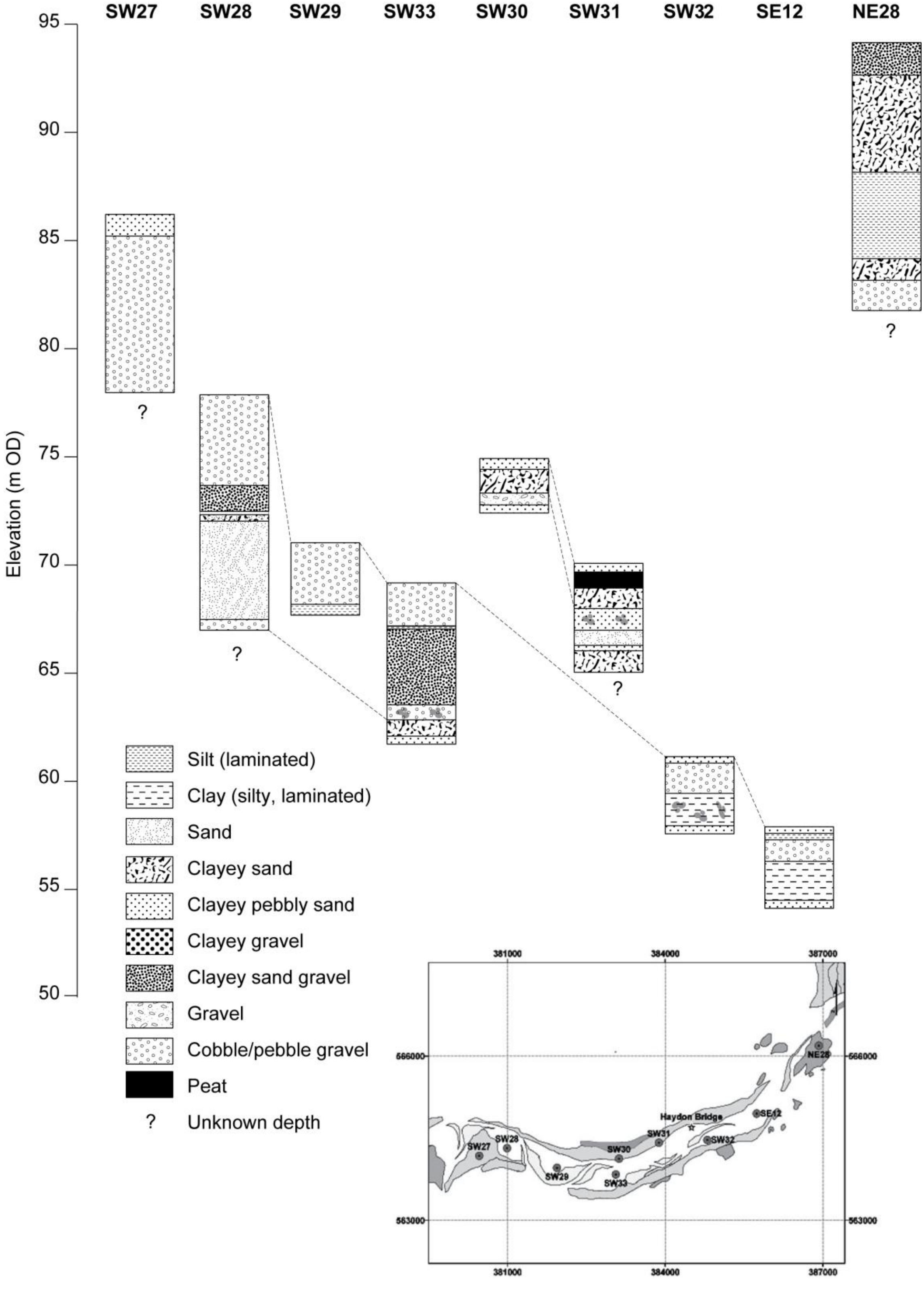


FIGURE 5.17: Stratigraphic relationship based on IMAU borehole data of the sediments underlying the terrace sequence along the South Tyne Valley between the Allen confluence and Fourstones. Borehole locations shown on inset map.

overbank deposits overlying an intercalated sequence of red and grey clay and clayey gravels, possibly related to episodes of dominance by the ice streams originating from the Lake District (red) and the North Pennines (grey). Passmore and Macklin (2000) recorded a cut-bank section in the upper South Tyne at Lambley that showed the same interaction between Lake District till and local till in the basal sequence. Borehole NE28 (Figure 5.16) records the sequence through the outwash terrace up to 25m thick, consisting of intercalated gravels, clayey sands, silts and sandy gravels.

Interpretation

Diamicton

The diamicton that underlies the cut-bank exposure is overconsolidated and stiff and is interpreted as a subglacial till (cf. Clarke *et al*. 2008). It is not a supraglacial melt-out or flow till, which are often slumped, with downslope stratification and dewatering structures (Benn and Evans 1998; Menzies 2002). The clasts contained therein are striated and faceted, which suggests prolonged glacial transport (Ward and Thomson, 2004).

Upper terrace

The sequence underlying the upper terrace is interpreted as a fluvial deposit. The lower 2m of the unit is interpreted as channel floor (Miall, 1978). The gravels developed as bedload transported along the channel floor, with subsequent infilling by finer sediments. The imbrication suggests clast rolling along under tractional forces (Johannsson, 1976). The iron staining, as previously explained, is interpreted as confirmation of deposition on the channel floor (Smith, 1990). The overlying 3m of gravel are interpreted as debris or flood flow deposits (Miall, 1978). They are weakly stratified, matrix-supported and poorly sorted, containing many boulders, suggesting that they were deposited by hyperconcentrated flood flows (Nemec and Steel, 1984).

The localised occurrence of planar stratified gravels towards the top of the sequence is interpreted as a linguoid bar or deltaic growth as bar front avalanching (Bluck 1979; Miall 1978) during falling stage flow. The unit includes horizontally stratified sands, indicative of bar top sand wedges (Forbes, 1983). The gravels at the top of the delta are interpreted as lag gravels (Bluck, 1979) deposited during low flow stage.

Unit 3 underlying the terrace surface is typical of a wandering gravel bed river, characterised by bar gravels, overlain by sand, occasional channel fills of sand or silt and channel floor gravels (Church, 1983). The concave sandy lenses within the gravels of T3 are interpreted as drape suspended sediments deposited from slack water in chute channels. The gravels are interpreted as flood flow in the main but towards the top of the sequence, where the gravels posses a tabular and sheet-like geometry, they are interpreted as bedload sheets (cf. Whiting *et al*., 1988). The interfingering of gravels and sands is interpreted as bar-type gravels overlain by falling stage sands.

Terrace remnant

Unit 2 is sedimentologically similar to those recorded at Farnley and at Crawcrook. The horizontally stratified sands contain a fining upward sequence and intercalated drapes of silt and clay. This unit is interpreted as a glaciofluvial deposit. The clay layers appear to be reworked and disaggregated diamicton rather than drapes. The moderately sorted nature of the unit suggests transport and deposition by water, rather than settling out of suspension in a ponded or glaciolacustrine setting. Due to the limited exposure of this unit, it is again difficult to appreciate fully the depositional context. Unit 3 is interpreted as a fluvial deposit. The sequence comprises channel floor gravels overlain by extensive deposition of overbank silty sands (Deslodges and Church 1987; Forbes 1983).

The units described as infilled scours are interpreted as chute channels cut through the overbank sediments during a flood event. The first channel is infilled with a flood deposit that contained a high density sediment load including soft sediment clasts, probably ripped out from nearby channel banks, and is interpreted as a debris flow (Nemec and Steel, 1984). The second infilled channel contains a sequence more suitably interpreted as a bar deposit, characterised by imbricate lower gravels and overlain by a coarsening up sequence of gravels and capped by cross-stratified sands. The sequence is tentatively interpreted as a bar, which prograded in a slough pool (Bluck, 1979) during high stage flow.

Lower terrace

Following the reasoning outlined above, Unit 1 is interpreted as a fluvial deposit. The cobble gravel is interpreted as bedload deposited on the channel floor (Miall, 1978). The silty sands were deposited as a thick layer of overbank material (Desloges and Church, 1987).

5.2.5 Haltwhistle to Hexham

The sedimentary sequence

There were no exposed sections in the landforms (mounds and flat surfaces) above the river terraces in this reach. IMAU borehole logs (Lovell, 1981) were examined to provide subsurface data. In general, they reveal a tripartite sequence of silt, clayey sands and gravels, and gravels. The detailed stratigraphy of the boreholes is shown in

Figures 5.17 (NE28) and 5.18. Diamicton is intercalated within the units, and often caps the sequence. The sequences are mainly in excess of 25m thick, although some of the sediments onlap onto bedrock, only forming a thin veneer (<2m thick). In general, the gravels are coarse, matrix supported (the matrix texture is fine to coarse sand), and may be angular or rounded. The clayey sands are fine-grained and up to 9m thick. The silts are up to 8m thick, and comprise both silt and fine-grained sands. The diamicton is hard or stiff clay, grey-brown in colour. Concentration of clasts, consisting of pebbles and boulders, is unknown. Lithologically the diamicton contains sandstone, limestone, greywacke, granite and dolerite.

Interpretation

The sequences are interpreted as glacigenic, formed in an ice-contact environment as proximal subaqueous outwash from receding ice. On ablating debris-rich ice, sorting is crude, with deposition from both debris flow and fluvial processes (Benn and Evans, 1998). The gravels exhibit evidence of fluvial sorting and probable reworking, but some gravels suggest little sorting suggesting limited transport. The clayey sand and gravels are interpreted as subaqueous glaciofluvial/lacustrine deposits, the silts indicating periods of suspension settling. The diamicton may represent basal till due to its high compaction, deposited due to ice sheet fluctuation, a debris/mass flow deposit that has slumped off the ice or even solifluction, but without examining the fabric and clast form, it is impossible to determine. The sequences appear to be chaotically bedded, reflecting changes in depositional environments that are not contiguous. Proximity to ice is inferred both from the chaotic nature of the sequences and the intercalation of the diamict. Evidence for both localised ponding and sorting by fluvial processes suggest temporary meltwater impoundment as the position of the ice and sediments change.

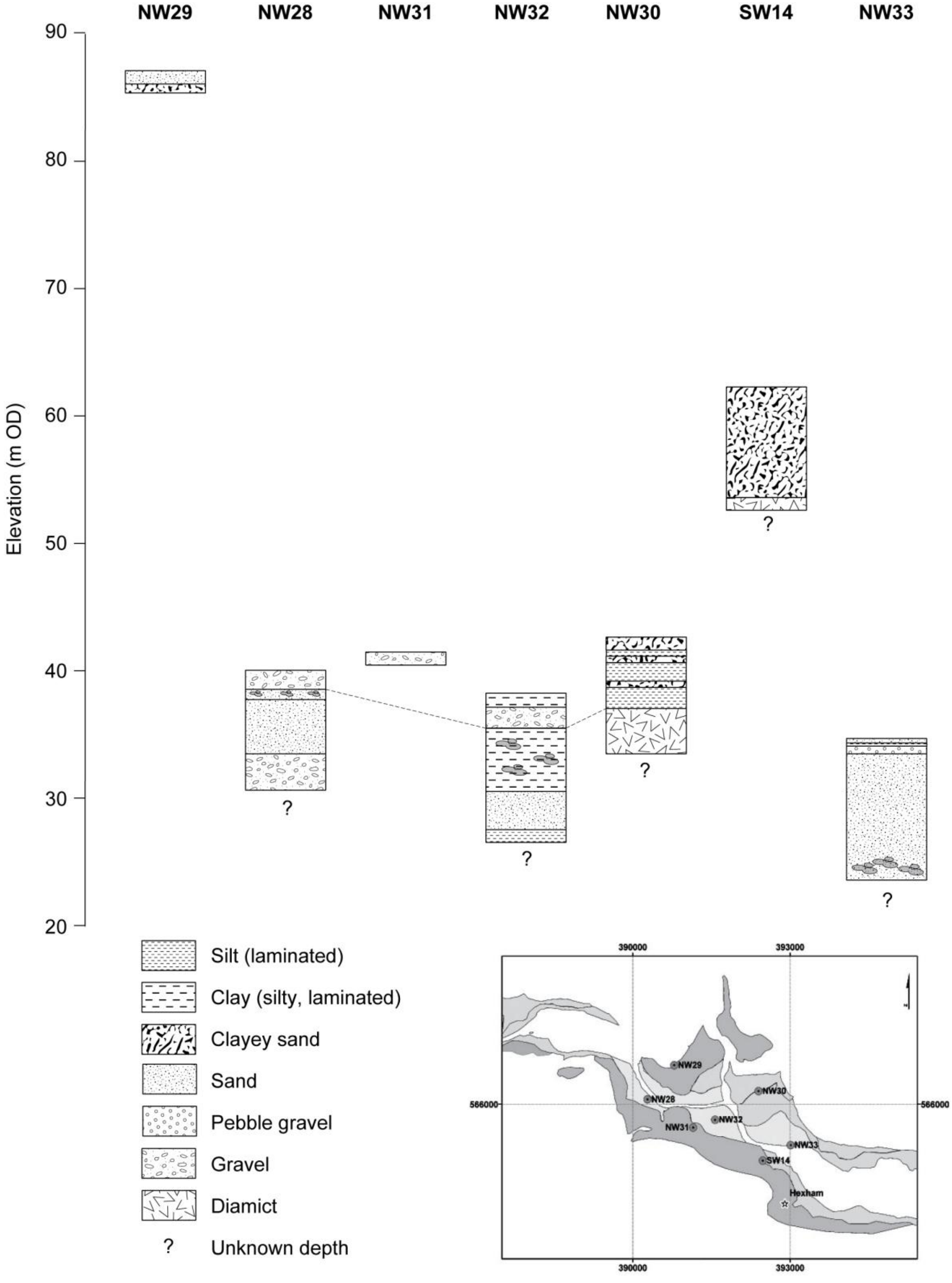


FIGURE 5.18: Stratigraphic relationship based on IMAU borehole data of the sediments underlying the glacigenic complex in the vicinity of the North/South Tyne confluence. Borehole location map inset.

5.3 Geochronology

Attempts to date the sequences were made using OSL dating techniques. OSL ages are presented in Table 5.1. OSL ages are given in years before present (BP). Uncertainties given with the age may represent those associated with the analysis procedure rather than the date (Huntly and Lian, 1999). Thus ages are accurate but subject to uncertainties associated with statistical uncertainty (precision), which reduces the reliability of the date (Lowe and Walker, 1998).

Interpretation

Crawcrook Quarry: Location 1

Three OSL samples were taken sequentially up the face: (1) two from Unit 1 (LV161; LV162); and (2) one from Unit 2 (LV163) (Fig. 4.2). None of the samples was able to provide a luminescence date due to the poor luminescence qualities of the quartz (see Appendix 3 on CD for laboratory report). The quartz had a weak fast component, which meant that the grains sampled did not show suitable luminescence intensity for dating. The samples had low sensitivity, which reduced the possibility of obtaining a date, and finally, noise from an impurity in the crystal quartz lattice interfered with the signal. The laboratory concluded that they had never previously seen so many disadvantages to OSL dating as in the quartz grains from this site (B. Mauz, personal communication, 2006). This was probably due to low sensitivity (i.e. poor bleaching and dosing cycles) of the quartz grains or the bedrock lithology (Carboniferous sandstones). Lukas *et al*. (2007) encountered similar problems with samples from proglacial sediments in the northwest Scottish Highlands.

Farnley Haugh Scar cut-bank section: Location 1

Three OSL samples were taken sequentially up the face of Terrace 4 (Figure 5.19): (1)

FIGURE 5.19: Photograph of the main Farnley face showing the location of the 2 OSL sampling sites (X2734; LV164).

TABLE 5.1: OSL ages for the Tyne Valley.

LAB #	SITE /TERRACE	LATITUDE (N)	LONGITUDE (W)	ELEVATION (M OD)	DEPTH (M)	SAMPLE FACIES	OSL AGE (AND RANGE) (KA CAL. BP)
X2730	4Stones /Terrace 2	54°59'59"	02°11'20"	49.5	0.7	Sh	8.2ka±0.9 (9.1-7.9ka)
X2731	4Stones /Terrace 2	54°59'59"	02°11'19"	46	5.5	Sh	48.1ka±10.3 (58.4-37.8ka)
X2732	4Stones /Terrace 2	55°00'01"	02°11'14"	49	2	Sh	18.6ka±2.0 (20.6-16.6ka)
X2733	4Stones /Terrace 1	55°00'03"	02°11'05"	44	0.9	Sh	6.5ka±1.1 (7.6-5.4ka)
X2832	4Stones /Terrace 2	55°00'01"	02°11'12"	50	2	Fl	20.4ka±4.0 (24.4-16.4ka)
X2734	Farnley HS /Terrace 4	54°57'54"	01°59'47"	37	2	Sh	10.3ka±1.1 (11.4-9.2ka)

from the glaciolacustrine sands (Unit 1) (LV165); (2) from the interstratified bar top sands (Unit 2) (LV164); and (3) from the overbank sands at the top of the sequence (Unit 2) (X2734).

No luminescence date was possible from the basal glaciolacustrine sands due to similar sample problems encountered with those at Crawcrook. Sample LV164 yielded an age range of between 26 and 20kacal. BP, with an uncertainty of ±13%. This age is unreliable not only because of the absolute uncertainty associated with the estimate but also because in the geological context the sampled sands lie above the glaciolacustrine sequence, so should be younger than the LGM at 21ka cal. BP. The sample age most probably reflects incomplete bleaching prior to deposition. Sample X2734 yielded an age range of between 11.4 and 9.2cal. BP, with an uncertainty of ±10%. Even with this uncertainty, the age is accepted because other workers have accepted small uncertainties between ±2 to ±12% of the sample age (cf. Spencer and Owen, 2004) and the date is in the age range expected for development of terraces during the early Holocene. The formation of T4 suggests valley floor refilling continued into the early Holocene. The calcite sample yielded a U-series age range of between 26 and 6ka cal. BP. A 60% uncertainty meant the age was unreliable.

Fourstones cut-bank section: Location 1

Five OSL samples were taken from various units which underlie the terrace sequence (Figure 5.20). Sample X2731, taken from the glaciofluvial sands (Unit 2), which underlie the undifferentiated Terrace, yielded an OSL age range of 57.8-38.4ka cal. BP, with an uncertainty of ±21%. The age was rejected due to the absolute uncertainty being so large; uncertainties >12% are unreliable for constraining the timing of deposition, and it appears other workers do not accept ages with uncertainties above

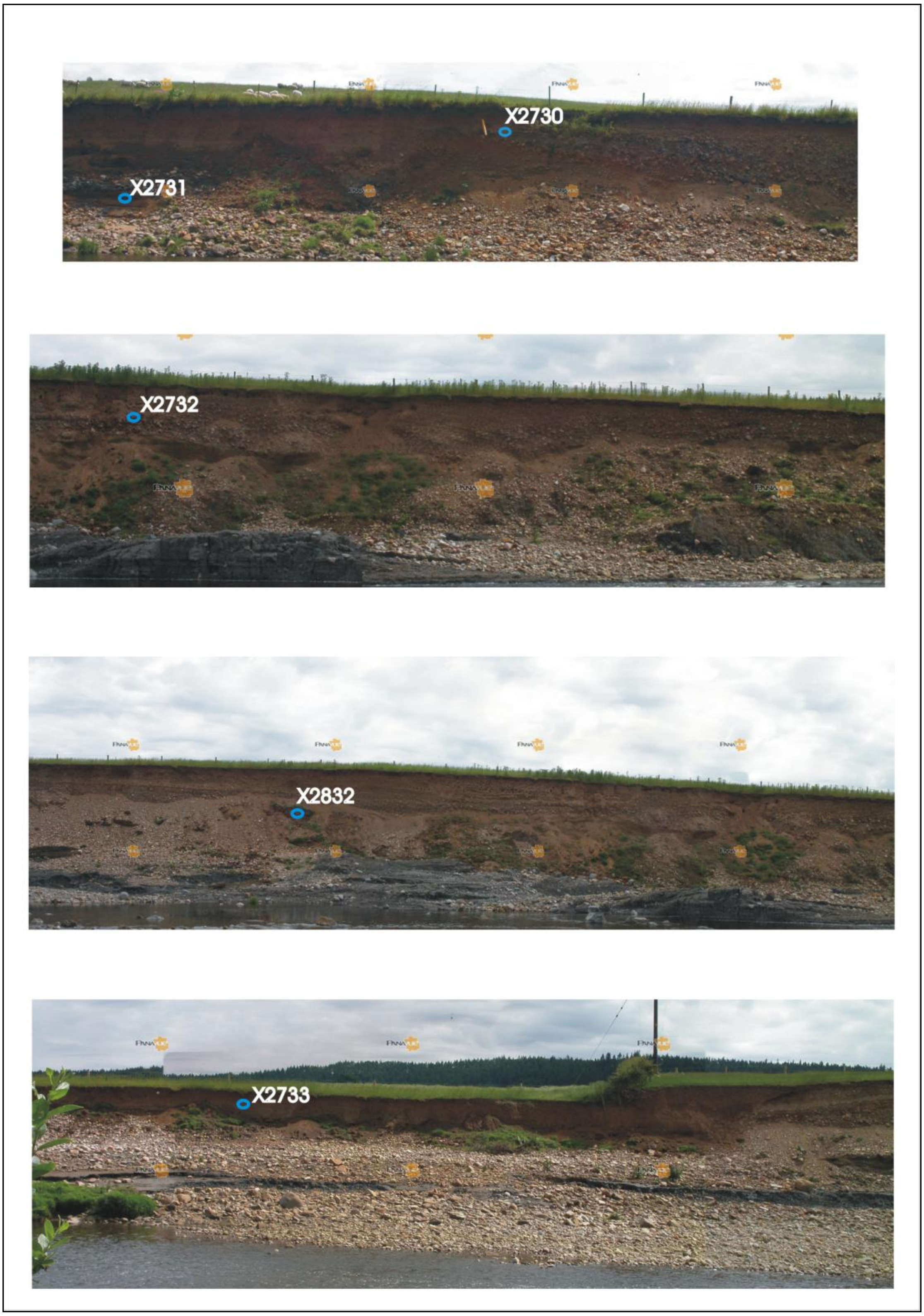


FIGURE 5.20: Photograph showing OSL sample locations(X2730-2733; X2832), and ages derived for the terraces (T2, T1) at Fourstones.

this limit (see section 4.5.1.4). Secondly, the age suggests deposition during MIS 3 and subsequent preservation of the sequence even though the region was completely inundated by ice during MIS 2. The age may be interpreted as incomplete zeroing (i.e. the grains retained a latent signal from previous exposure to light) of the sediment prior to redeposition.

Sample X2732 returned an age range of between 20.6 and 16.6kacal. BP, with an uncertainty of ~10%. Whilst the age is associated with an acceptable uncertainty, the geological context precludes it being acceptable. The sequence is not proglacial but a flood deposit, which suggests reworking and incorporation of earlier sediments. From this, it can be inferred that deglaciation in the Tyne Valley took place between 20.6 and 16.6ka cal. BP. Sample X2832 returned an age range of between 24.4 and 16.4ka cal. BP, and with a large uncertainty of ~19% this age was deemed unreliable. Sample X2730 returned an age range of between 9.1 and 7.9ka cal. BP, with an uncertainty of ~10% it was accepted, suggesting aggradation continued into the early Holocene. Sample X2733 returned an age range of 7.6-5.4ka cal. BP, with an uncertainty of ~16%. The large uncertainty suggests the age is unreliable. Independent age control obtained from below the equivalent terrace at Lambley (T4) on wood fragments yielded a ^{14}C age of between 1430-1100 BC, suggesting aggradation began after ~3.6 cal. yrs BP (Passmore and Macklin, 2001). Further work is needed to constrain the chronology for the Tyne Terraces, as single dates provide limited reliability.

5.3.1 Summary

There is no definitive chronological support for the glacigenic deposits, although they have been relatively dated to before 18ka cal. BP. Direct chronological control for the Tyne sequence is limited; the next sediments in the sequence are the river terraces, T4-

T1, which began aggrading after ~11.4ka cal. BP. From the two accepted dates for T4 and T2, it may be inferred that two cycles of valley floor incision and aggradation took place over a 4k year period during the early Holocene (11.4-7.9ka cal. BP). Whilst the age range of sample X2732 (20.6-16.6ka cal. BP), taken from Terrace 2 at Fourstones, is rejected from a geological context, it may suggest retreat and deglaciation in northern England was contemporaneous with that across the British-Irish Ice Sheet (see section 1.2.2). The mineral grains were reset during that period suggesting opportunities for bleaching occurred. However, further work on the sequence is needed to improve the chronological control and elimit erroneous ages.

5.4 Palaeoecological analyses

Samples taken for palaeoecological analyses contained no organic material. No pollen grains were found in any of the four samples taken from the coarse silt channel deposits of Terrace 2 at Fourstones. No beetle remains were found in the glacial lake sequence at Crawcrook (Location 2) or in the coarse silt channel deposits of Terrace 2 at Fourstones.

5.4.1 Interpretation

It is perhaps not unexpected that no organics were present in the pond deposit at Crawcrook associated with ice-marginal deposition. However, Coleopteran species are often found in association with palaeochannels, so the lack of finds is surprising. It suggests either conditions were not suitable for their habitat or there are preservation issues probably associated with acidic sediments.

5.5 GIS Visualisation of NEXTMap data: a combined field and remote sensing approach to landform mapping

Comprehensive landform mapping has not been carried out in the field area except for limited mapping in the early part of the last century, and more recently with only the Holocene valley floor receiving attention (cf. Passmore and Macklin, 1997; 2001). However, with renewed interest in mapping glacial landforms in Britain generated by the BRITICE project (cf. Clark *et al*., 2004a), and the availability of remotely sensed datasets at the appropriate scale, the present investigation into the landforms/landscape development of the Tyne Valley is timely. In terms of feature identification, spatial scale of representation relative to landforms being investigated is important. Features that form the focus of this investigation range in scale from 10^{-1}-10^{-2} km^2 (i.e. medium- to small-scale). Satellite imagery (e.g. SRTM, 90-30m grids) and 1:63000 scale air photos are suited to identification of large-scale glacial lineations (flutes, drumlins, bedrock grooves) although at this relatively coarse scale the quality of the analysis is limited and medium-scale features (e.g. ice-stagnation environments) are invisible (Napieralski *et al*., 2007). NEXTMap data, with a 5m grid resolution, was chosen for this study as medium-scale features are visible. This resolution allows investigation of local detail in the glacial history, which compliments the big picture mapping currently underway as part of the BRITICE project. Using GIS (ArcMap), the ‘raw’ NEXTMap data was converted into an output using visualisation techniques, from which information about the area could be deduced, and ideas about landforms and landscape development could begin to be produced and tested. Figure 5.21 summarises the methods that were used to analyse the NEXTMap data: stages 1 to 5 illustrate the steps that were taken to determine which the best approach to visualising the data was.

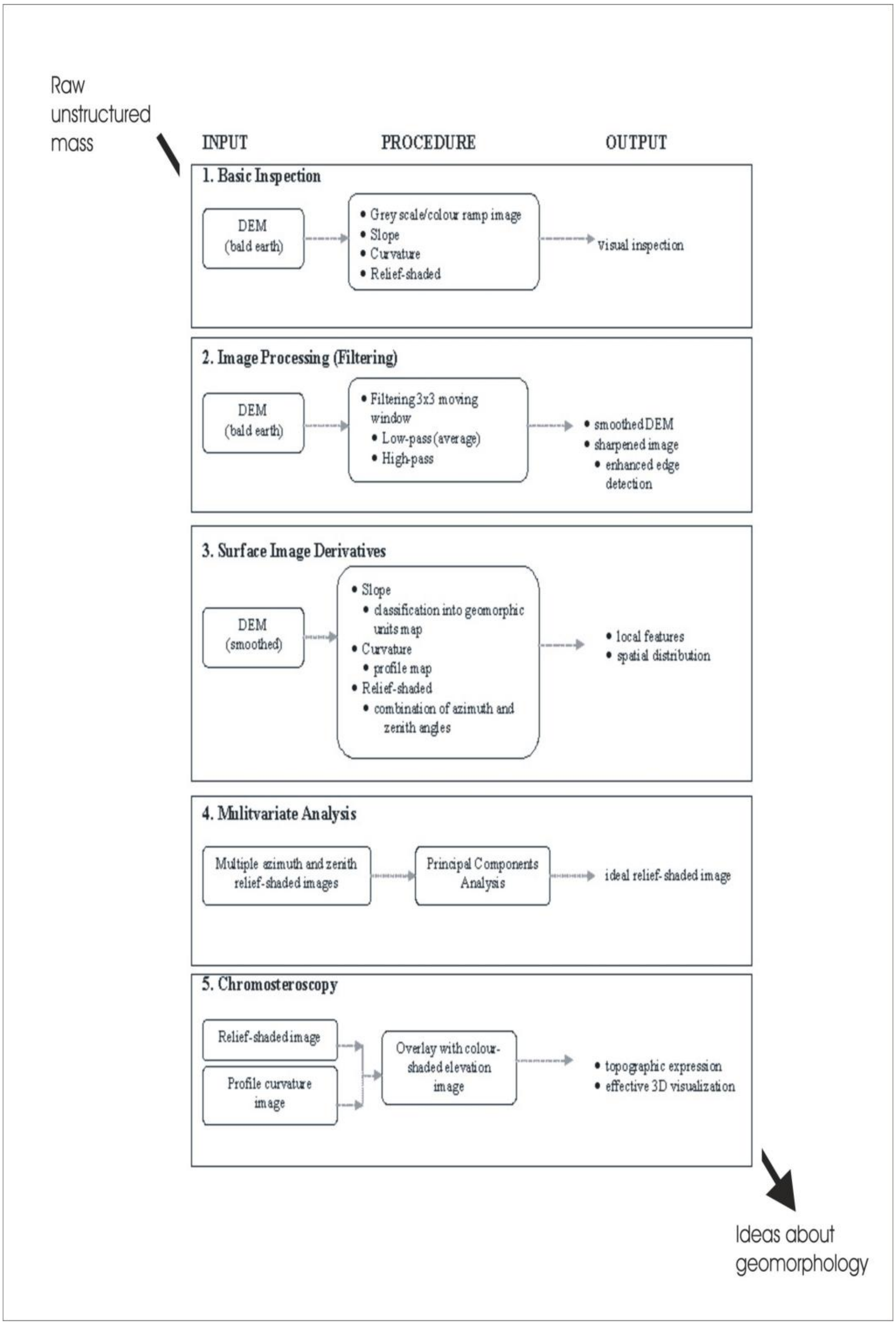


FIGURE 5.21: The analysis procedure for the systematic preparation of the NEXTMap data to enable identification of medium-scale morphological features.

5.5.1 Assessing the visualisation techniques

The first phase of the assessment was to evaluate the visibility of different landscape features in the visualisations. This was achieved by mapping a control area around Crawcrook and Stocksfield. The field map was used as the true and accurate record of the morphology of the control area against which the NEXTMap map was checked, and assessed through qualitative comparison.

5.5.1.1 Spatial filters

High- and low-frequency spatial filters offered no more information that the thematic visualisation model developed from the surface derivatives (Figure 5.21; Stage 3).

5.5.1.2 Surface derivatives of the NEXTMap data

By generating relief-shading, slope and curvature surface derivatives of the ‘raw’ NEXTMap data, it was possible to identify morphological elements of the landscape.

Relief-shading

Traditionally, air photo interpretation provides a stationary view of the landscape at a single azimuth and zenith angle, whereas relief-shading provides multiple options to analyse critically the data from any combination of observation direction and elevation In order to determine which relief-shading combination provided the best rendition of the data so that the landforms were easily identifiable (i.e. which enhanced rather than diminished landforms), a PCA analysis was applied to each azimuth angle combination model, to create a components image and determine which zenith angle contributed the most information to the image. The statistics generated for each analysis provided a clear indication of which zenith and which azimuth angle was contributing the most information to the image. In every combination, Principal Component 1, which is dominated by a zenith angle of 15˚, accounted for over 70 per cent of the total variance, and by an azimuth angle of 315˚ (i.e. northwest), accounted for 54.94% of the variance in the output image. Therefore, in terms of feature identification and mapping the ideal relief-shaded image in this context was illuminated from a northwest direction (315˚) and at a low elevation (15˚). The low zenith angle mimics winter sunlight levels, and landforms show up more clearly (Figure 5.22).

Curvature (change in gradient)

The curvature model of the area essentially highlighted the major features such as the river, its tributaries and the outline of the quarries; subtle features were more difficult to discern. The profile curvature map, which defines flow and therefore, influences erosion and deposition, highlighted the features of interest most clearly. For example, slope-breaks were easily recognised as light and dark edges in the visualisation model (Figure 5.23).

Slope (gradient)

The slope output model can be displayed either as a stretched or classified raster; a stretched output comprises a continuously coloured surface, which is stretched to optimise contrast (Figure 5.24), and a classified output displays the elevation data as a colour ramp dividing the data in a series of classes (Jones *et al*., 2007). Re-classification of the classified slope model into the geometric units (see chapter 4, section 4.2.3) highlighted the features more clearly by delimiting their outline through the break in slope. The classified slope map reveals that the steepest slopes are coincident with valley edges, demarcating incised valleys, and the quarry outlines. Continuous breaks of slope that parallel the river were associated with terrace edges.

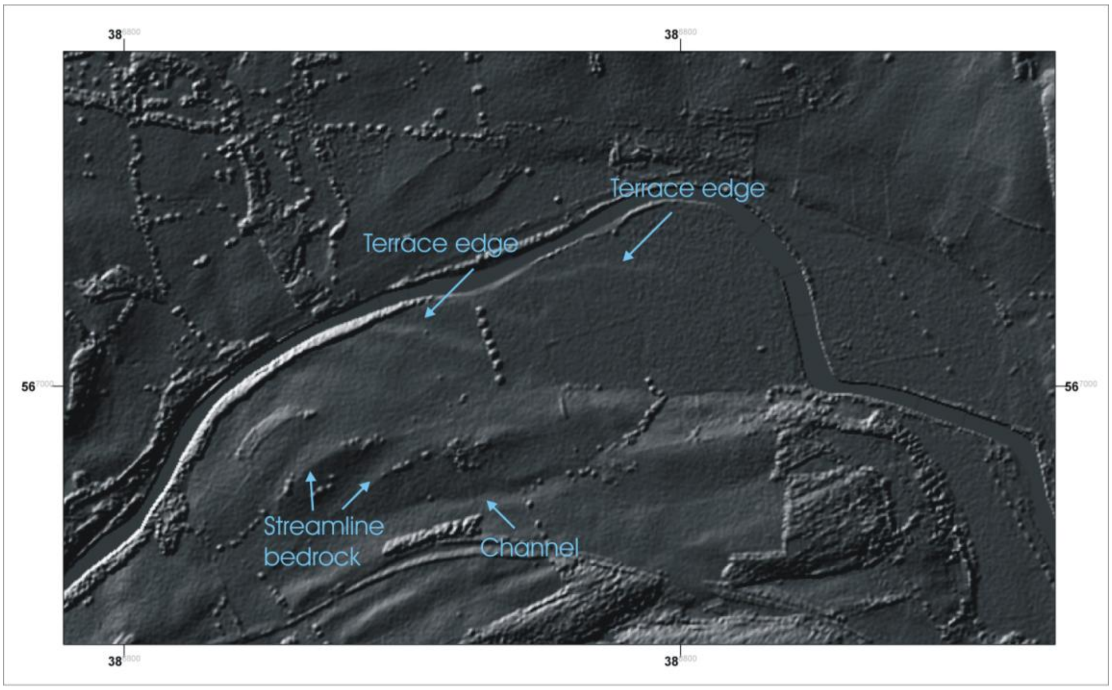


FIGURE 5.22: Shaded relief image of the NEXTMap (DSM) tile for the area around Fourstones. Labelled to highlight features identified.

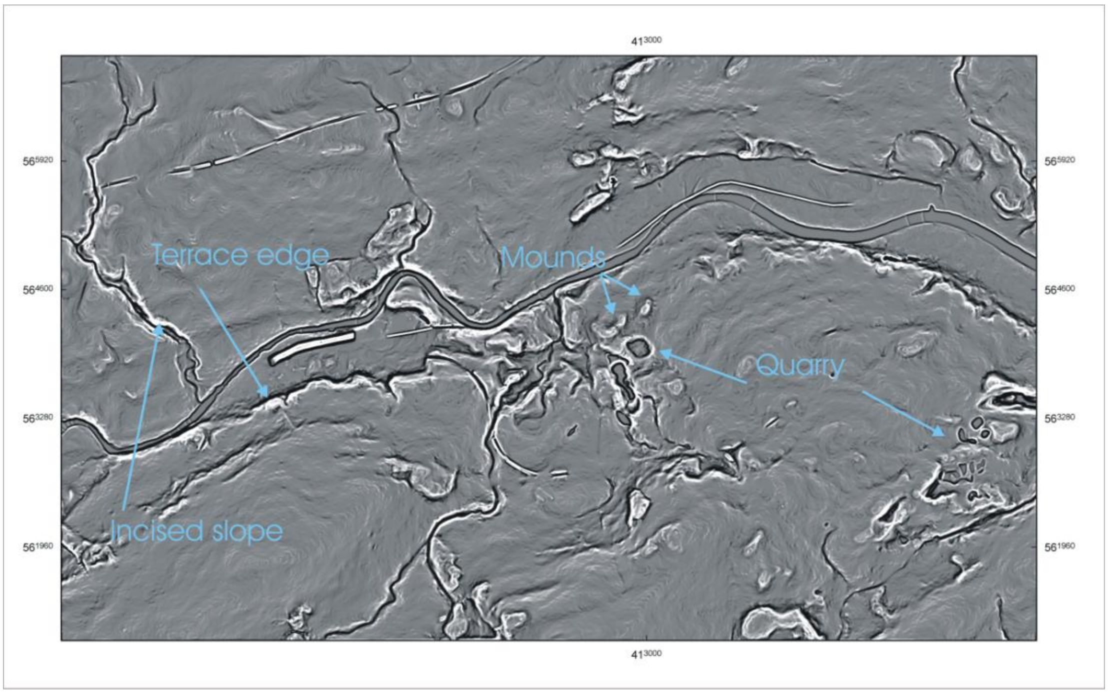


FIGURE 5.23: A profile curvature model of the NEXTMap data (DTM) for the area around Crawcrook. Slope breaks are represented by light (valley edge) and dark (valley base) lines.

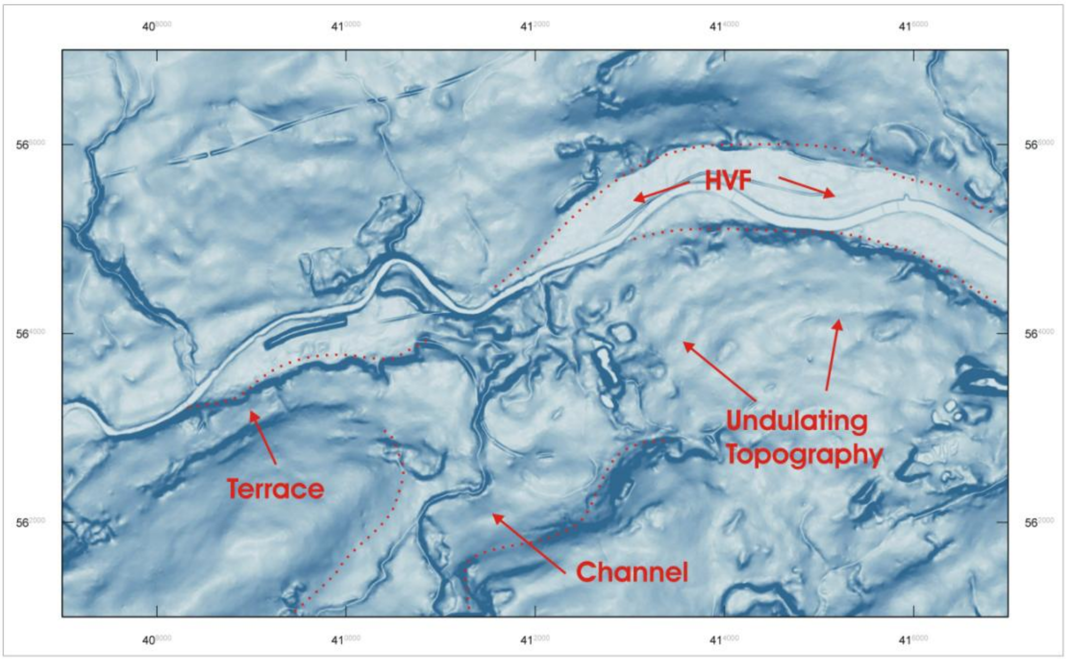


FIGURE 5.24: A stretched slope colour-shaded relief model at Crawcrook, which optimises topographic expression. Circular, ovate and irregular features, demarcated by dark outlines, represent the quarry workings. Annotated to highlight features, HVF is Holocene Valley Floor.

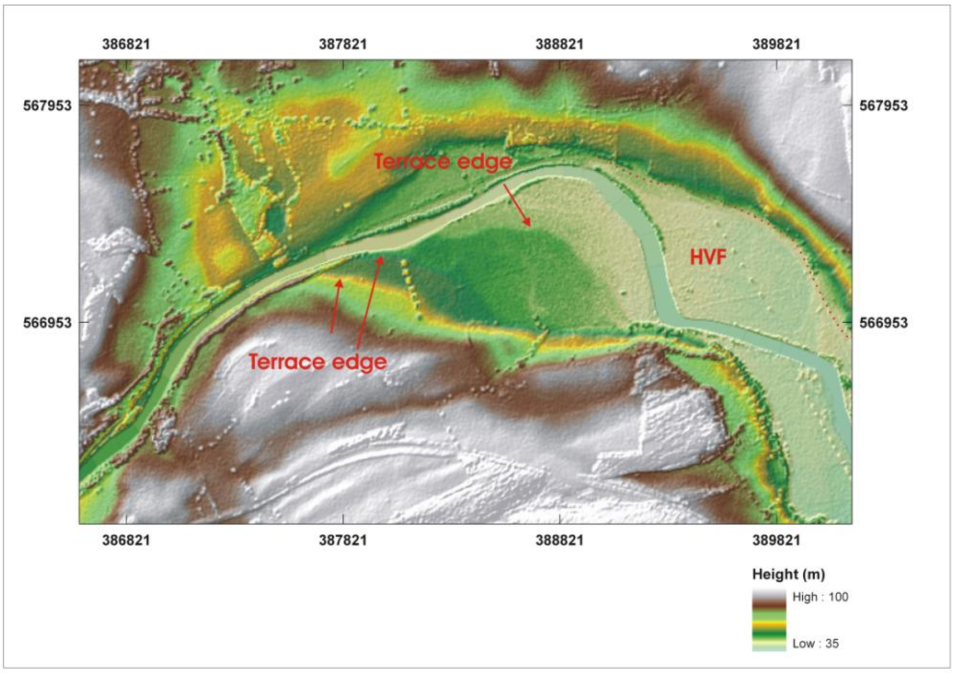


FIGURE 5.25: Colour shaded-relief model NEXTMap (DSM) tile of the area around Fourstones, annotated to illustrate the features mapped. Colour ramp set to reflect heights between 35 and 100 m, giving greater definition to features that fall between those height ranges to be mapped; above the upper range of 100 m, white represents all other elevations.

Chromosteroscopy

It was found that topographic expression was optimised and feature detection enhanced by overlaying the relief-shaded and curvature visualisation models with a colour-shaded elevation or stretched slope model to give a stereoscopic appearance to the landforms and creating a virtual landscape that had a 3D illusion to it. A colour-shaded thematic model is effectively a 3D visualisation, enhancing the chromosteroscopy (cf. Toutin and Rivard, 1995), giving a more realistic rendition to the image (Figure 5.25).

5.5.1.3 Summary

Mapping was undertaken in three stages. The colour-shaded relief model was examined first. From this model, landforms associated with deglaciation (Figure 5.25), and fluvial activity, were successfully highlighted. Secondly, overlaying the slope and curvature models with the colour-shaded elevation model was a useful tool for highlighting more subtle fluvial features, such as Holocene cut and fill units in the South Tyne Valley, and giving a three dimensional view to the surface. Finally, the colour-shaded relief model overlain with the mosaicked air photograph provided a useful aid for clarifying features identified on the DSM. The value of visualising both the surface and bald-earth NEXTMap was the ability to view the image where the vegetation/surface clutter had been removed, which may have been masking underlying topographic features. Air photos provided a useful overview of the area, rendering a realistic dimension to the topography. The subdued and subtle topography of the mounded elements of the landscape were more difficult to identify from the visualisations. By mapping the visualisations in ArcMap an area of valley floor ~ 80 km^2 was covered (Figure 5.26).

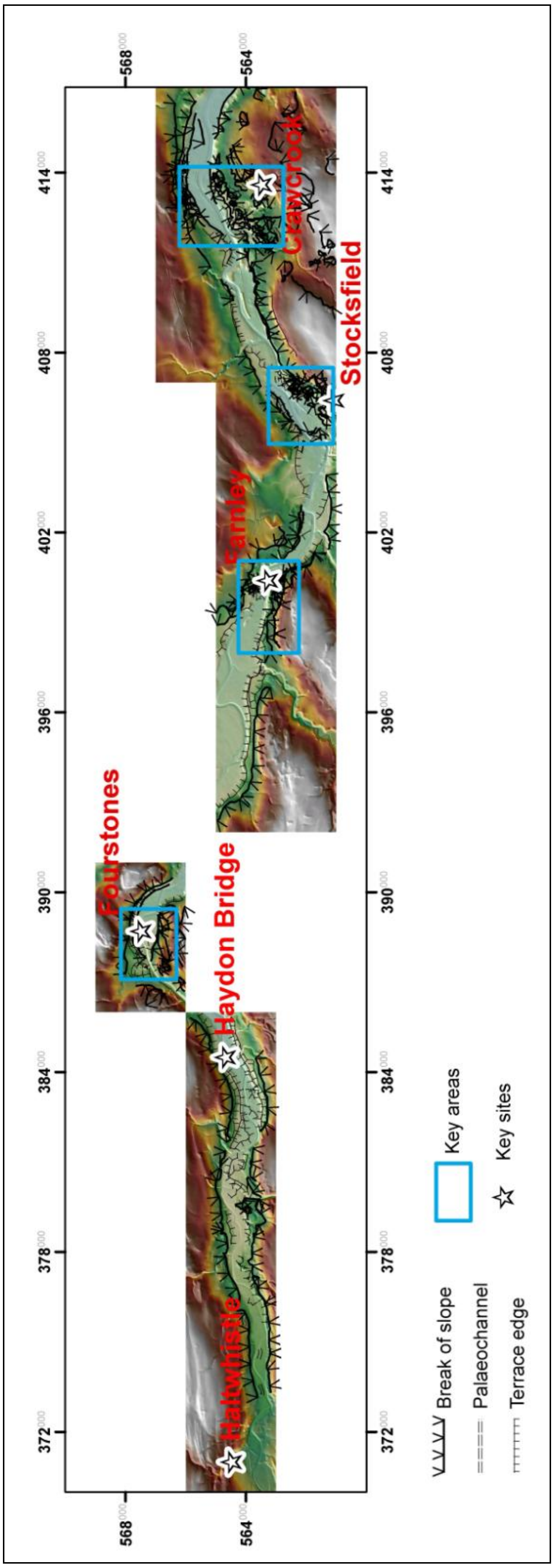


FIGURE 5.26: Illustration of the key areas mapped in the field and the area covered by visualisation of the NEXTMap data. See Figure 5.32 for more detail of the area mapped between Haltwhistle and Fourstones. See CD for full version of the NEXTMap geomorphological map.

5.5.2 Verifying the accuracy of the operator

With all techniques there must be some testing of both the methods used and the output derived. The NEXTMap geomorphology map was created through visual inspection of thematic models from which landforms were identified and the digitised. Both the field map and on-screen digitising was carried out by the author, therefore ensuring consistency and minimising operator error. The GIS method, like field mapping, still relies upon subjective interpretation, and the output is highly dependant on the skills and experience of the 'mapper' in terms of geomorphology and GIS. Although all the mapping was carried out by the author, and errors can be assumed to be minimal, a check for operator consistency and accuracy in terms of feature identification was carried out as the author's initial GIS skills were low. Therefore, the on-screen digitising was carried out twice, with a two week interval in between mapping, and the output compared by overlaying the resultant maps. It was found that the features identified during the initial mapping phase from the relief-shaded visualisation model were almost consistently identified during the second attempt. The skills of the author improved throughout the course of developing the visualisations and mapping, therefore, images were re-checked on more than one occasion.

5.5.3 Validation of the NEXTMap geomorphology map

The validation was performed to test whether the results of the NEXTMap geomorphological map provided reliable/accurate information on the geomorphology of the area under investigation. To assess how well the features could be identified from the thematic visualisation models, a qualitative comparison was made between the NEXTMap map and the field map using on-screen overlays to highlight any differences between features (delimited by polylines) identified in maps. It is worth noting that there is limited independence here, because both the NEXTMap map and field map were produced by the author and therefore, there will have been some personal bias in terms of feature recognition and interpretation. The two maps were overlain, to assess where the polyline symbols were coincident (though not an exact match) between each map and where they differed. The number of polylines identifying the same morphological feature was calculated with the result expressed as a percentage shown in Table 5.2.

TABLE 5.2: A comparison between features mapped using NEXTMap DSM data with those mapped using traditional field based approaches for the test area (Lower Tyne Valley).

FEATURE	TERRACE EDGE	MOUND OUTLINE	VALLEY EDGE
Field map	10	21	8
NEXTMap map	5	16	8
	50 %	76 %	100 %

The NEXTMap map, which resulted from careful analysis of the three visualisations of the NEXTMap data, had between 50 and 100% agreement with the field map. Table 5.2 indicates that there was a high degree of accord between the two maps in terms of valley edges and mound outlines. This is due to the fact that the valleys are highly incised, creating a major difference in slope and gradient of these features that is clear to see in the visualisations. The mound outlines were picked out with a relatively high success rate. However, it is clear that the subtleties (i.e. degraded edges, subdued expression) of these features cannot be as accurately delimited from NEXTMap data, and therefore, these features will always have to be checked in the field. The low success rate associated with terrace identification from the NEXTMap data is attributed to the fact that when field mapping, Holocene terraces (cut and fill) were visible to the author (who has considerable experience at field mapping) but are, in fact, relatively small-scale features in terms of their elevation i.e. ~1-2 m. Thus, they are not always visible on the NEXTMap data, which has a 5 m grid resolution. However, major terrace edges, which are delimited by significant breaks in slope (as is the case when field mapping), were picked up by the visualisation analyses. Comparison of the NEXTMap map and the field map indicate that the terrace edges and feature outlines correspond with major breaks in slope, identified in the surface derivatives of the NEXTMap data. These results confirm that the thematic visualisations of the NEXTMap data produce good quality renditions of the landscape, and that the map produced from them is an accurate though not a complete representation of the landforms on the ground. Thus, ground-validation of the maps produced is a requisite of this method.

Secondly, to assess the accuracy of the map produced from visualisation models of the NEXTMap data, a geomorphological map was produced (through identification of feature outlines) where no previous field mapping had been carried out by the author. The area selected for this assessment was the South Tyne Valley between Bardon Mill (NY 780 643) and Haydon Bridge (NY 829 655). Validation of the visualisation technique for mapping was then achieved by comparison of the NEXTMap, used as a base map, against the landforms in the field (Figure 5.27). Previous workers have compared feature identification from visualisation techniques to pre-existing field maps (cf. Smith *et al*., 2006), but at the time of the survey no one had employed the techniques to produce a morphological map from remotely sensed data (i.e. NEXTMap or LiDAR) to validate in the field. However, Jones *et al*. (2007) have just published a paper on geomorphological mapping using LiDAR and carried out such an assessment.

Through a walkover survey of the area, the location of the feature boundaries and landforms were verified by assessing whether the lines digitised from the thematic

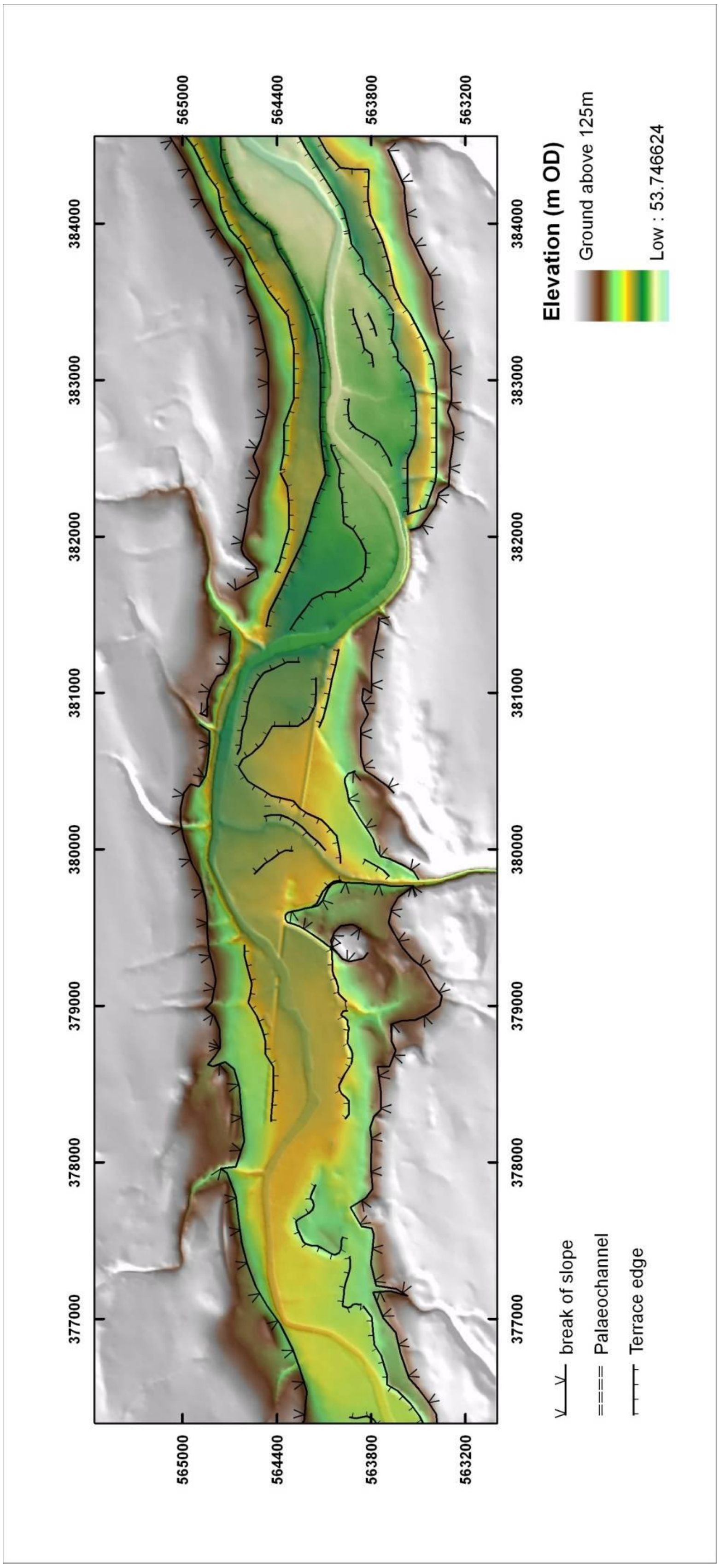


FIGURE 5.27: Geomorphological map of the area downstream of Bardon Mill, River South Tyne produced from visualisation of the NEXTMap data, overlying colour-shaded relief DTM. Terrace edges and break of slope have been delimited as black line symbols.

visualisation models were coincident with those landforms on the ground. The rigorous testing of the mapping method in the field gave a true evaluation of the mapping technique. Verification was carried out in the field for the South Tyne corridor. Where terrace remnants had been identified by visualisation there was very little discrepancy between the NEXTMap derived map and the features on the ground. Where discrepancies occurred, the landform was mapped in the field and the visualisation map subsequently updated. Thus, in terms of landform identification (particularly for river terraces) from visualisation of NEXTMap data, the geomorphological map was a success. In part, this is likely to be related to the prominence of the features as well as the methods used to visualise the landscape. The field check was successful in demonstrating the application of the method, and for the scale of landform under investigation, it was an appropriate approach to employ. Identification of small-scale features, i.e. Holocene cut and fill terraces, can be better mapped using LiDAR as demonstrated by Jones *et al*. (2007) due to the higher resolution (<1m) compared to NEXTMap.

In conclusion, visualisation mapping gives a close, but not complete, representation of the landforms on the ground. It has been shown that the visualisations techniques applied to the NEXTMap data have resulted in the production of an accurate geomorphological map with a high level of coincidence between features identified and those mapped in the field. Geomorphological mapping produced from visualisation models of the NEXTMap data have been successful. Using both methods (visualisation and field verification) the final output was a geomorphological map of the landforms along the Tyne Valley. The advantage of this approach over traditional field mapping is that it enables the amount of time spent in the field to be minimised whilst still producing high quality information. Validation of the NEXTMap map, over an area *c*.16km^2, took less than a day to check in the field. Given the excellent spatial coverage of remotely sensed DEMs now available for the UK, visualising the landscape and mapping large areas can be achieved relatively easily, cheaply and quickly. Finally, visualising the landscape through remotely sensed DEMs, can give access to areas that may otherwise be difficult to gain entry to on the ground.

DEM visualisation is about developing big ideas about landscape evolution. Whilst it is not intended to replace traditional field mapping, its contribution in landscape studies and understanding broad scale development is invaluable. Comparison with other studies (e.g. Smith *et al*. 2006; Jones *et al*. 2007) indicates that the approach used to visualise the data adopted here is becoming a standard and that in similar terrain, NEXTMap data is the appropriate scale to view the landscape and map features. A combination of NEXTMap, air photo and field mapping is recommended to provide the best source of information for data capture of medium-scale glacigenic landscapes that will result in the production of an accurate geomorphological map.

5.6 Landform assemblages along the Tyne Valley

Through the combination of field mapping, DEM visualisation and air photo interpretation, the landform assemblages that crop out along the Tyne Valley have been identified (Figure 5.28A,B) The assemblages represent proglacial and postglacial development. Outside the study area, individual glacial landforms have been identified by other workers (cf. Lunn 1980; Douglas 1991). Proglacial landform assemblages are most prominent in the lower Tyne Valley, in the vicinity of the towns of Crawcrook and Stocksfield. The postglacial landform assemblage comprises the Lateglacial and Holocene valley floor and occurs along the length of the research area between Haltwhistle and Crawcrook.

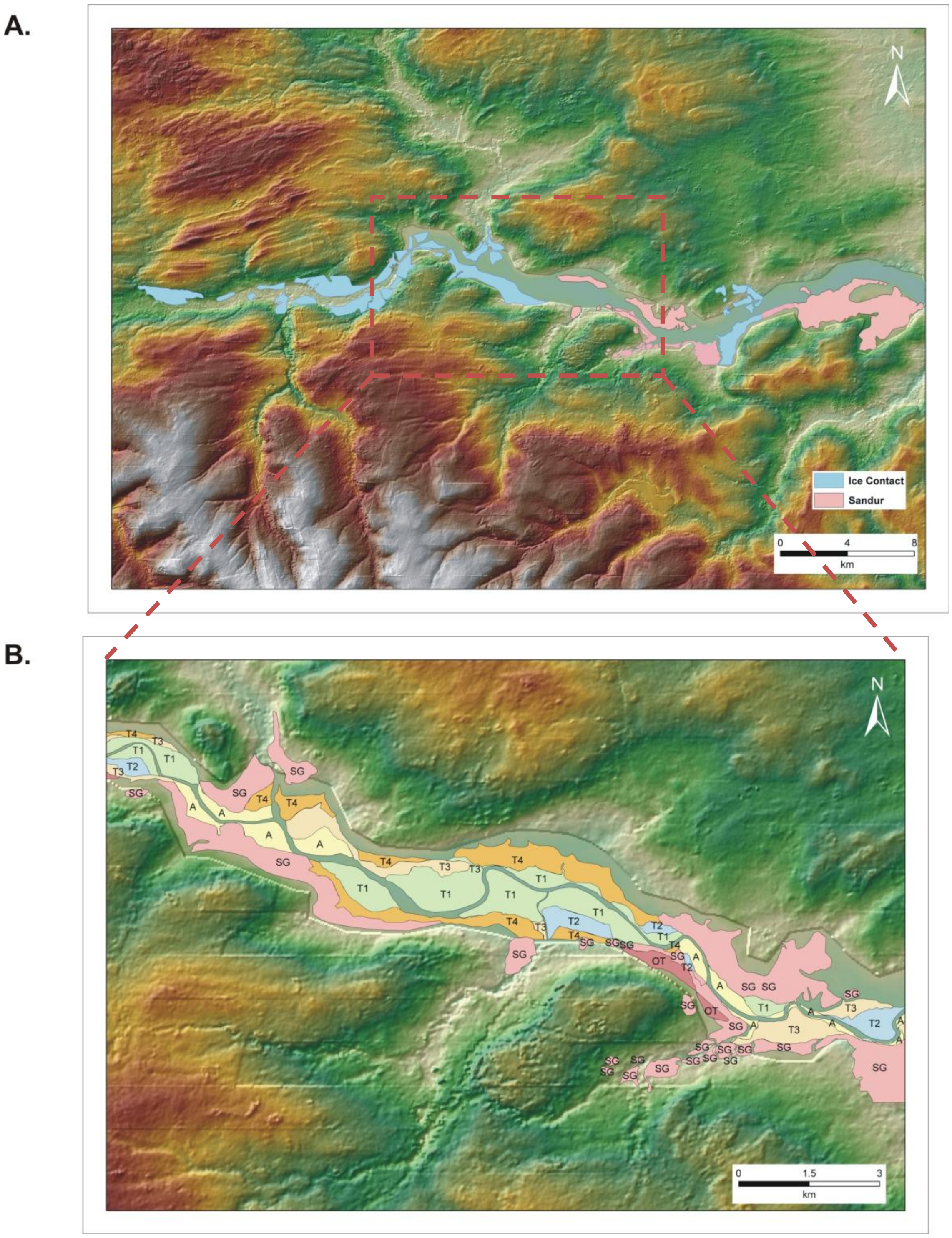


FIGURE 5.28: A. Map of sediment-landform assemblages along the Tyne Valley. B. Surficial geology map of part of the Tyne Valley (highlighted by the red box in A). A - Alluvium, T1-T4 - Terraces 1-4, SG - Glacial sand and gravel, OT - Outwash Terrace. (Background is colour-shaded relief DEM, 25m grid, source: www.landmap.ac.uk).

5.6.1 Glacial landforms

The most prominent glacial landform assemblage comprises a pattern of subglacial bedforms (large-scale streamlined glacial landforms) in the northwest area of the Tyne Basin, between the South and North Tyne Valleys. No bedforms are visible down-ice of this area. These streamlined landforms (drumlins) occur in bedrock and superficial deposits. Two ice stream flow sets can be identified: (1) from Kielder (NY 818 769) to Wark (NY 857 773) in a south easterly direction and (2) from the Tyne Gap to Warden in an east-north easterly direction (Fig. 5.29A). Large meltwater channels occur between the drumlins, which are up to 2.5km long (cf. Frost and Holliday, 1980). The flow sets are convergent in the Wark area and terminate before the present day Tyne confluence is reached. The bedrock (Figure 5.29) and the superficial features indicate different periods of ice flow dominance, and the flow sets in this area are currently under investigation at the University of Durham. These large-scale features are interpreted as glacial lineations and relate to the flow direction of palaeo-ice streams. Ice streams are associated with a zone of fast flowing ice, scouring the bedrock and producing subglacial bed deformation. The landforms reflect fast ice flow, which may be interpreted as a response to rapid climate change during deglaciation, but also could simply reflect internal ice sheet dynamics. However, these features were not investigated as part of this study. Evidence of glacial smoothing of the landscape is apparent throughout the main Tyne Valley. On the north and south side of the lower Tyne Valley between Crawcrook and Stocksfield, and along the South Tyne Valley between Haltwhistle and Fourstones, there is indication of ice smoothing and moulding of the underlying sediments at the ice/sediment bed interface in the form of small-scale fluting, smoothed slopes and surfaces.

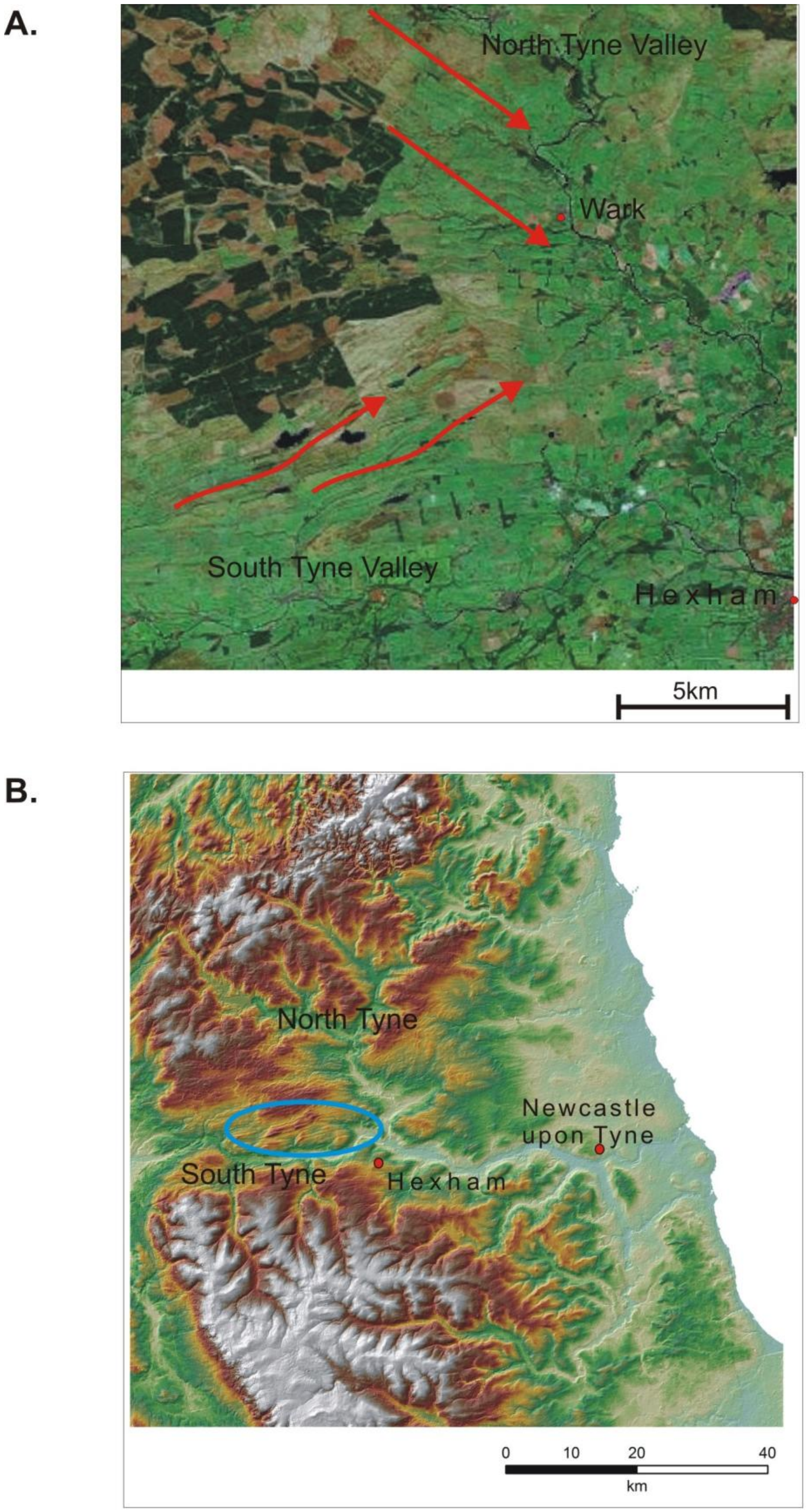


FIGURE 5.29: Glacial lineations (palaeo ice-streams) in the North and South Tyne Valleys visible on both A. air photos (red arrows; source: www.multimap.com) and B. a 25m grid scale relief-shaded DEM (blue ellipse; source: www.landmap.ac.uk).

Subglacial drainage is evident from nye channels (cf. Benn and Evans, 1998) that arise on the flanks of the valley and across the high ground, sometimes crossing watersheds. At Fourstones, a nye channel ~1.5km in length has been incised into the bedrock. Southwest of Stocksfield, a spectacular meandering meltwater channel, flowing west-east, is traceable over ~2km (Figure 5.30A). To the southeast of Stocksfield towards the Derwent Valley, further examples of channels can be identified, such as the Beldon or Whittonstall channels (Peel 1949; Sissons 1958). Sissons (1958, 1960) identified and interpreted these features as subglacial meltwater channels, representing drainage route ways within the ice, which were eventually lowered onto the surface as the area deglaciated.

There is an absence of identified moraines in the Tyne Valley. The absence of moraines suggests stagnation by *in-situ* downwasting. The presence of till deposits between the glaciofluvial and glaciolacustrine sediments suggests stagnation zone retreat (cf. Currier, 1941), with an extensive zone of stagnant ice leaving behind glaciolacustrine and outwash deposits, and limited active ice depositing till, in some cases over the outwash deposits.

5.6.2 Proglacial landform assemblages

Due to the dynamic nature of the proglacial environment and subsequent river valley development, preservation of intact proglacial landform assemblages is low, and they remain as dissected remnants of sandur and lacustrine deposits. The Crawcrook complex (Figure 5.31) is exposed on the southern flank of the lower Tyne and is traceable further east and southeast into the Derwent Valley. Much of the surface morphology has been destroyed by extensive quarrying of the sands and gravels over the last 40 years but the complex extends over 24 km^2, forming a continuous landform-

A.

B.

FIGURE 5.30: A. Glacial channel location (denoted by blue ellipse) in the Tyne Valley, southwest of Stocksfield, identified from air photo analyses. B. Subglacial geomorphology (esker?) in the Derwent Valley, southwest of Greenside, identified from air photo analyses (denoted by blue ellipse). (Air photos from www.multimap.com).

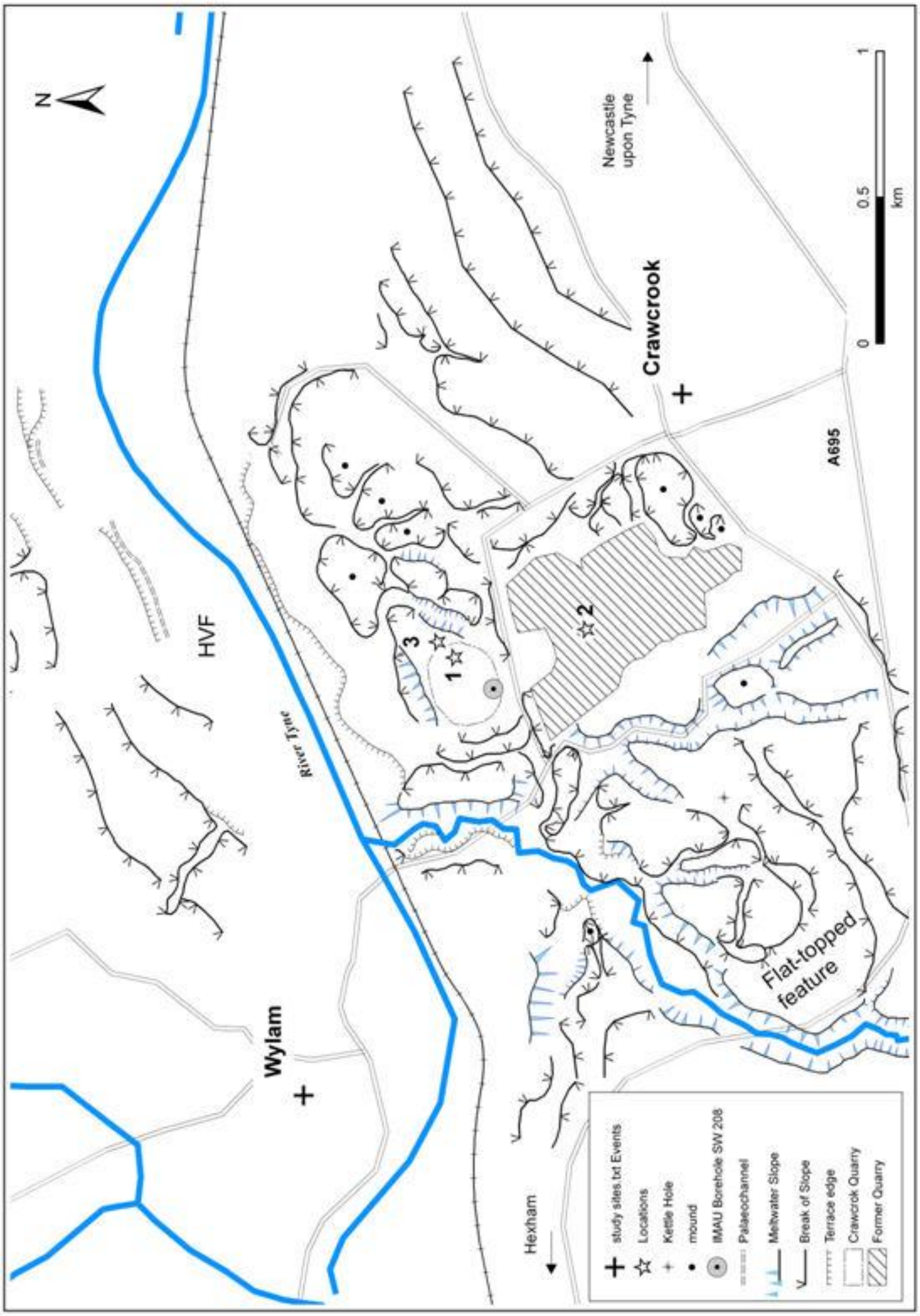


FIGURE 5.31: Geomorphological map of the area around Crawcrook. It shows the locations of the exposures recorded, the arrangement of mound topography, and distribution of meltwater channels. The flat-topped surface and ribbed moraine mentioned in section 5.6.2 are also indicated. HVF denotes Holocene valley floor.

sediment assemblage around the towns of Crawcrook, Greenside and Blaydon. The sediment body is in excess of ~30m thick, but varies locally. The topography is characterised by a series of subdued, irregular and low-amplitude mounds, dissected by channels. Previous workers have interpreted the undulating mounds of the Crawcrook complex to be the product of esker and ice-contact (kame) deposition (Francis 1970, 1975; Lunn 1980) or through deposition by ephemeral, subglacial streams (Mills and Holliday, 1998). Based on the sedimentology (section 5.2) the main body of the complex is interpreted here as a proglacial sandur that was subsequently dissected and eroded, forming the mounds. Other features of the complex include flat-topped terraces, kettleholes and ridges, surface channels and incised valleys. To the south of Greenside, there is a distinct group of small, sinuous ridges which bifurcate and anastamose over <1km (Figure 5.30B). This localised morphology within the complex may be Francis's (1970) esker but there was no access/exposure to these features or their internal sedimentology.

Approximately 0.5km to the west of Crawcrook, a dissected, channelised flat-topped deposit occurs ~3km in lateral extent. Its surface lies at ~59m OD, and it is dissected by two steeply incised channels, now occupied by the Stanley Burn and Bradley Dene. Smaller channels flow across its surface, possible indicating development as a delta top associated with the lake, and the larger, incised channels, which dissect its outer flanks, probably occurring after the surface was abandoned. Its surface slopes gently northeast towards the current quarry. A few possible kettleholes can be seen on its surface and across parts of the wider complex, suggesting proximity to ice. Such a laterally extensive, flat area is indicative of a deltaic surface or subaqueous fan (Figure 5.31), and given the close associated with lacustrine sedimentation recorded in the quarry this is the most likely interpretation as subsurface data is limited (Figure 5.6). Given the elevation of the lacustrine deposits, the lake level must have been, at least, 60 m OD. Based on the extent of the landform-sediment assemblage, it is envisaged that the lake was localised, impounded to the east by ice in the Tyne Valley, with an irregular, retreating ice margin along the North Pennines to the southwest. The location of the ice margin is speculative as ice contact slopes were not recorded but the proximity to the ice is clearly indicated in the sedimentary record.

A smaller proglacial landform assemblage has been mapped further west around Stocksfield, and the complex extends over ~2 km. The topography comprises a series of subdued, subparallel ridges and mounds cut and degraded by channels. The mounds reach an altitude of 55 m OD but are smaller in areal extent than those around Crawcrook. The mounded landforms are abruptly terminated by the Holocene valley floor. This assemblage is contiguous with the larger Crawcrook complex, but has been differentiated as ice-marginal rather than proglacial on the basis of sedimentology (see section 5.2). To the east of Stocksfield, the mounds are inset below a major cut slope that forms part of the valley side (Figure 5.32). The mounds have been incised by meltwaters, flowing off the valley side in both small and large channels. On the north valley side, west of Ovingham (NZ 082 636), the IMAU report (Giles, 1981) delimits outcrops of sand and gravel. Morphologically, these features are indistinct but form a discrete landform assemblage that extends over ~4 km^2. Given the lack of sedimentological detail, interpretation is difficult, although their disconnected nature suggests formation as ice marginal sediments, particularly as they are bounded by till, confirming ice cover. Due to their occurrence in close association with the ice-marginal sediments identified at Stocksfield, it is tempting to assign the same interpretation, but there is no borehole evidence to confirm this. The subaqueous deposits and ice marginal topography at Stocksfield indicate the presence of an ice-

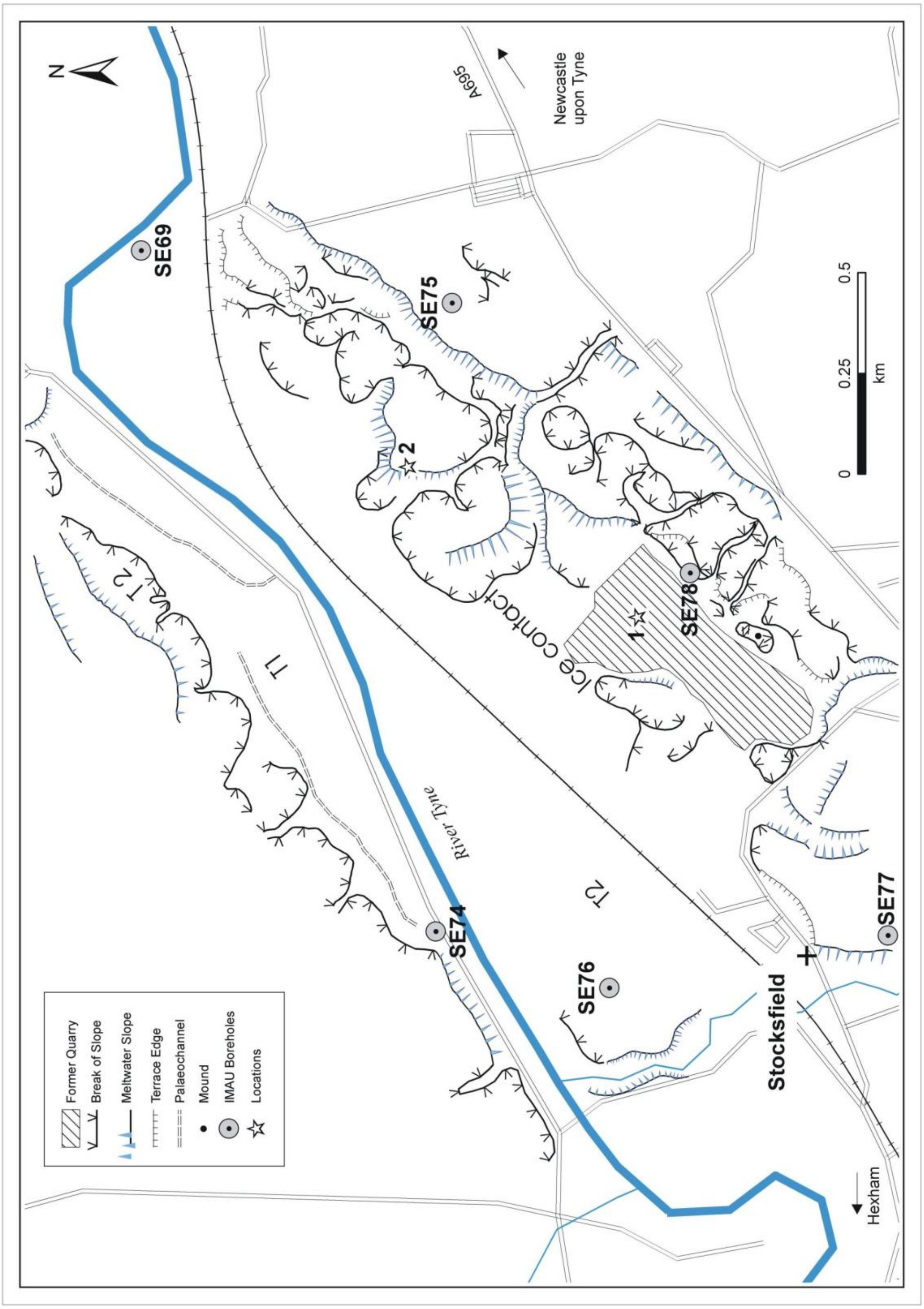


FIGURE 5.32: Geomorphological map of the area around Stocksfield. It shows the locations of the exposures recorded, the extent of the undulating topography, the cut-slopes and the degraded edges of the mound landforms. Extensive postglacial valley floor reworking is indicated due to the lack of features.

margin. The subglacial channels identified in the vicinity of Stocksfield would have delivered meltwater and sediments into the Tyne valley. Localised ponding of meltwater is probably associated with active ice situated in the lower Tyne Valley forming an ice dam as the ice downwasted and temporarily preventing drainage (Smith and Ashley, 1985).

On the north side of the valley at Farnley, mounded landforms associated with dissection by meltwater channels outcrop behind the terraces (Figure 5.33). The landforms are associated with an extensive spread of glacigenic sediments (Lovell, 1981), characterised by mounds up to 55 m OD, surrounded by flat terrain ~35-40 m above the present river level. Between Bywell (NZ 049 617) and Thornborough Haugh (NZ 004 635) the landform-sediment assemblage is ~4 km long and up to 1 km wide. The sediments lie against the high ground to the north and are curtailed on the river side by a major scarp slope. There were no exposures in the mounds at Thornborough Haugh and the quarry workings were being landfilled, thus the underlying sedimentology could not be examined to determine the genetic origin of the features. On the south side of the River Tyne above Farnley, an extensive flat-topped terrace with a steep bluff, ~1.4 km in length, was identified between Dilston Vale (NY 979 633) and Farnley Gate (NY 999 630) and can be traced downstream to Riding Mill (NZ 019 617), where it pinches out. The surface of this upper terrace lies at ~70 m OD, ~40 m above the modern river level. Its surface has been dissected and degraded by small channels, probably both penecontemporaneous and post-abandonment, and it has a prominent bluff, which is interpreted as fluvial down-cutting. This terrace was not identified by the IMAU survey (Lovell, 1981), however, here it is identified as a feature of the proglacial landform assemblage that extends around Farnley. No IMAU boreholes were recorded through the landform assemblage, thus it was not possible to

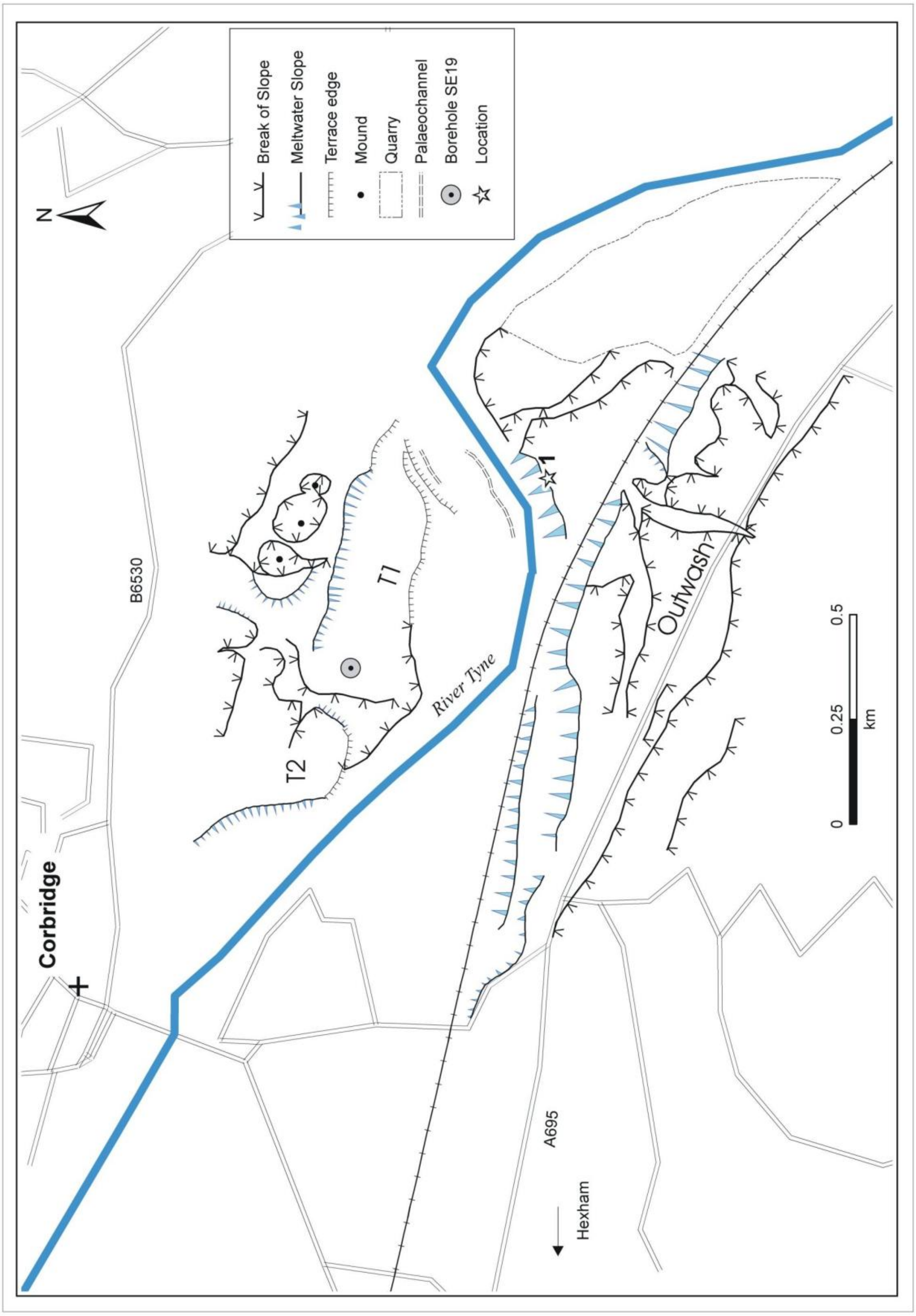


FIGURE 5.33: Geomorphological map of the area around Farnley Haugh Scar. The location of the cut-bank exposure is denoted by location 1. Note the intricate dissection of the outwash surface.

determine whether it comprises glaciofluvial or glaciolacustrine sediments. However, ground survey of the recently ploughed surface revealed well-rounded cobble and pebble gravel (Lake District igneous, Shap granite) near the surface. The clasts showed evidence of glacial weathering (i.e. striations, flat-iron shape), reflecting reworking or out-washing of glacigenic material. The landform-sediment assemblage is interpreted to be an outwash complex (see section 5.2.3), and the sandur has been preserved as an outwash terrace (OT) on both sides of the river. The mounds are the erosional remnants of the outwash complex, similar to those associated with Crawcrook and Stocksfield. Incision and dissection of the sediments to form the outwash terrace probably occurred as the proglacial river responded to changing conditions at the ice front as deglaciation continued.

Within the mid Tyne Valley, near Aydon (NY 998 670), the sedimentary sequence recorded from borehole NE40 (cf. Lovell, 1981) has been interpreted as an outwash fan deposit. Nearby there is evidence of laminated clays and peat basins (cf. Mills and Holliday, 1998). Taken together, this sequence has been interpreted as subaqueous outwash into a lake and ice mass wastage providing further sedimentary evidence for temporary lake development as part of the deglacial process.

Landform assemblages are more discrete along the South Tyne Valley. The area between Warden (NY 910 667) and Haltwhistle (NY 709 645) is characterised by a subdued, undulating topography, and although mounds are present, they occur as localised hills (Figure 5.34). The glacigenic assemblage drapes the till covered valley sides high above the contemporary river, which has become entrenched within its river terraces. Between Fourstones and the Allen confluence, the assemblage is almost continuous on both sides of the river; though upstream of this confluence the

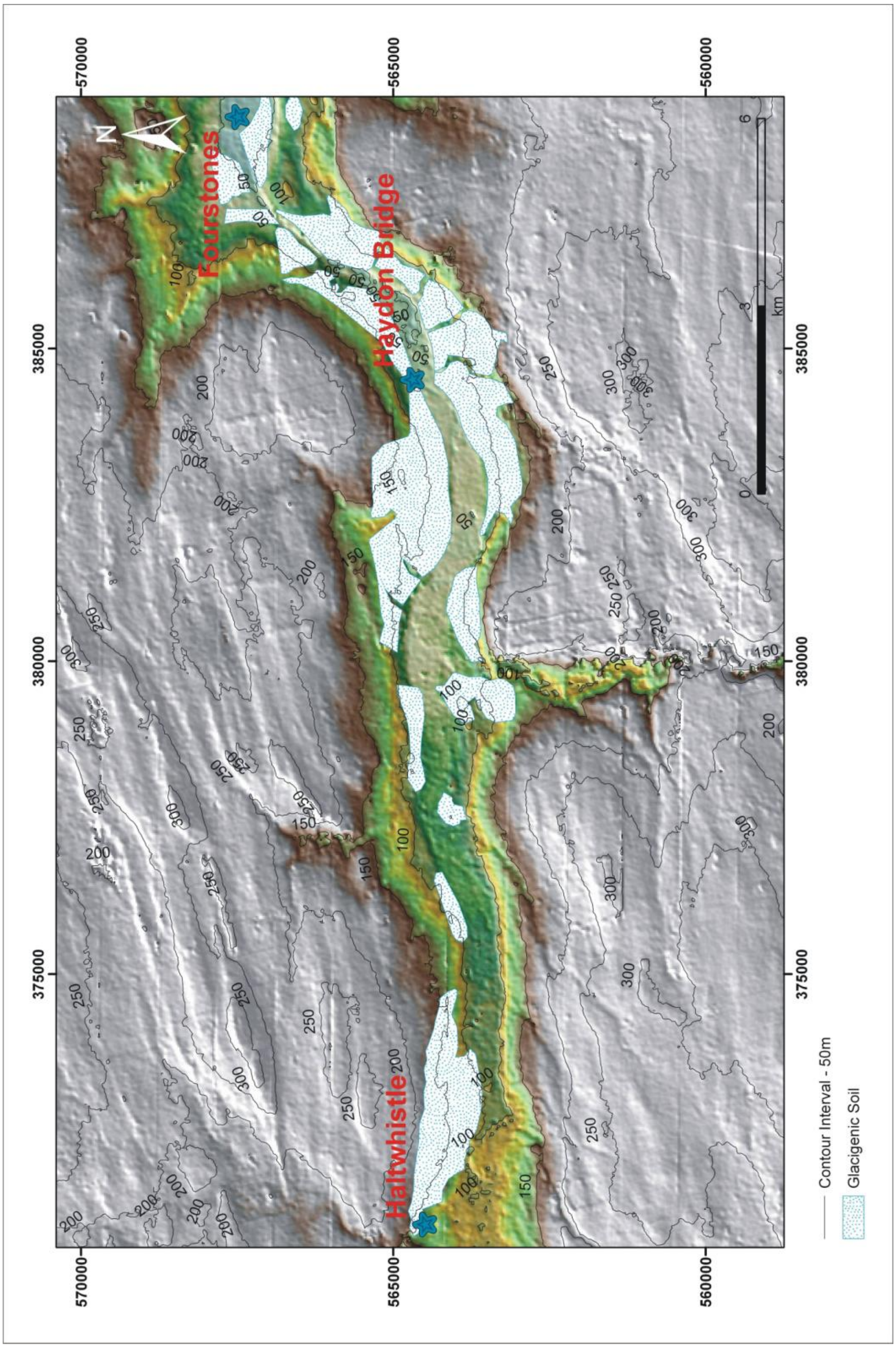


FIGURE 5.34: Map showing the lower South Tyne Valley between Haltwhistle and Fourstones and the distribution of glacigenic soils (identified by Jarvis, 1983). Colour-shaded relief DEM (25m grid) is landmap (source: www.landmap.ac.uk). Colour ramp set to reflect heights between 35 and 150 m OD.

assemblage becomes disjointed. In the vicinity of the North/South Tyne confluence, the area comprises localised mounds, rising to ~100 m OD and associated with a kettlehole.

Sedimentologically, these features comprise sands and gravels (see section 5.2; Figure 5.18), which appear chaotically deposited. King (1976) suggested that these deposits represent an esker, although both Lovell (1981) and Lunn (1995) interpreted the deposits as ice-marginal, rather than subglacial. Part of this assemblage comprises a flat-topped surface on the south side of the valley at Fourstones, lying at 68 m OD, approximately 35 m above the present river bed, and was interpreted as an outwash terrace in this study. There were no subsurface records for this terrace, though ground survey of the terrace surface following ploughing revealed well-rounded cobble gravels, and the soil survey (cf. Jarvis, 1983) interpreted the soils as glacigenic. This terrace might be correlated with similar terraces identified in the North Tyne Valley at an elevation of 30 m above river level (cf. Peel, 1941), but the latter terraces have not been investigated as part of this study. This landform assemblage is delimited to include the area to the west of Fourstones up to Haltwhistle.

Without detailed sedimentological data, it was impossible to be absolute about the genetic origin of the assemblage. However, based on the limited evidence from the IMAU survey boreholes (cf. Lovell, 1981) and field survey at Fourstones, the assemblage is interpreted as an ice-marginal complex, comprising glaciofluvial outwash and glaciolacustrine deposits resulting from localised ponding. The complex represents deposition along the South Tyne Valley as the ice retreated and stagnated (Figure 5.34). The complex was subsequently incised, probably towards the end of deglaciation, by meltwaters discharging from the ice front across the watershed reworking some of the lower deposits to form the indistinct outwash terraces and mounds.

5.6.3 Postglacial landform assemblages

These comprise the contemporary floodplain bordering the River Tyne and the Lateglacial/Holocene river terraces that are the incised remnants of former floodplains and/or glacigenic sediments, and extend the length of the research area (Figure 5.28B; Figures 5.33 - 5.39). The floodplain is omnipresent and bounded by terraces. Four distinct terrace levels have been identified along the length of the Tyne Valley between 20 and 4 m above present river level. These have been referred to as Terraces 1-4, where T4 is the highest. The terraces are preserved as both laterally continuous and fragmentary elements, and are most continuous along the South Tyne and mid Tyne Valley. The major fluvial terraces are the upper ones (T3 and T4), with the other two (T1 and T2) being relatively minor in terms of areal extent and height above river level. The major terraces are laterally extensive, semi-continuous and paired between Melkridge (NY 737 645) and Farnley Haugh Scar, below which they become fragmentary.

Terrace 4

The highest terrace (T4) is most extensive and its surface lies, on average, 20 m above present river level and, ~10m below the surface of the glacigenic landforms. The elevated position of this terrace above the present river level would suggest it is most likely to be Lateglacial in age in accord with similarly elevated terraces recorded elsewhere in northern England (cf. Tipping 1995; Howard *et al*. 1999). In the South Tyne at Fourstones, the surface of T4 is at 69m OD, ~20 m above the present river bed (Figure 5.35). Well-rounded gravel (sandstone) is present at the surface but no exposure is available in the cut-bank section. In the mid Tyne at Farnley, the surface of

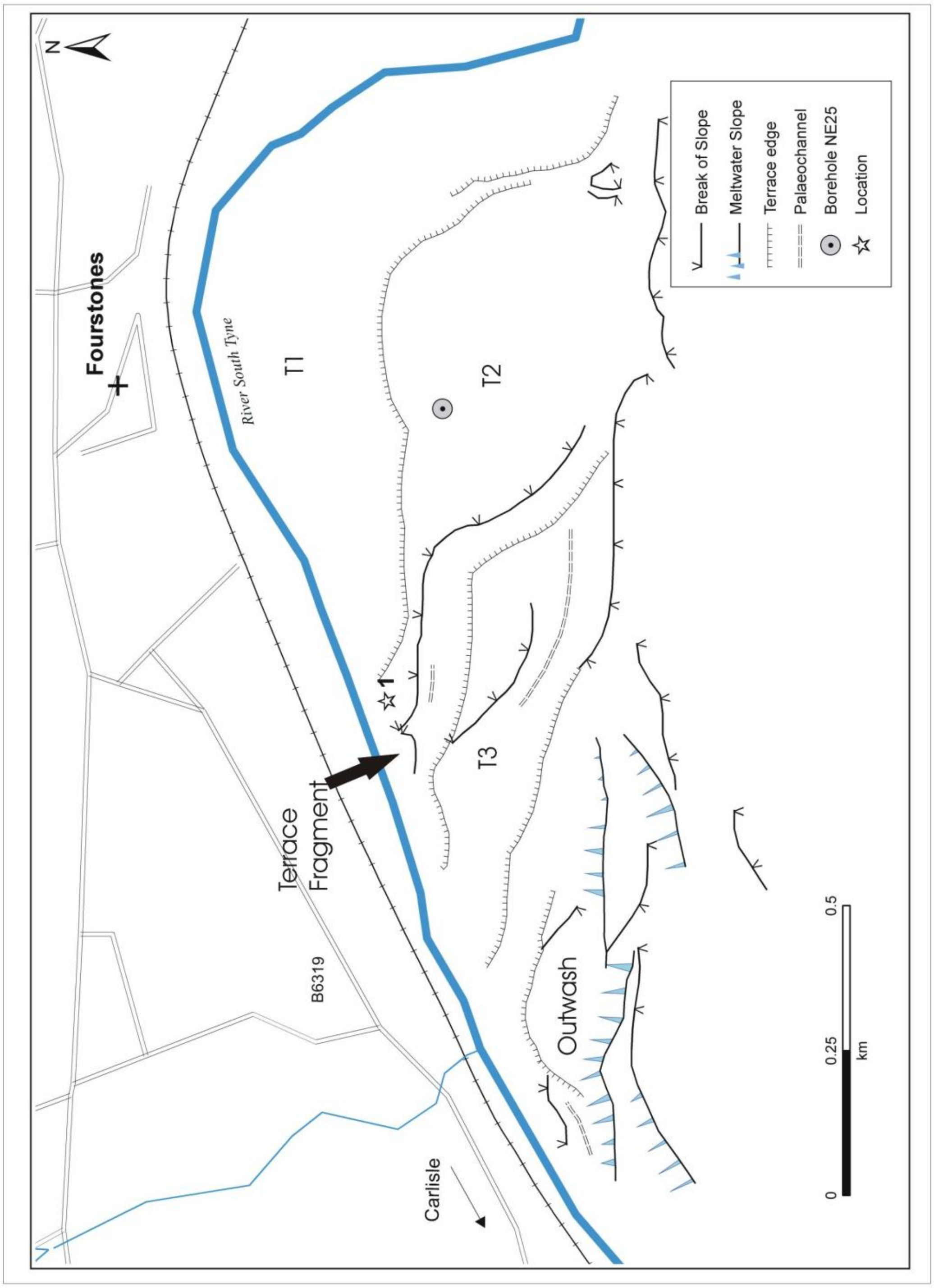


FIGURE 5.35: Geomorphological map of the Fourstones reach. The location of the cut-bank exposure is denoted by location 1. See Figure 5.35 for clarification of terrace units.

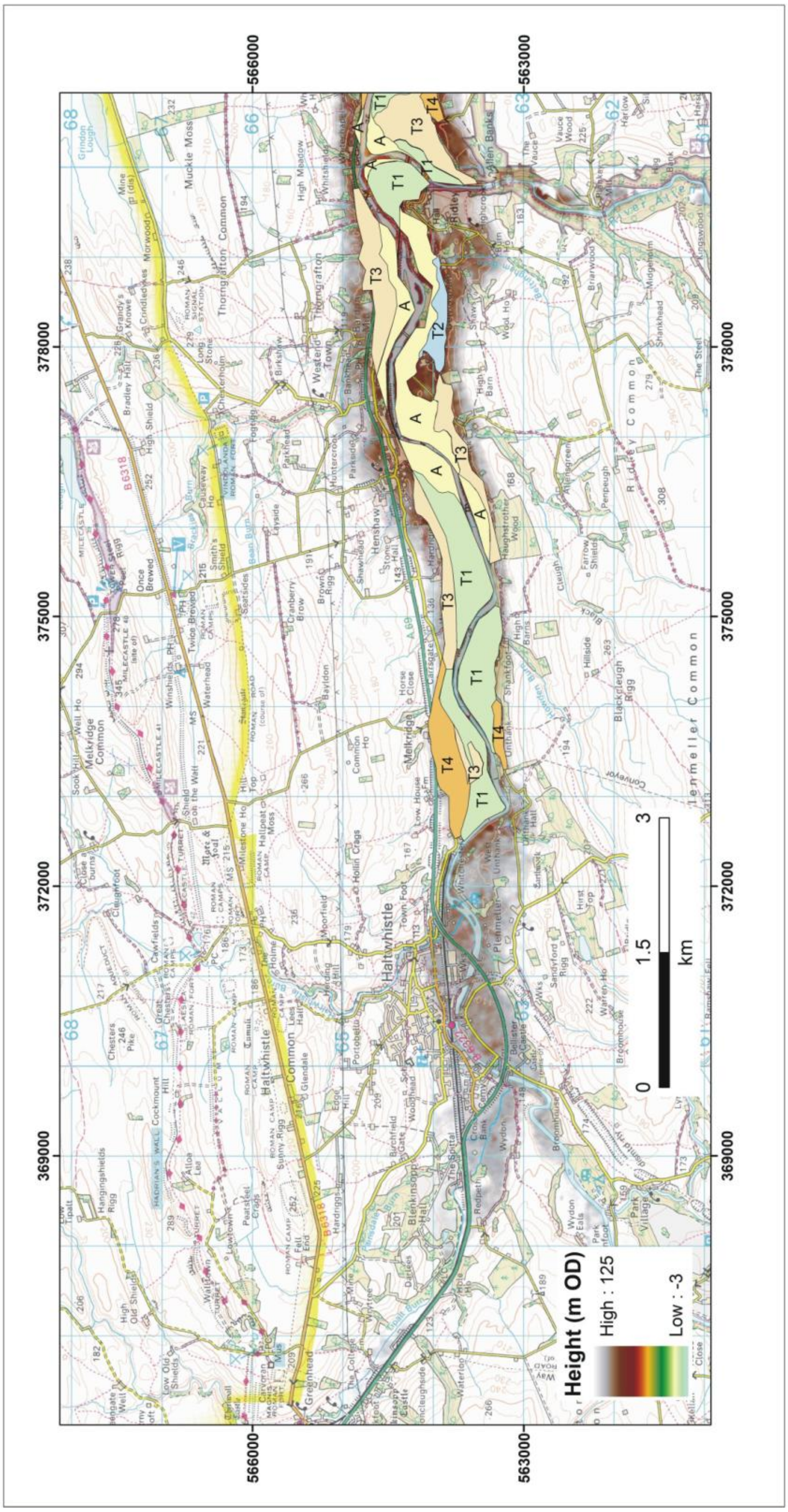


FIGURE 5.36: Geomorphological map of valley between Melkridge and Haydon Bridge showing late Pleistocene and Holocene river terraces and the active channel floor. OS 1:10000 Tile downloaded from digimap (www.edina.ac.uk/digimap), and relief shading generated from 25m DEM grid (www.landmap.mimas.ac.uk).

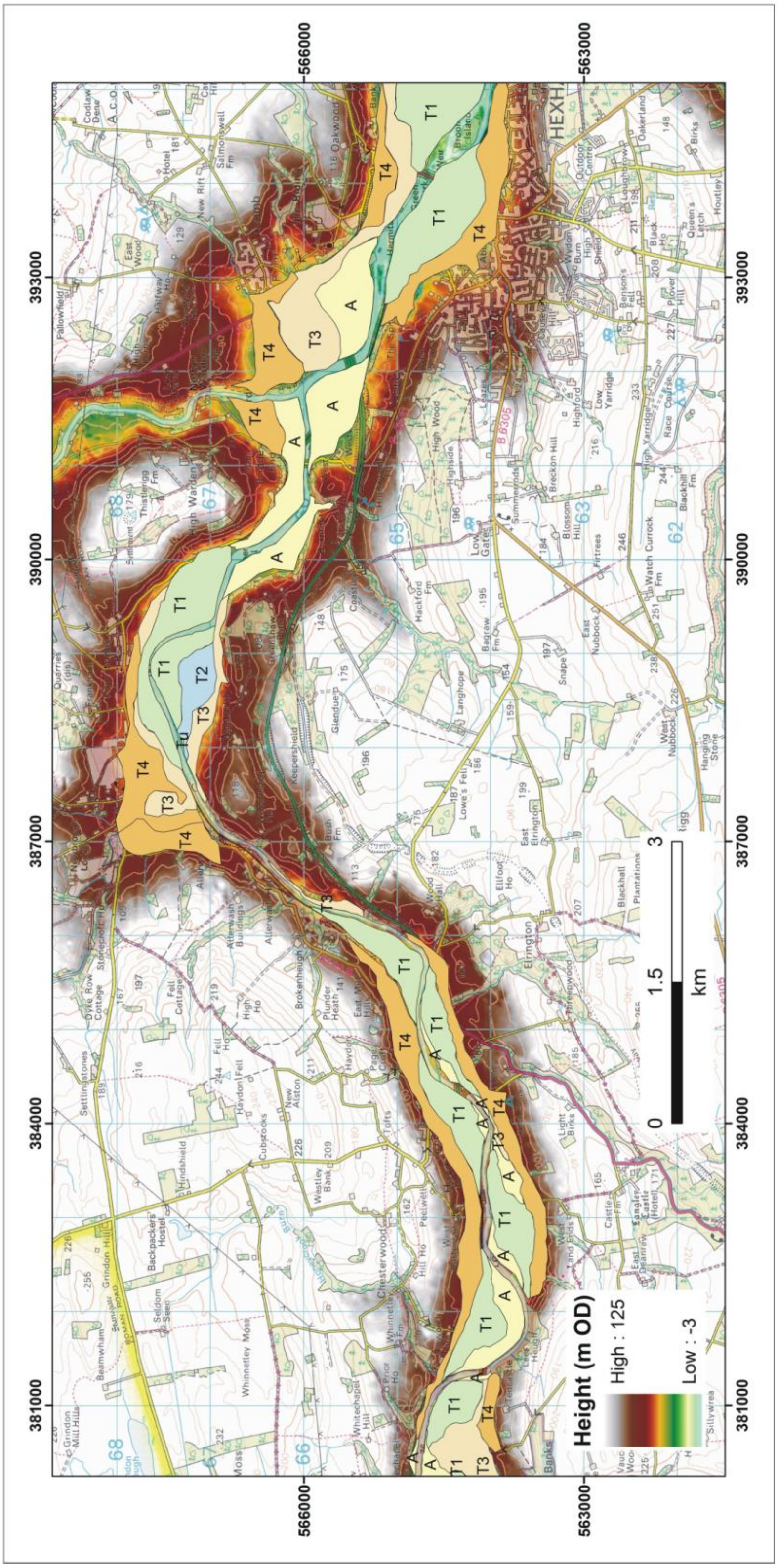


FIGURE 5.37: Geomorphological map of valley between Haydon Bridge and Hexham showing late Pleistocene and Holocene river terraces and the active channel floor. OS 1:10000 Tile downloaded from digimap (www.edina.ac.uk/digimap).

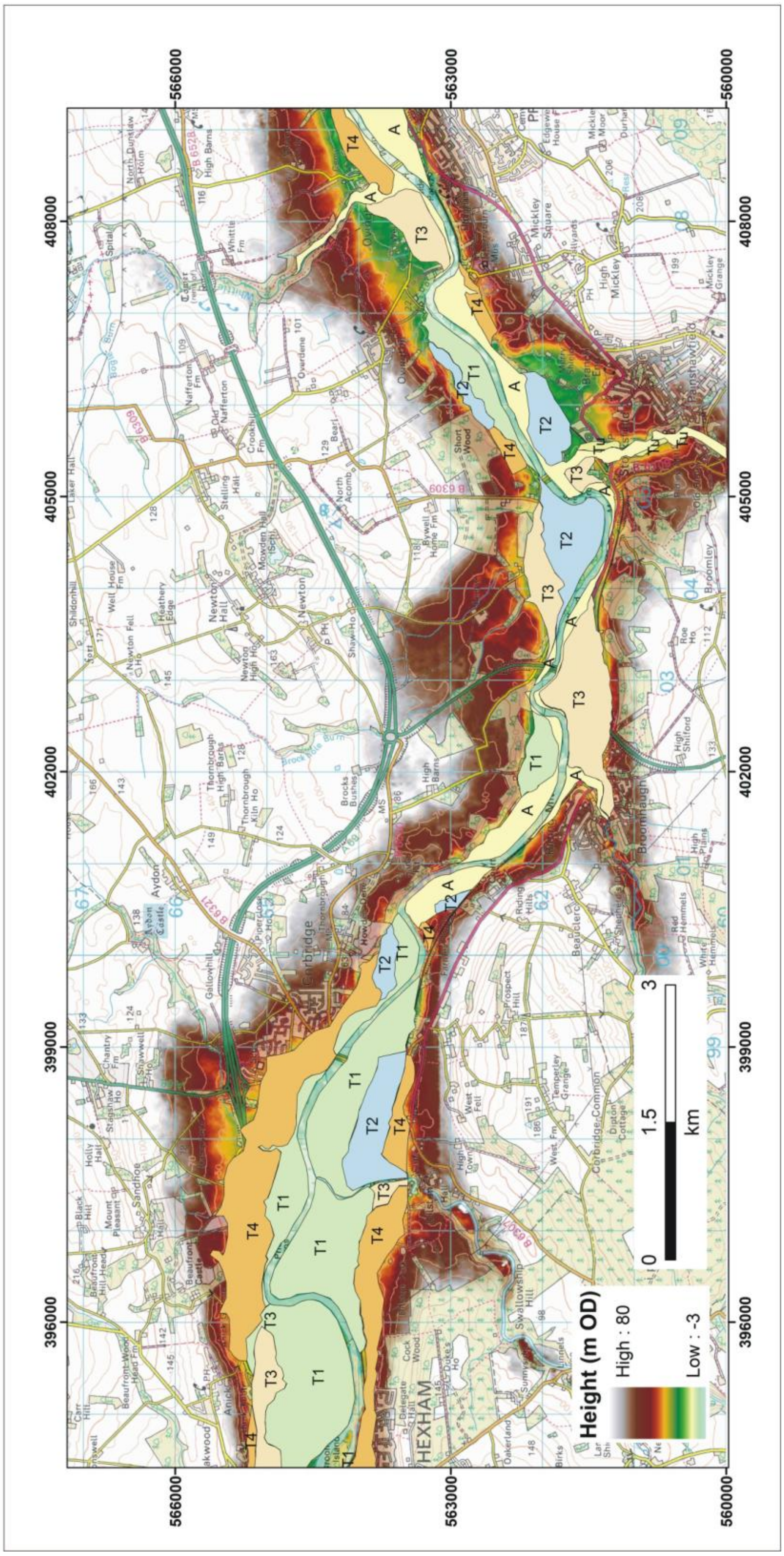


FIGURE 5.38: Geomorphological map of valley between Hexham and Corbridge showing late Pleistocene and Holocene river terraces and the active channel floor. OS 1:10000 Tile downloaded from digimap (www.edina.ac.uk/digimap).

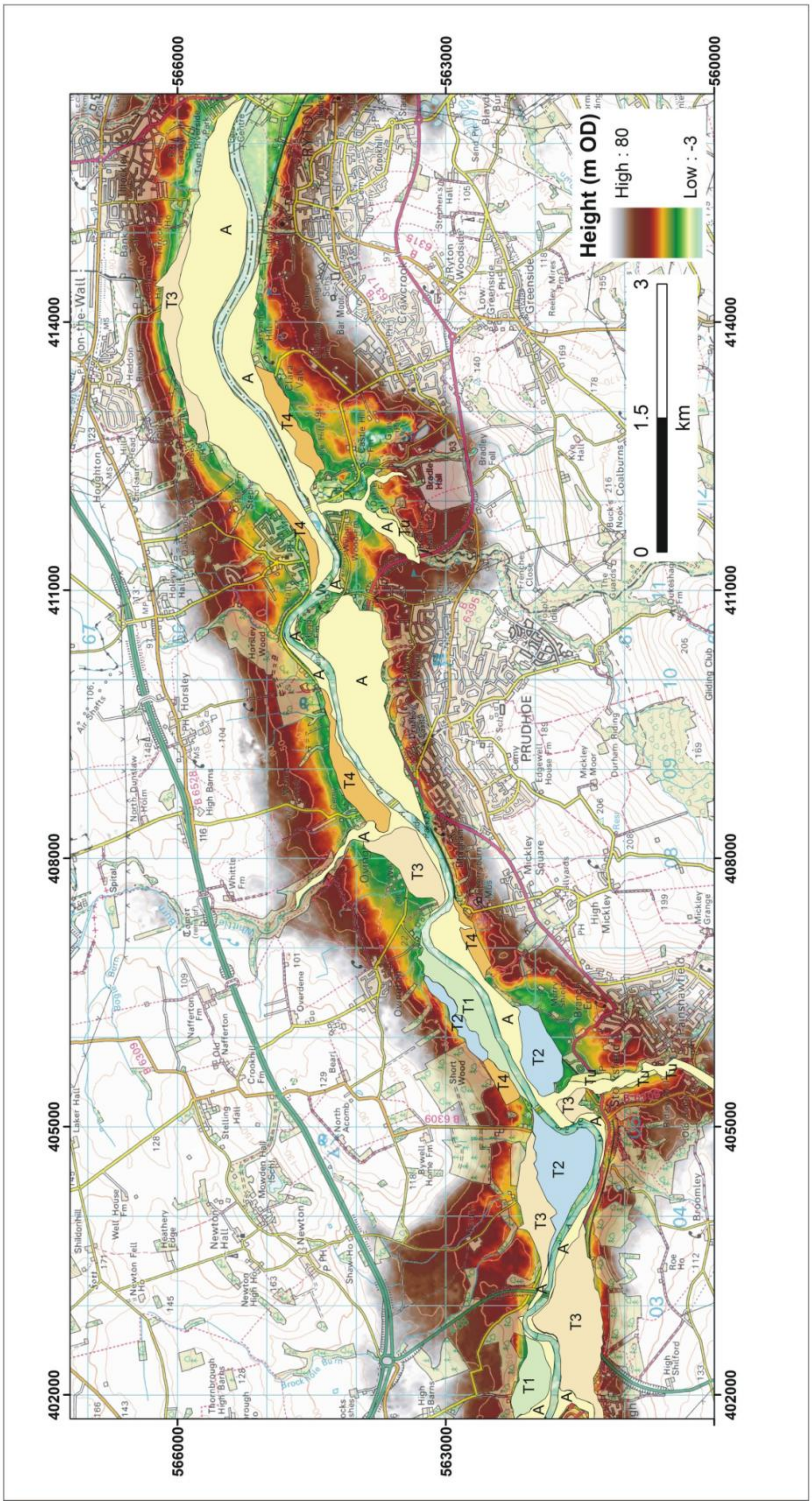


FIGURE 5.39: Geomorphological map of valley between Corbridge and Blaydon showing late Pleistocene and Holocene river terraces and the active channel floor. OS 1:10000 Tile downloaded from digimap (www.edina.ac.uk/digimap).

T4 lies between 40-35 m OD, approximately, 15-20 m above the present river bed and is paired (Figure 5.33). Local scouring at this location accounts for the differential incision. In the lower Tyne at Crawcrook, the T4 surface lies at 19 m OD, ~15 m above the present river level.

Terrace 3

Inset ~10 m below T4 is Terrace 3 (T3), which is less extensive than T4, and its surface lies, on average, 10 m above present river level. T3 has well defined bluffs and surface palaeochannels, although these are not always well preserved or clearly visible. In the South Tyne at Fourstones, T3 lies at 55 m OD, ~10 m above present river level and is paired (Figure 5.35). At Stocksfield, T3 lies at 35 m OD, approximately, 10m above the present river bed (Figure 5.32). The terrace is not laterally extensive at this reach, and its pronounced bluff may be related to some quarrying (Giles, 1981).

Terrace 2

The least extensive terrace identified was Terrace 2 (T2). This is preserved at a few localities in the vicinity of Fourstones, Farnley Haugh Scar and Stocksfield, and was identified through intensive ground survey (Figures 5.32, 5.33, 5.35). Its surface lies, on average, 8m above present river level and, ~2 m below the surface of T3. T2 is tentatively correlated with the Passmore and Macklin's (1994; 1997) Willy Wood terrace identified by at Farnley, which lies 9-8m above present river level. The terrace has been dated to between 4940 and 4600 BC (^{14}C), with aggradation continuing between 1350 and 550 BC (palaeomagnetic dating; Macklin *et al*., 1992). In the South Tyne at Fourstones, T2 lies at 52 m OD, approximately 8 m above present river level. In the mid Tyne at Farnley, T2 lies at 28 m OD, approximately, 8 m above the present river bed. In the lower Tyne at Stocksfield T2, although rather degraded by channels running off the valley side, lies at 22 m OD, is ~8 m above present river level and is paired.

Terrace 1

The lowest and, therefore, most recent terrace in the sequence was identified as a Holocene, cut-and-fill alluvial unit (T1). Its surface lies, on average, 4-5 m above present river level, is omnipresent and, is ~4 m below T2. In the South Tyne at Fourstones, T1 lies at 49 m OD, approximately, 5 m above present river level (Figure 5.35). This terrace is laterally the most extensive at this location, and is paired. In the mid Tyne at Farnley, T1 lies at 24 m OD, ~4 m above present river level. In the lower Tyne at Stocksfield, the Holocene cut-and-fill terrace lies at 18 m OD, approximately, 4 m above the present river level, and palaeochannels are clearly visible on the surface of the alluvial unit (Figure 5.32). This unit has been correlated with Passmore and Macklin's (2000) T4, identified in the upper South Tyne Valley at Lambley, and where wood fragments from the gravel member were ^{14}C dated to 1110-1360AD.

Local Terraces

A fragmentary remnant of a former terrace (Tu) was identified in the South Tyne Valley at Fourstones inset into T2, and was differentiated on the basis of sedimentology (see section 5.2.4). Its surface lies at 52 m OD, ~ 8 m above present river level (Figure 5.35).

5.7 Terrace Sequence: reconstruction of terrace profiles

From the altitudinal points collected from the NEXTMap DEM, long profiles of the River Tyne and its terraces were created (Figure 5.40). A river is only able to incise uniformly if it is able to transmit all of the base-level fall from its mouth upstream

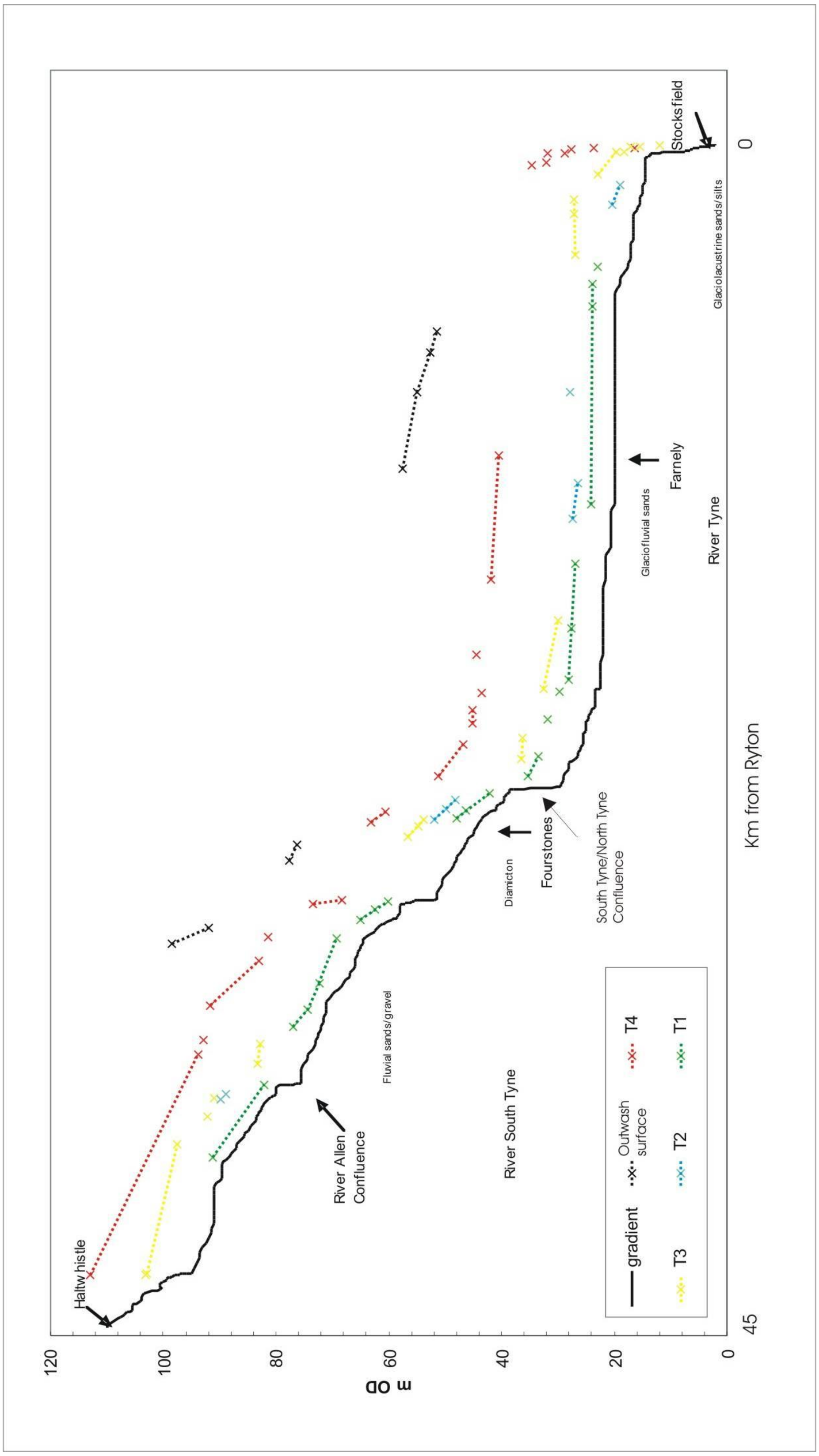


FIGURE 5.40: Reconstructed long profiles of the river terraces and the contemporary River Tyne. Key sites are marked, and the subsurface sediments indicated.

along its length; in most cases this does not happen giving rise to knick points. A major knick point exists in the vicinity of the North/South Tyne confluence, and is observed on all of the river terraces and present day gradient. Above the knick point, the gradient is steep and below it the profile flattens out, essentially showing an exponential decay curve. Knuepfer (2004) identified this type of profile as typical of river valleys affected by glaciation, where the headwaters have been over steepened by ice, and the lower, flatter profile, is indicative of conditioning by glaciofluvial outwash. The knick point divides the upper and lower Tyne, and the IMAU survey report (Lovell, 1981) suggests there is sedimentary evidence for a moraine at this location but there was no topographic/morphological evidence found in this study. A second knick point lies in the main Tyne Valley, upstream of Stocksfield, and is present in all four terrace profiles and the contemporary channel profile. The knick point has not migrated upstream from its location in T4.

Interpretation

The location of the knick point at the confluence of the North and South Tyne is interpreted to relate to increased erosion downstream of the confluence rather than demarcating the location of an ice margin (J. Shaw, personal communication, 2007). Knick points along the South Tyne profile also correspond with the confluence of tributary streams. The knick point above Stocksfield relates to a lithological discontinuity in the bedrock that has been exploited by the river rather than relating to a change in base profile induced by eustatic or isostatic change. Knick point location has remained static during the formation of all four terraces.

The difference in elevation between the terraces, and above and below the confluence knick point, remains unchanged along the length of the profile. The greatest height difference between terrace surface elevation (and height above present river level) occurs between T4 and T3. The amount of incision during the early Holocene was (probably) driven by high rates of glacio-isostatic uplift which subsequently declined throughout the Holocene, reflected in the reduced height difference between younger terrace surfaces. There is no change in the base profile, and the terraces are not out of grade with each other or the contemporary profile. Neither is there any downstream convergence of the terrace profiles, which would suggest sea level rise and burial of older units by the onlapping alluvial wedge. Furthermore, there is no evidence of tilting in the profiles that could indicate localised response to uplift within the basin. The presence of terraces, however, indicates there has been continued incision throughout the Lateglacial and Holocene periods, suggesting long-term regional uplift, with the only candidate being postglacial isostatic rebound. The ice reached a thickness of ~1 km over the region, and rebound has declined to a present day uplift rate in the order of ~0.5 mm/yr for Northumberland (Shennan, 1989). Lambeck's (1995) theoretical isobases for the British Isles predict a rise of ~40 m between 12 and 7ka BP for the region (see Figure 2.10). Thus, isostatic rebound combined with sea-level rise (up to the mid Holocene, cf. Plater and Shennan, 1992), coastal shortening and climatic change are all potential drivers of incision. The fact there is no change in the base profile over the last ~10k years suggests the river has been able to keep pace with these changes, and maintain its profile.

In the Tyne estuary below Blaydon, the valley is steeply incised, and this incised profile continues to the coast at Tynemouth (cf. Cumming, 1977). The drainage of Glacial Lake Wear in the Tyne/Wear lowlands (see section 1.2.2), towards the end of deglaciation (cf. Smith, 1994), possibly contributed to incision of the Tyne and Wear, as the proglacial rivers responded to fall in local lake level. However, steeply incised valleys (denes) are characteristic of the rivers that drain southeast Northumberland and the Durham coastline (cf. Horton *et al*., 1999). Widespread incision along the coastal fringe must be related to isostatic rebound following deglaciation, and combined with current data from the river terrace profiles provides a strong argument for this mechanism. However, further work is required to combine data from the coastal ‘denes’, the other east draining rivers and the Tyne Valley before contributing further to that debate.

5.8 Quantification of valley infill, sediment storage and reworking since deglaciation

This section provides a critical review of the prediction and spatial analysis method employed to generate the bedrock DEM and sediment thickness models, which were used to estimate amounts of sediments infilling the valley. In order to examine the behaviour of the river since deglaciation, the amount of sediment delivered, stored, reworked and exported from the valley was calculated. By combining borehole data on terrace thicknesses with the areal extent of the terrace units, the volume and rates of sediment storage and transport were quantified. Huisink (1999) employed this method to examine the relationship between climate change and fluvial response in the Maas Valley, Netherland. It is generally accepted that, during deglaciation following the LGM, increased meltwater discharge and sediment availability resulted in peaks of sediment transport (Lewin and Macklin, 2003), and the rapid adjustment of the glaciated landscape to non-glacial conditions i.e. paraglaciation (*sensu* Church and Ryder, 1972). As the climate ameliorated, and the ice began to retreat from the drainage basin, fluvial systems initially responded to a deglacial (nival) regime. The newly exposed landscape would have been highly unstable, with enhanced sediment supply to rivers through weathering and erosional processes, such as gelifluction, and readily accessible coarse sediments from the un-vegetated, steep slopes. With the onset of full interglacial conditions (15-12.9ka cal. BP), sediment availability and discharge start to decrease; however, landscapes (and rivers) continue to experience paraglaciation (cf. Ballantyne, 2003) for a time following deglaciation and systems, are therefore, out of equilibrium during this period. The onset of the Younger Dryas (12.9-11.5ka cal. BP), represented a return to cold, dry periglacial conditions that resulted in incision followed by aggradation as the rivers responded to the climatic shift from warm to cold conditions, and the dominance of seasonal nival discharges (Lewin *et al*., 1995). At the Younger Dryas-Holocene transition, temperatures rose very quickly from mean July temperatures of <9^0 to 15-17^0 (Coope *et al*., 1998), and precipitation also increased by 50% (Dansgaard *et al*., 1989). The early Holocene period is generally thought to be a period of quiescence in the upland fluvial record (cf. Macklin and Lewin 1993; Lewin and Macklin 2003). Johnstone *et al.'s* (2006) latest assessment of ^{14}C dated alluvial units shows activity in lowland and unglaciated catchments, but there is little evidence from upland basins to date. This could be related to a lack of identified and dated terraces from this period or, possibly, indicates a period of reworking/erosion rather than inactivity. The Lateglacial/early Holocene transition between 11.7-7.9ka cal. BP was one of continued climatic instability; three periods of instability are identified in the Greenland ice cores during the early Holocene at 9.5ka, 9.3ka and 8.2ka cal. BP, and correspond to cooling events of several degrees (Rasmussen *et al*., 2007). The 8.2k event is thought to represent a more dramatic cooling event that was associated with a longer-term cooling between 8.6-8ka cal. BP (Rohling and Pälike, 2005) and, like the Younger Dryas Stadial, is associated with meltwater discharge from the Laurentide ice sheet (Barber *et al*., 1999).

Although the sediment storage and reworking calculations made in this study were limited by the availability of data, the results provide order of magnitude estimates of reworking and sediment export. Combined with the geochronological framework, the data was then examined for relationships to regional climatic change and/or local drivers.

5.8.1 Mathematical modelling: evaluation of the approach of polynomial and linear prediction

The purpose of modelling was to predict the shape of the valley bottom prior to the last glaciation as a means of estimating the amount of subsequent sediment infill. Two approaches to valley profile prediction were adopted, these being: (1) polynomial curve; and (2) linear regression. Because of the lack of freely available[9] borehole data, which would detail elevation at bedrock, the valley prediction model was employed as surface elevation datasets were held (e.g. NEXTMap and landmap). Whilst the approaches have in principal been shown to be successful in Alpine glaciated valleys (cf. Hinderer 2001; Schrott *et al*. 2003) they were based upon the fact that the upper valley sides had not been glaciated and reflected a pre-glacial profile. In contrast the Tyne Valley was completely overrun by ice and the valley side profiles have a stepped profile and an asymmetrical form, thus only the lower valley sides were considered in this study.

The evaluation of the linear and polynomial generated DEMs (see section 4.6.1) was carried out as a visual inspection of the output surfaces. In both cases, the interpolation methods were experimented with and qualitatively assessed in order to ascertain the most appropriate DEM surface. A comparison of the interpolation techniques indicated that geostatistical analysis (Ordinary and Universal kriging) provided the most reliable

[9] Subsurface data is held by the BGS and available for purchase

representation of the data in terms of characterising the spatial variation in bedrock elevation. This was because the spatial model is developed from the actual data and based on measured correlation and weighted measurements of nearby points.

In terms of model generation, the interpolation was carried out on the full set of data generated from the polynomials and linear regressions. The linear regression approach (cf. Hinderer, 2001) to valley profile prediction resulted in a poorly smoothed DEM surface. Despite revising the spatial interpolation model a number of times, the output surface was always represented with a series of mounds and hollows giving a 'bulls-eye' pattern to the surface (Figure 5.41). It was concluded that, whilst the spatial interpolation model was as good as it could be (i.e. kriging variances were negligible), ultimately, the low resolution of input data points meant the surface would never be accurately represented. Linear regression could be reliably applied where a V-shaped valley profile existed (indicating little glacial modification: cf. Hirano and Aniya 1988, 1990; Harbor 1990), however, in the case of the Tyne Valley, only a limited number of cross-sections met this criteria.

The polynomial curves approach (cf. Schrott *et al*., 2003) was also refined through several iterations of spatial interpolation modelling. Initially, the surface was modelled from the points generated from the polynomials curves. However, the polynomial curves were overpredicting the depth to bedrock when checked against known bedrock z values derived from borehole logs. Thus, the output surface was not an accurate representation of the bedrock surface. Therefore, in the final model, the polynomials curves were forced to go through bedrock elevation points where they were known from neighbouring depth data. By conditioning the model in this way the output surface was most likely to reflect the true bedrock elevation or at least the closest

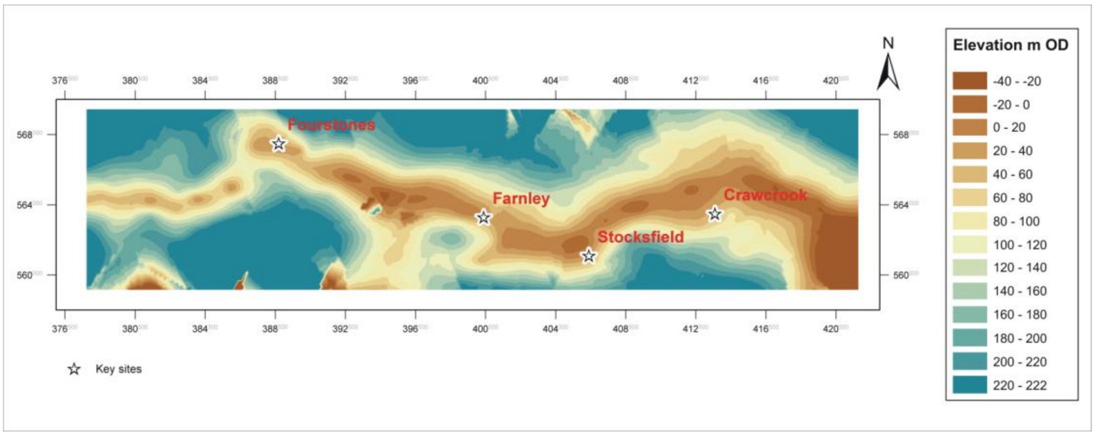


FIGURE 5.41: Bedrock DEM generated from the linear regression approach, using kriging. The valley floor is represented as a series of hollows. Key sites are marked on the map. This model was rejected, and not used in the analyses.

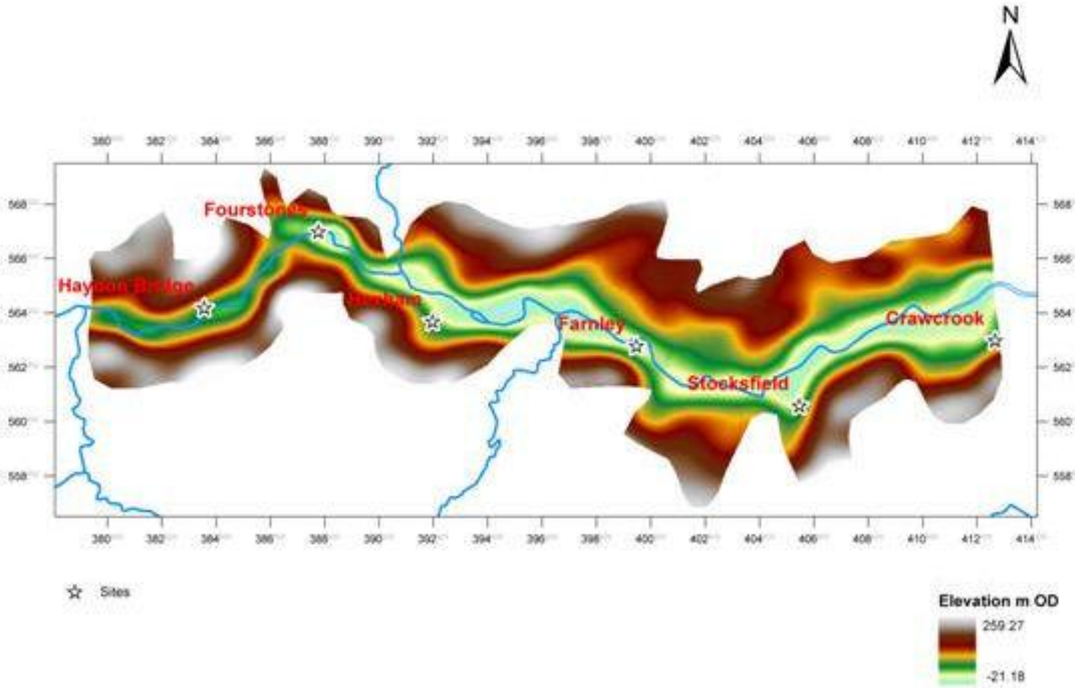


FIGURE 5.42: Bedrock DEM, generated using kriging from points predicted using polynomial curves and conditioned with known z values from borehole data. Sites are marked on the map. This was the model used in analysis.

approximation. The adjustment to neighbouring depth data resulted in a more accurate estimation of sediment thickness, which was not possible when the polynomials were not adjusted. This, of course, results in limited independence in the output DEM model but at least provides a good, overall prediction of bedrock. Kriging variances were used to examine the accuracy of the output surface, rather than using a validation set (between 10-20%) removed from the full data set, which would have reduced the size of the data set for the interpolation. Once a satisfactory interpolation model had been determined, and a surface generated, kriging variances indicated the error between the input value and the prediction value to be negligible. Thus, the interpolation model was accepted as an accurate representation of the input data.

The bedrock DEM (Figure 5.42) predicts bedrock elevations in the lower Tyne to be ~-22m OD, and ~3m OD west of Hexham; the predictions are less reliable for the South Tyne Valley. The model predicts a valley width of ~ 2km, widening slightly towards the confluence with the Team Valley (i.e. The Wash, cf. Wood and Boyd, 1863), which is in keeping with known pre-glacial valley bedrock elevations between -30 and 0m OD, and widths of between 2-3km along the main Tyne Valley (cf. Cumming, 1971). However, whilst the polynomial modelling approach is relatively quick and simple to apply, and required minimal data, it was judged unsatisfactory in the case of the Tyne Valley because the Tyne Valley did not fit the a priori assumption that glaciated valleys are parabolic in shape and can be modelled by a polynomial curve (cf. Aniya and Welch, 1981). Even though the method produced results, it is concluded that the predictions are speculative and that where bedrock elevation data from boreholes is available this should be used to generate a bedrock surface when a valley has been completely overrun by ice.

5.8.2 Spatial distribution of sediment within the river valley

The difference in elevation between the bedrock DEM and the surface DEM enabled the volumetric infill to be calculated and from the resultant DEM the spatial distribution of the sediment storage could be examined. The spatial distribution of sediments in the South Tyne and Tyne Valley are shown in Figure 5.43. The thickest accumulations of sediments are located in: (1) the mid Tyne Valley between Hexham and Dilston, and the lower Tyne Valley in the vicinity of the Crawcrook; and (2) the lower South Tyne Valley, in the vicinity of the Fourstones reach. The spatial pattern of sediment transfer would suggest storage is greatest where the valley widens out. At Fourstones, South Tyne, sediments have accumulated immediately downstream of a valley constriction and represent an alluvial basin. Within the main Tyne Valley, downstream of the North/South Tyne confluence, a laterally extensive (~2km) zone of sediment accumulation occurs between Hexham and Corbridge, and is the contemporary storage zone (alluvial basin) of the modern river. Sediment accumulation zones are also associated with tributary valleys where they enter the main Tyne Valley, depositing their bedload as alluvial fans. The greatest thickness of sediment stored in the valley, up to ~55m, occurs in the two alluvial basins at Fourstones and Hexham. In general, postglacial river valleys in glaciated uplands, are characterised by wide basins infilled with alluvial sediments separated by narrow reaches, giving a distinct ‘hour glass’ configuration to the valley floor (cf. Macklin, 1999).

In the lower Tyne Valley, between Stocksfield and Blaydon, the sediments are almost exclusively stored on the south side of the valley, and are up to 55m thick. The predilection to storage on the south side is related to the accommodation space which existed during deglaciation. The pre-glacial route of the Tyne flowed south of the

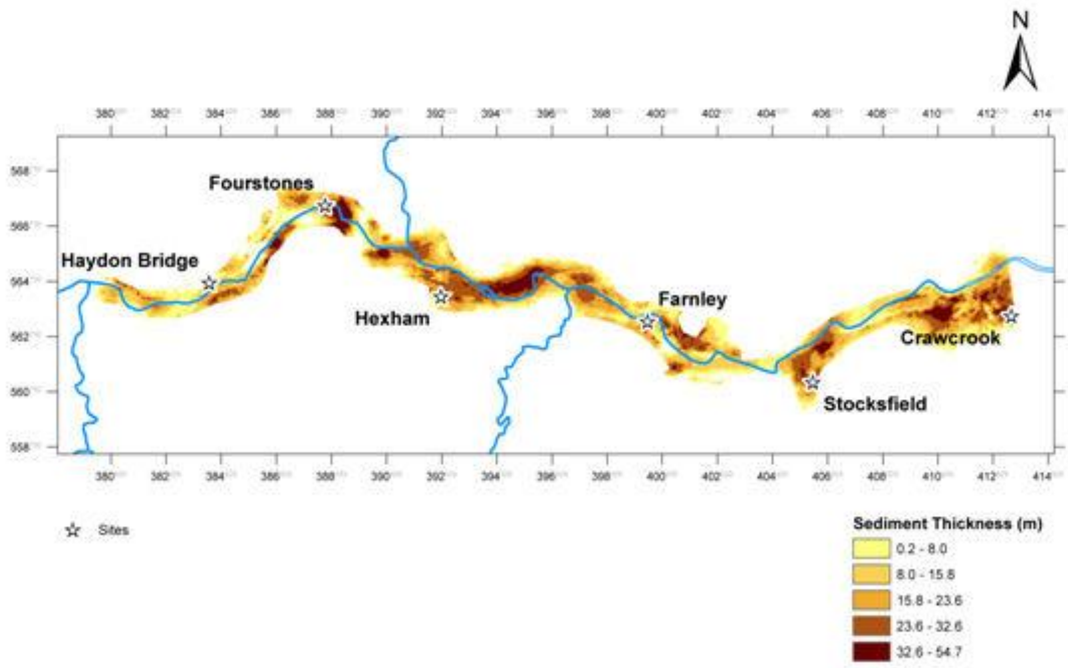


FIGURE 5.43: Calculated sediment thickness derived from the geomorphometric (polynomial curve) analysis. The thickness was estimated by subtracting the surface DEM and the bedrock DEM. Sites are marked on the map.

present day route, and this was infilled, giving an asymmetrical distribution to the sediments (cf. Cumming, 1971).

5.8.2.1 Sediment storage

Since the disappearance of the late Pleistocene ice from the Tyne Valley, the river has been responding to conditions set as the legacy of glaciation (cf. Macklin and Lewin, 1997). Deposition of the valley infill has occurred under paraglacial conditions, the influence of which has continued into the present interglacial. Within the Tyne Valley, the total volume of sediment stored as Pleistocene valley infill (including till) between the Allen confluence and Crawcrook, has been estimated by subtracting the surface DEM from the bedrock DEM to be 0.85km^3. The size of the drainage basin during deglaciation was probably larger than the present day drainage area, as sediment and meltwaters entered the South Tyne catchment *via* the Tyne Gap, which was probably draining north eastern Cumbria and the Lake District ice-margin, but this is unquantifiable. However, by redistributing the valley infill to the area of the present day drainage (~2.9km^2), the (average volumetric) sediment loss (i.e. denudation) was estimated to be ~29tm^{-2}a^{-1}[10] for the post-LGM period, ~20ka. In catchments currently undergoing deglaciation, denudation rates of 2700∀710tkm^{-2}a^{-1} and 1800∀350tkm^{-2}a^{-1} (Hodgkins *et al.*, 2003), and +460tkm^{-2}a^{-1} (Etzelmüller, 2000) have been estimated, illustrating the figure for the Tyne catchment is a likely underestimate. The average figure cannot account for higher rates which would have occurred during the paraglacial period.

Volumetric estimates of fluvial sediment reworked and net erosion, and volumetric loss (denudation) for the Tyne Basin were calculated for each phase of development by employing OSL ages derived from the research and other available dates from the Tyne Valley (e.g. Passmore and Macklin, 2001). Calculations and justification for data used in the calculations are given in section 4.6.2 and the results are presented in Table 5.3. Based on the calculations, the total volume of sediment stored in the study area was estimated to be 0.84km^3, of which 0.42km^3 is stored as glacigenic sand and gravel and 0.39km^3 is stored as alluvium. Postglacial development is reflected in the river terraces that flank the valley sides. Thus >50% of the sediment currently stored is directly related to deglacial deposition.

TABLE 5.3: Volumetric estimates of Holocene fluvial storage and total valley fill storage in the study area.

TOTAL FLUVIAL STORAGE KM^3	TOTAL VOLUME SEDIMENT STORAGE KM^3	TOTAL AREA KM^2
0.39	0.84	70.21

[10] Bulk density for the calculation was assumed to be 2.02, sand with gravel (wet).

By summing export from each phase of development, based on the maximum floodplain area (see Table 5.4), a volume of 0.48km^3 has been removed (loss) from the system since deglaciation. A comparison between present day (total) sediment storage and export shows that >50% of the glacigenic sediment deposited has been removed from the system. Storage is approximately equal to export since deglaciation.

5.8.2.2 Volumetric estimates of sediment reworking and export

Table 5.4 summarises the volumetric estimates of sediment erosion (export) and deposition (reworking) for each phase of terrace development, and this is shown graphically in Figure 5.44. The maximum floodplain for each terrace was determined by summing the surface areas of the terrace, all the younger terraces and the contemporary channel. The minimum floodplain was simply the area of the terrace. These then gave upper and lower estimates of sediment reworking and export rates for each phase of terrace development.

From the chronology, there is clearly a hiatus in valley floor development between the aggradation of the outwash sediments and the first recorded episode of fluvial development, 10ka cal. BP. Sediment reworking (Table 5.4; Figure 5.44) was highest during the early Holocene (11.4-7.5ka cal. BP) with 0.96km^3 of sediment reworked during the T4 to T3 phase. This is twice as large as reworking throughout the majority of the Holocene (T2 to T1 phase), with 0.51km^3 of sediment reworked, and is three times as large as reworking during the late Holocene (historic) period (T1 to Alluvium phase), with only 0.28km^3 of sediment reworked. Sediment reworking estimated for the development of T4 and T3, gives a 25% increase in reworking from 0.59 to 0.82km^3 during the trenching of T4 and development of T3. This was followed by a 50% reduction in reworking during the succeeding phase of development, with a rate of

TABLE 5.4: Volumetric estimates of sediment reworked and exported from storage for each of the terrace units in the Tyne Valley during the last ~12k years.

Floodplain	Sediment reworked during each phase (m^3)	Sediment exported (m^3)	OSL Age Range (ka cal. BP)
OT-T4			
Max Floodplain	0.59	0.38	~11.4-9.2
Min Floodplain	0.12	0.08	
T4-T3			
Max Floodplain	0.81	0.30	
Min Floodplain	0.14	0.05	
T3-T2			~9.1-7.9
Max Floodplain	0.42	0.05	
Min Floodplain	0.13	0.02	
T2-T1			
Max Floodplain	0.35	0.09	~7.6-5.4
Min Floodplain	0.15	0.01	
T1-Alluvium			
Max Floodplain	0.20	0.03	
Min Floodplain	0.15	0.02	

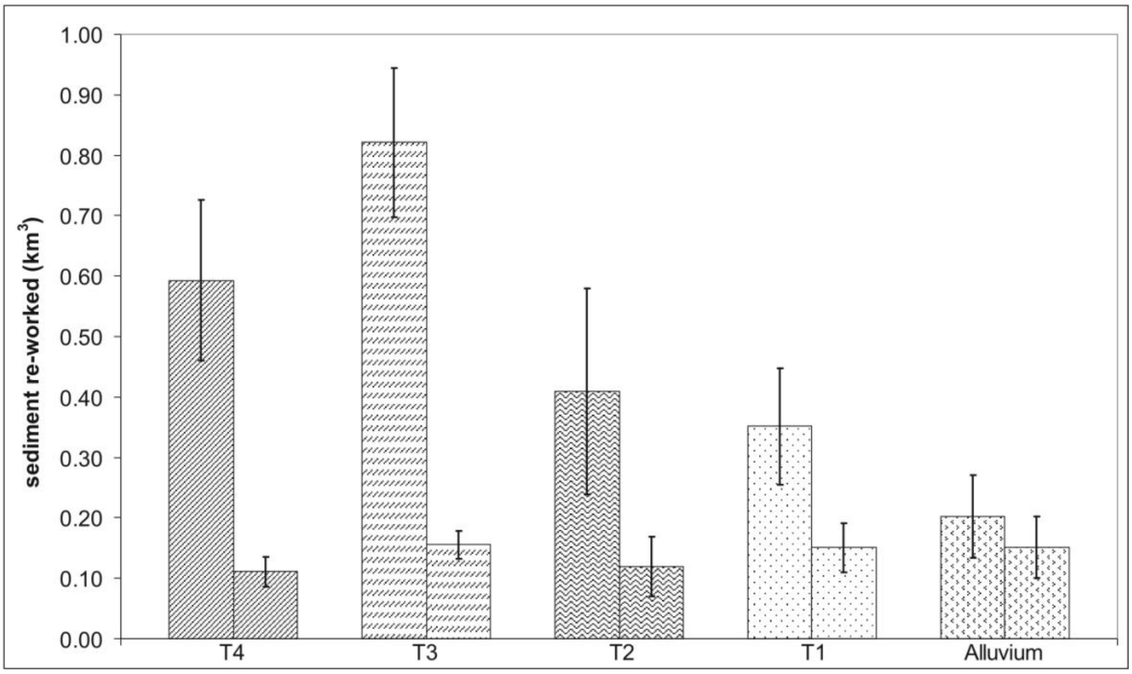


FIGURE 5.44: Volumes of sediment reworked for each phase of sediment unit development based on estimates of minimum and maximum floodplain size and sediment thicknesses and error bars depicting standard error (see section 4.4.2 for explanation).

0.41km^3. Volumes of sediment export decline from 0.38 to 0.03km^3 throughout the Holocene, suggesting that sediment is being stored for longer and less is available to the river as it becomes progressively entrenched and confined within its own terraces.

Incision was calculated by summing the depth to base of the terrace unit and the height difference between the terraces. Valley floor development, in terms of incision (erosion) *vs*. aggradation (deposition) since deglaciation, is characterised by net incision following deglaciation through to the development of T3, *c*. 8ka cal. BP, and during the development of T1, ~7-1ka cal. BP. Net incision has been approximately equal to net aggradation during the formation of T2 and the alluvial valley floor (Table 5.5). Based upon summing the thickness of alluvium of the lower terrace unit and the height difference between the two terrace surfaces, depth of incision below the surface of OT before valley floor refilling began to form T4 was ~22m. Whereas the amount of incision achieved in the following phase of activity increased by a third, and ~32m of downcutting took place below the surface of T4.

TABLE 5.5: Net amount of incision or aggradation during the Lateglacial, where ratio is calculated as the amount of incision to aggradation.

PHASE	RATIO (INCISION:AGGRADATION)	MODE
OT-T4	1.8:1	Net Incision
T4-T3	1.4:1	Net Incision
T3-T2	1:1	Incision = Aggradation
T2-T1	1.2:1	Net Incision
T1-Alluvium	1:1	Incision = Aggradation

5.8.2.3 Volumetric sediment loss (denudation)

Estimates of sediment loss (sediment export from the catchment per unit of catchment volume), based on reference bulk density values to represent probable sedimentological characteristics, are summarised in Table 5.6 and in Figure 5.45. Rates of sediment loss increased during the Lateglacial/early Holocene period, followed by an exponential Decrease throughout the Holocene period, until close to the present day (Figure 5.45). Mean sediment denudation rate for the period between incision of outwash sediments and development of T4 (~8k years) was estimated to be 28.8tkm^{-2}a^{-1}. Mean denudation increased by two thirds, up to 93.6tkm^{-2}a^{-1}, during the development of T3 (<1k years) and is interpreted to be the peak of reworking following the Younger Dryas Stadial. During the development of T2 (~1k years) denudation decreased to a sixth of the previous level, to 14.8tkm^{-2}a^{-1}. The development of T3 represents the most dynamic period of sediment erosion, reworking and export in the post glacial period. The highest rate of loss took place between 10-8ka cal. BP. The river was actively down-

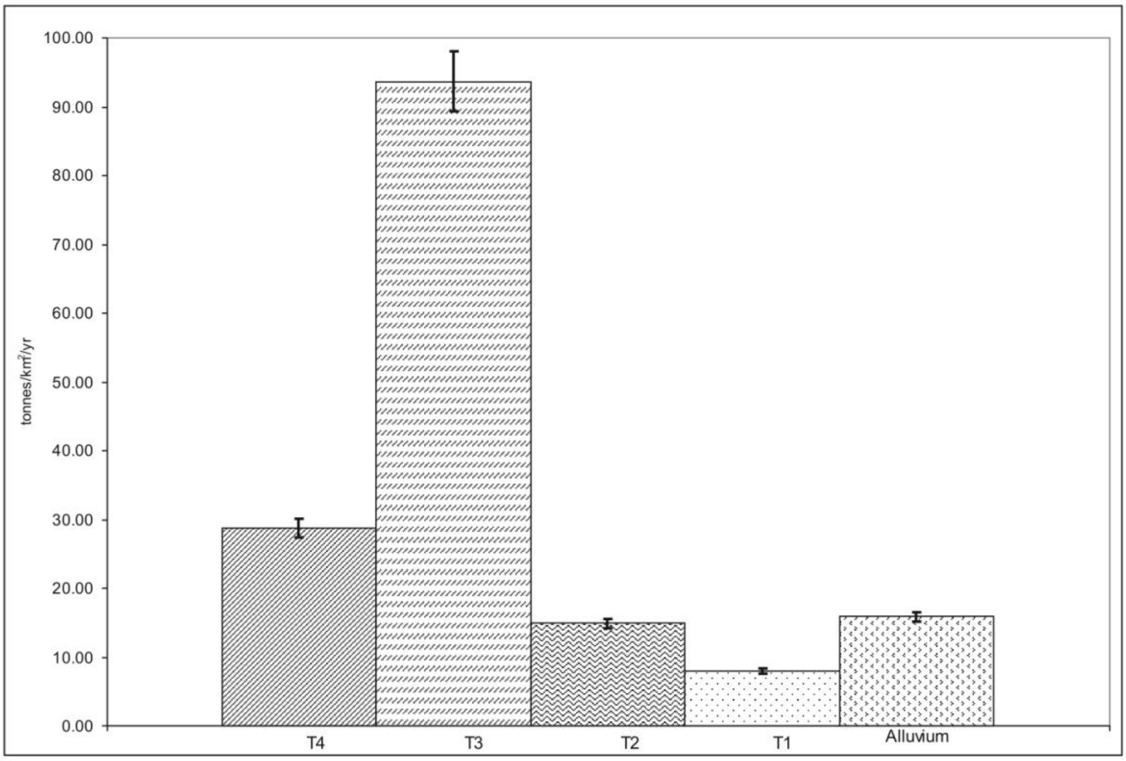


FIGURE 5.45: Volumetric estimates of sediment denudation (loss) for each phase of valley floor development. Standard error is denoted by the bars are included to provide an indication of the cumulative error associated with the calculation. Sediment denudation following at the start of the Holocene peaks during the early Holocene (T3), and then declines, rising slightly towards the present day (alluvium).

cutting through T4 and re-filling the valley floor with glacially derived sediments. A second phase of incision followed, resulting in the formation of T3, before a further period of valley floor aggradation. The peak of reworking is associated with incision and erosion of the largest floodplain, with up to ~22m incision and ~11m of aggradation. Subsequent floodplains are inset below the oldest terrace (floodplain) surface, and sediment loss from the following terraces is constrained by smaller floodplains, and by decreased amounts of incision between terraces.

Interpretation

Based on the OSL ages for the terraces (see section 5.3) and their sedimentology, it is suggested that the alluvial paraglacial cycle is not recognisable in the Tyne Valley record due to the lack of alluvial records for this period following deglaciation. This

TABLE 5.6: Volumetric estimates of sediment yield based on different sediment compositions of units for each phase of terrace development (errors associated with the calculations are shown). FP is floodplain, see section 5.8.2.2 for definition of min and max floodplain area.

	TONNES/KM^2/YEAR													
	Sand (loose)	**Sand (dry)**	**Sand (wet, loose)**	**Sand (wet, packed)**	**Sand w/gravel (dry)**	**Sand w/gravel (wet)**	**Gravel (dry)**	**Gravel (wet)**	**MEAN**	**STDEV**	**SE**	**Upper**	**Lower**	**95%**
OT-T4														
Max FP	23.05	25.61	30.74	33.30	26.42	32.34	26.93	32.05	**28.80**	3.77	1.33	31.42	26.19	2.61
Min FP	4.32	4.80	5.76	6.24	4.95	6.06	5.04	6.00	**5.39**	0.71	0.25	5.88	4.91	0.49
T4-T3														
Max FP	74.94	83.27	99.93	108.25	85.87	105.13	87.54	104.19	**93.64**	12.25	4.33	102.13	85.15	8.49
Min FP	14.20	15.78	18.93	20.51	16.27	19.92	16.58	19.74	**17.74**	2.32	0.82	19.35	16.13	1.61
T3-T2														
Max FP	11.89	13.21	15.86	17.18	13.63	16.68	13.89	16.53	**14.86**	1.94	0.69	16.21	13.51	1.35
Min FP	3.45	3.84	4.61	4.99	3.96	4.84	4.03	4.80	**4.32**	0.56	0.20	4.71	3.92	0.39
T2-T1														
Max FP	6.34	7.04	8.45	9.16	7.26	8.89	7.40	8.81	**7.92**	1.04	0.37	8.64	7.20	0.72
Min FP	0.46	0.51	0.61	0.66	0.52	0.64	0.53	0.64	**0.57**	0.07	0.03	0.62	0.52	0.05
T1-Alluvium														
Max FP	12.68	14.09	16.90	18.31	14.53	17.78	14.81	17.63	**15.84**	2.07	0.73	17.28	14.40	1.44
Min FP	9.51	10.56	12.67	13.73	10.89	13.33	11.10	13.22	**11.88**	1.55	0.55	12.95	10.80	1.08

suggests early episodes of reworking, probably characterised by vertical incision and lateral migration, which may have removed the Lateglacial valley floor. This in keeping with findings from other upland glaciated river basins (cf. Lewin and Macklin, 2003). Therefore, sometime after the retreat of the ice, ~20-16ka cal. BP, the valley floor was trenched, and the glacigenic sediments were incised, eroded and exported.

From the terraces preserved in the Tyne Valley, the analysis suggests sediment yields were highest at the start of the Holocene. OSL ages from T4 and T2 suggest extremely rapid accumulation of the valley fill, and T3 must have developed over a short period. This extremely dynamic period of fluvial activity, only lasting ~4k years, which resulted in extensive reworking, incision and erosion, may represent secondary paraglaciation (*sensu* Ballantyne, 2002a, b). There are five possible causes for this fluvial response: (1) climate change, and linked to that (2) increasing vegetation cover, (3) crossing of intrinsic thresholds and therefore, complex response, and (4) a lagged response to deglaciation, and linked to that (5) continued isostatic uplift. Although the Tyne Basin was no longer glaciated, regionally the area was under the wider influence of the Fennoscandinavian and Laurentide ice sheets, which continued to exert their influence on the North Atlantic climate. The period of transition from deglaciation to the early Holocene is marked by deep incision on a basin wide scale, but with no evidence of fluvial aggradation. The river then responded with two cycles of incision and terrace aggradation between 11.4 and 7.3k years ago. Within the OSL age range, terrace formation may be in-phase with major climate change, but further work is required to refine and improve the OSL chronology. However, even with dating uncertainties, the sediments and terrace formation suggests the earliest phase of postglacial development was characterised by aggradation. As the climate ameliorated, precipitation increased and as vegetation cover was still very open, glacigenic sediments were readily available and mobilised into the system, with aggradation following. However, as the climate 'flipped' to brief colder, drier conditions during the early Holocene, and probably combined with continued glacio-isostatic uplift, the river incised. During the early Holocene, the system would have been in disequilibrium, and this combined with the other factors, enabled the cycles of incision and aggradation to occur. Sedimentologically, T4 and T2 are characterised by wandering gravel bed planform but include major flood deposits indicating instability within the system.

The two major terraces (T3 and T4) formed over a relatively short period during the Younger Dryas and the early Holocene, whilst for the majority of the Holocene the river has been moving across the valley floor, intermittently incising in response to periods of dynamic climatic and anthropogenic change (cf. Macklin *et al*. 1992a; Macklin 1999) but, in general, achieving little in comparison with the early Holocene period. The development of T2 is related to lower rates of sediment reworking, an order of magnitude decrease in sediment export, and approximately equal rates of incision and aggradation. T2 is characterised by lateral migration and flood events, indicating periods of instability and extreme events. The formation of the post 8k year alluvial valley floor is characterised by net incision, and follows the formation of T2 and the aggradation of T1, and represents the majority of development throughout the Holocene. The formation of the cut and fill units is widely attributed in the literature to both climatic forcing (cf. Macklin and Lewin 1993; Johnstone *et al*. 2006) and anthropogenic disturbances within the catchment (cf. Macklin, 1999). Historic rates of reworking coincide with both increase climatic instability i.e. the Little Ice Age (cf. Rumsby and Macklin, 1994, 1996) and increased anthropogenic activity within the drainage basin (cf. Rumsby, 1991). However, the subtleties of each incision and aggradation episode during the mid Holocene to historic period are outside the current analysis because cut and fill alluvial units were not specifically identified; they have been discussed in detail elsewhere (cf. Passmore and Macklin, 1994, 1997, 2001).

5.9 Summary

The main findings have been summarised below, the implications of which are discussed in chapter six.

1. Sedimentological evidence indicates glaciolacustrine and glaciofluvial deposition in a proglacial and lacustrine environment. Small, temporary ice marginal lakes formed in the pockets of ice-free areas that developed along the Tyne Valley between Hexham and Crawcrook as the ice downwasted.
2. The mounded landform assemblages in the lower Tyne Valley developed as erosional remnants following dissection by meltwaters from the ice front, and as ice marginal lakes drained. Meltwater flow was generally in an west-east direction. The Crawcrook complex is, therefore, considered to be the product of post-depositional erosion and not supraglacial letdown during ice stagnation.
3. The eastern limit of the Crawcrook complex may mark the inland ice position of the coastal ice. The presence of subglacial till overlying sandur deposits indicates a possible ice-marginal position.
4. The extent of aggradation in the lower Tyne Valley suggests the area was fed by active or stagnating ice located further up valley (i.e. mid-Tyne and lower South Tyne).
5. Limited sedimentological (borehole) data from landform-sediment assemblages that outcrop in the lower South Tyne valley suggest they developed as ice-contact features. However, this remains tentative as no actual sections could be investigated.

6. Glacial lineations and drumlins (as recorded by earlier workers) are interpreted as ice advance bedforms because the event stratigraphy has not been established for these landforms.
7. The absolute age of the outwash deposits, and therefore an event chronology, in the lower Tyne Valley remains unknown due to problems encountered with OSL dating (i.e. poor quality quartz).
8. The sedimentary sequence in the Tyne Valley represents stagnation chronology of the landform-sediment assemblages in the Tyne Valley remains unknown.
9. Alluvial paraglacial response is reflected in the aggradation of glacigenic valley fill sediments and alluvial fans, both of which remain undated.
10. River terrace development began with the formation of T4 during the Lateglacial/early Holocene period (11.4-9.2ka cal. BP). There is no evidence for earlier fluvial activity during the Bølling-Allerød period.
11. There is evidence for rapid terrace development during the early Holocene period through the incision and aggradation of two terraces during a 4k year period between 11 and 8ka cal. BP.
12. Fluvial reworking and sediment yield showed an initial increase in erosion during the early Holocene, with an exponential decline to the present day. This is interpreted to represent both rejuvenated and renewed paraglacial response, and is linked to climatic instability and anthropogenic activity during the Holocene whereby glacial sediments have been remobilised or transferred within the system.
13. The terrace sequence reflects baselevel change since deglaciation. Terrace longitudinal profiles do not appear to record the low (and rising) postglacial sea-level (see section 2.3.1.1.), although relative sea-level change is recorded in the lower valley alluvium (e.g. RSL index points). However, baselevel change

is indicated through the continued incision of ~20m by the river during the last ~11.4k years. This was largely driven by glacioisostatic uplift and, to a lesser extent, the influence of climate change and anthropogenic activity.

Chapter Six

Deglaciation of the Tyne Valley and early postglacial response

6.1 Introduction

This chapter presents a discussion of the new evidence from the deglacial sequences in the Tyne Valley and proposes the mode of deglaciation based on this data. Existing models for regional deglaciation are presented and evaluated in light of the new findings, and a regional model of deglaciation is discussed. Results are set in the context of the response of the BIIS during deglaciation. The problems associated with deriving a date for northern England and the lack of a chronological framework is presented, with possible solutions offered. Existing dates for deglaciation, derived from the margins of the BIIS, are discussed with implications for the mode of deglaciation. Paraglaciation and the apparent hiatus in the sedimentary record between the end of deglaciation and the Younger Dryas period is explored. Early postglacial river response is discussed in the context of the terrace sequence, and the various mechanisms for terrace development and their mode of formation are evaluated. The chronology for early postglacial river terrace development is compared to the climatic record (e.g. GRIP), and the results from the Tyne Valley river terrace sequence are set against the wider UK upland pattern of alluvial development and the possible cause of differences between the sequences is discussed.

The chapter is concluded by returning to the aims, hypotheses and conceptual models set out in chapter 1. A summary review of the findings of the research is presented to illustrate the aims of the thesis have been met, to ascertain which hypotheses should be accepted or rejected, and whether the conceptual models are an accurate representation of the mode of deglaciation in the Tyne Valley.

To provide a context for the following discussion a summary of the sequence of events across the Tyne Basin during and following deglaciation is provided in Table 6.1, which gives uncalibrated (yrs BP) and calibrated (cal. yrs BP) dates and the associated landforms developing at that time.

TABLE 6.1: Relative chronology and sequence of landform-sediment assemblage development along the lower South Tyne and Tyne Valleys.

RELATIVE DATE	SOUTH TYNE VALLEY	TYNE VALLEY	LOWER TYNE VALLEY/COASTAL ZONE
~21ka BP	Active ice continues to flow along the valley	Downwasting and active retreat. Ice-contact landforms develop	Glacial Lake Bradley and Crawcrook complex develops
~19ka BP	Downwasting ice, stagnation. Ice-contact landforms develop	Proglacial and paraglacial reworking of sediments. Outwash terraces form	Coastal ice impounds lowlands. Glacial Lake Wear forms
~15-11.5ka BP	Unknown development	Unknown development	Unknown development
~10ka cal. BP	T4 develops in response to incision and aggradation due to climatic instability and decline in glacial sediments towards the end of the paraglacial period		
~8ka cal. BP	T2 forms through continued instability. River Tyne becomes entrenched in its own terraces		

6.2 Models of deglaciation

There are two modes by which deglaciation can take place:

(1) Active retreat, whereby ice gradually retreats in stages but continues to be nourished by snowfall. If the ice periodically ceases movement or re-advances ice marginal landforms (moraines) or sediments (stratigraphy) develop, which may be

linked to climatic oscillations, such as those induced by subsequent Dansgaard-Oescher and Heinrich events (cf. Bond *et al*., 1993); or

(2) Stagnation and *in situ* downwasting, where there is no order to decay and downwasting is widespread across the ice mass (cf. Ward and Rutter, 2000). Local ice stagnation results in ice becoming progressively confined in valley bottoms. As the ice downwastes *in-situ*, recessional moraines are not formed. Once the ice has become stagnant it is disconnected from the regional climate and if sediments were dated they would produce a local history of ice decay.

The traditional view of deglaciation of the BIIS advocated by early Quaternary workers was of active retreat with the ice becoming confined to the valleys then in corries before finally disappearing (Raistrick, 1931). The assumption was that the ice front was dynamic, and both recessional and readvance stages were often identified from the sedimentary sequences. These ideas were largely informed by the Alpine models. The active retreat mode of deglaciation pervaded research until the 1970s when there was a paradigm shift towards downwasting and stagnation of the ice in response to temperature rise associated with rapid climatic amelioration (Sugden 1970; Coope and Brophy 1972; Sissons 1976; Coope 1977).

The mode of deglaciation across the BIIS varied and this reflects complex response and local conditions (both within the ice and topographic setting), both of which played an important role once deglaciation has begun. The local complexities that arose within the BIIS during deglaciation are demonstrable in two adjacent east-draining valleys in northeast Scotland, the Dee and the Don, where there is evidence of complex response in deglaciation. Aitken (1998) concluded the glacigenic sediments in the Don Valley formed through ice stagnation and *in situ* melting, whereas Brown (1994) interpreted the sequences in the Dee Valley as aggraded in an active retreat environment. Deglaciation cannot be interpreted by a "single model fits all" approach.

Through the work of McCabe and others, the onset of deglaciation and subsequent standstills and re-advances of the BIIS has been strongly linked to millennial and sub-millennial climatic oscillations in the north Atlantic climate (McCabe and Clark 1998; Knight 2003; McCabe and Clark 2003; McCabe *et al*. 2007). This response to climatic oscillations during deglaciation is recorded in ice-marginal sedimentary sequences and landforms on the Isle of Man (cf. Thomas *et al*., 2004) and in Ireland (cf. McCabe and Clark 2003). The mode of deglaciation was by actively responsive ice, which waxed and waned with the fluctuating temperature.

Previous models of deglaciation proposed for the Tyne Valley and Northumberland (outlined in section 1.2.2) advocate regional stagnation, downwasting and *in situ* stagnation (Clark 1970; Lunn 1995; Mills and Holliday 1998). Clark (1970) interpreted the deglaciation of Northumberland based on the regional stagnation model, supporting active ice downwasting and local stagnation. This mode of ice stagnation is encompassed by the stagnation zone retreat (SZR) model, which represents the interplay between both active and downwasting ice. Currier (1941) first introduced the SZR model based on landform assemblages in Massachusetts, USA related to the disappearance of the Laurentide Ice Sheet. The SZR model recognises that whilst the ice-margin stagnated the ice centre remained active. The model was later adopted by Mulholland (1982), who suggests that active ice shears a stagnant front that then downwastes to a critical thickness below which valley ice tongues can no longer support movement and the shear zone moves up-ice with a new suite of outwash deposits developing. Both Lunn (1980) and Mills and Holliday (1998) interpret deglaciation in the Tyne Valley as widespread downwasting of ice from the high ground with progressive lowering of the ice surface in lower areas. As it thinned, stagnant and dead ice masses became confined to the valleys. North Sea ice persisted along the coastal zone after the lowlands had become ice free, resulting in the development of Glacial Lake Wear (Smith 1981; 1994).

The evidence presented in this thesis demonstrates that the ice in the Tyne Valley decayed in both an active and stagnation (inactive) mode, and two phases of deglaciation have been recognised. From previous research, support for the existence of active ice during the onset of deglaciation in the Tyne Valley is found in the prevalence of subglacial channels (cf. Sissons, 1958, 1960, 1961). To the south of the River Tyne on the northern interfluves of the Northern Pennines there is a concentration of subglacial channels (cf. Raistrick 1931; Sissons 1958; Lunn 1980; Figure 6.1). These channels are discordant as they cut across the topography and tributary watersheds. The meltwater flowed in an ice directed, subglacial system; the divergence of the channels away from the main Tyne valley indicates the meltwater followed the pressure gradient and the thickness of the ice in the Tyne Valley. The ice slope was probably inclined to the southeast, following the main direction of ice outflow and the channels reflect the original flow direction from west to east, then southeast at the divide between the western and coastal ice lobes. The ice would have been thinner over the Northern Pennines than over the gentler relief of the south facing slopes of the Tyne Valley and in the valley itself, which explains why the network of channels developed on the Northern Pennines. These ice-directed meltwater channels are associated with the existence of a well-developed subglacial hydrology and a warm based ice sheet during deglaciation (Brown, 1993).

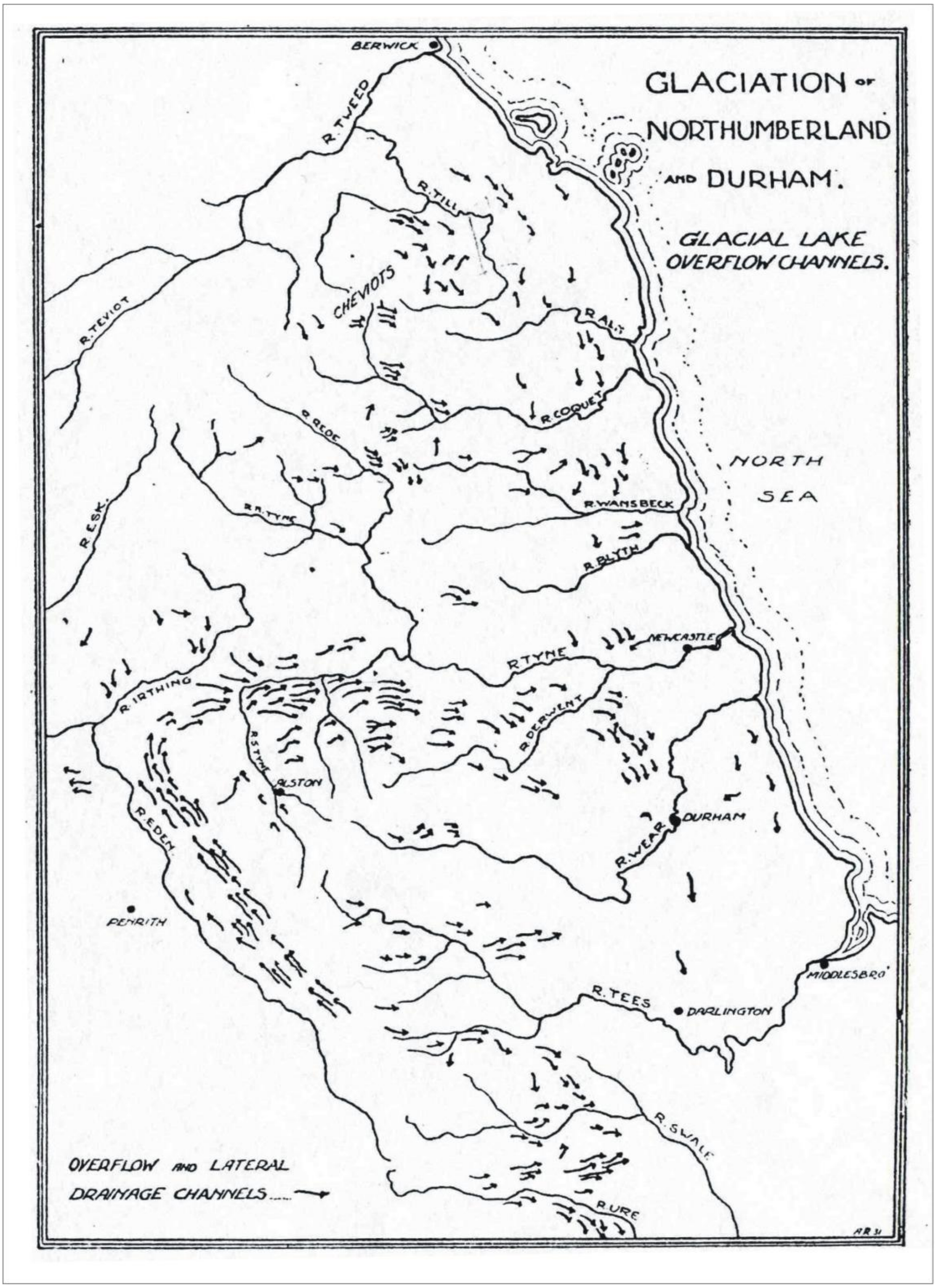


FIGURE 6.1: Raistrick's (1931) map illustrating the distribution of glacial lake overflow channels later re-interpreted by Sissons (1958) as subglacial channels.

The absence of moraines within the Tyne Valley (section 5.6.1) suggests retreat was either rapid (i.e. active) or by *in situ* stagnation. Assessment of IMAU borehole logs (Giles 1981; Lovell 1981) suggests there are two possible morainic sequences outcropping at Ovington, near Stocksfield, and Greenshaw Plain, east of Fourstones (Figures 5.10; 5.17). Without further investigation, the sedimentological evidence is equivocal as to whether the ice was wasting and stagnant or it was a still active ice mass. The deposits may simply represent the downwasting of ice *in situ* as it thinned and was confined to the valley bottom, rather than moraines. The sequences contain diamicton, as might be expected if the ice block was stagnating and releasing meltout tills as it collapsed.

The pattern of glacigenic sediments records evidence of ice marginal deposition and local centres of decay are recorded at Stocksfield and in the vicinity of Fourstones/Warden. The high sediment flux and chaotic style of deposition around these sites suggests an active ice margin (cf. Knight, 2003). The landforms were subject to post-depositional erosion by meltwaters. This was accompanied by the development of temporary glacial lakes as evidenced by the extensive sequence of laminated sediments (see section 5.2). The wealth of evidence for small glacial lakes in the Tyne Valley and its tributaries is unequivocal. Clearly the formation of temporary lakes has been a major factor of deglaciation in the Tyne Valley, and more widely across northeastern England. Meltwaters were dammed as thinning ice downwasted and temporary lakes formed. The development of proglacial lakes at the ice margin probably contributed to increased instability and wasting due to iceberg calving and lake down-draw destabilising the local ice-margin (Evans and Ó Cofaigh, 2003).

During the initial phase of deglaciation in the lower Tyne Valley a large ice-marginal lake (see section 5.2.1) developed, Glacial Lake Bradley. The sequence at the top of Glacial Lake Bradley was incised by channels infilled with coarse gravel indicative of subaerial exposure, and by basins infilled with thinly laminated silts and clays and dropstones indicative of ice rafting. Two-phases of development are proposed. In places the lake sediments onlap onto bedrock on the valley side (Figure 5.6), implying the lake was formed between downwasting ice in the valley and the ice-free valley side. Mills and Holliday (1998) suggest that, during early deglaciation, some lakes evolved from being totally subglacial, through a stage with significant floating ice cover, to a largely ice-free proglacial lake. Evidence from Glacial Lake Bradley partly supports this interpretation, with development at the ice margin and the formation of a subaqueous fan (see section 5.2.1). The sedimentological data suggests the lake was not supraglacial but ice proximal, as evidenced by the rainout from icebergs. The water surface level of Glacial Lake Bradley must have been at least 70m OD, based on the level of sediment accumulation and incision into the surface by channels; however, the basins contain dropstones indicating floating ice and reflecting a higher (unknown) water surface level. The damming mechanism for the lake was active ice in the lower Tyne Valley. Likewise in the Derwent Valley development of a glacial lake was attributed to ice in the valley bottom (cf. Allen and Rose, 1986). If the ice were stagnant the water level in the lake would have been subject to frequent fluctuations, as stagnant ice does not provide an impermeable dam; well-developed subglacial drainage could drain the lake (cf. Glen, 1954). There is no evidence in the sequence for draining/re-filling sequences. Rather, deposition appears continuous, providing additional evidence that active ice was involved at the early stages of deglaciation.

An ice margin is located on the Tyne/Derwent interfluve, evidenced by constructional landforms and dead ice topographies (Figure 6.2). Francis (1970) and Mills and Holliday (1998) have reported dead ice landforms (e.g. kettleholes and undulating topography) and stagnation (e.g. eskers) within the southeast quarter of the Crawcrook complex, which indicates deposition from active and stagnant ice at various stages of deglaciation. Whilst field observations from the sandur sequence do not provide direct evidence of ice contact, previous workers have inferred that ice tongues lay to the southeast of Crawcrook, based on the presence of ice contact slopes within the sedimentary sequence, in both the Derwent Valley (Allen and Rose, 1986) and the Blaydon Burn, a small tributary of the Tyne that dissects the complex (NGR NZ416562) (Francis 1970). This suggests the complex was ice-marginal. The presence of ice tongues (which probably developed as the high ground between the main Tyne Valley and its tributaries emerged, resulting in the ice sheet separating into discrete 'valley glaciers') is compatible with a topographic control model of deglaciation (cf. Mulholland, 1982) supported by Mills and Holliday (1998). The lack of similar ice contact landforms in the main Tyne Valley floor suggests stagnation was less prolific in the main valley.

There is evidence of a single esker, ~600m long and ~8m high, in Chopwell Wood (NZ 413 558) feeding this ice margin (Figure 6.2). Generally deposition at this margin is essentially confined to the southeast area of the Crawcrook complex in the Derwent Valley, and has taken place over dead ice in a stagnant zone (cf. Francis 1970; Mills and Holliday 1998). The deposition of till (probably supraglacial meltout) at the western extent of the esker suggests subglacial let down at the tunnel mouth, and the fall in height along the former ice line directions (southeast-south) probably reflects a time-transgressive nature of the esker as the ice margin retreated into the Derwent by

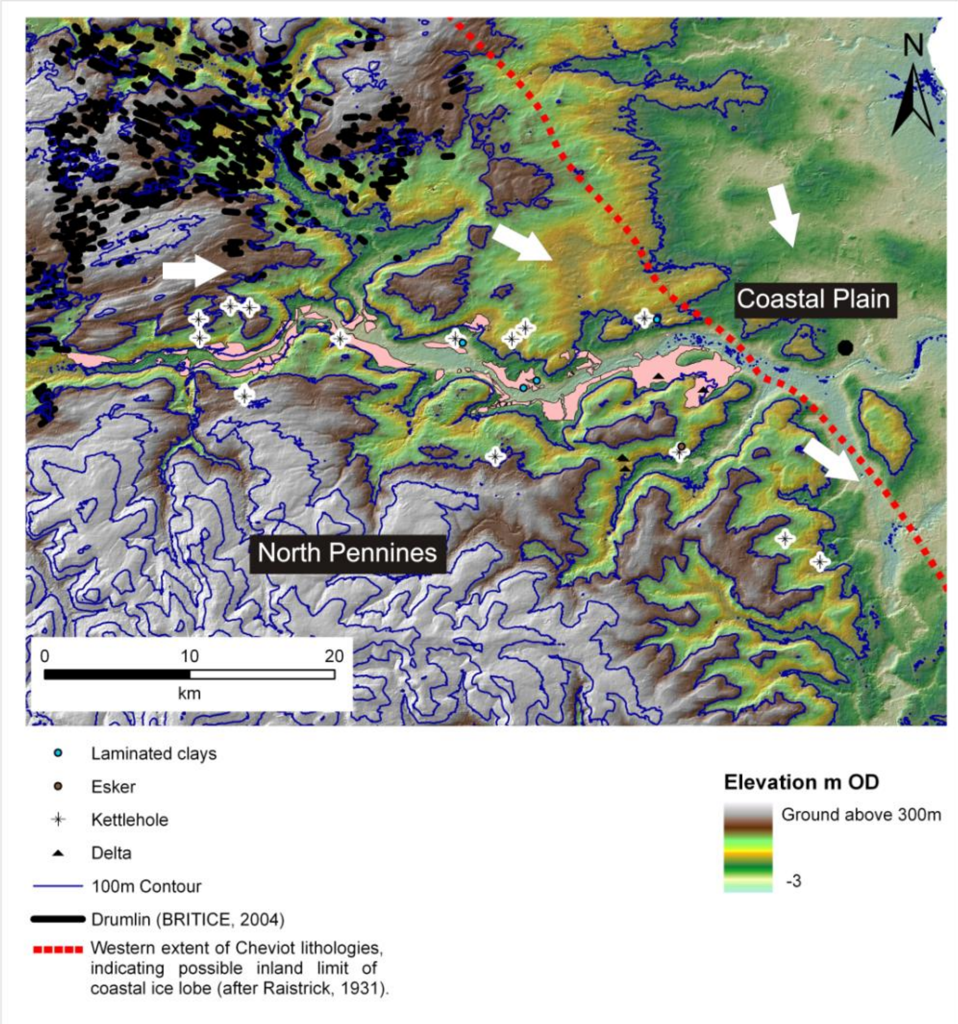


FIGURE 6.2: Map illustrating the distribution of ice-contact/marginal features (identified from a number of published sources and as part of this thesis).

stagnation zone retreat (cf. Delaney, 2002). The ridges southeast of Greenside (Figure 5.28B) are the only reported evidence of an esker in the Tyne Valley (Francis, 1970). The association of the esker with proglacial fluvial sediments was difficult for Francis (1970) to explain. Extensive quarrying prevented re-investigation of this sequence; however, the sequence suggests ice-marginal deposition associated with an active ice margin.

The formation of kettlehole basins is indicative of stagnant, isolated ice blocks that have downwasted *in-situ*, suggesting proximity to the shedding zone (cf. Mulholland,

1982) of the ice margin. The area around Hooker Gate, Derwent Valley (part of the Crawcrook complex), is reported to comprise a series of kettleholes (cf. Mills and Holliday, 1998). A possible kettlehole was identified (Figure 5.29) during the field mapping of the Crawcrook complex lower Tyne valley but could not be investigated. On the north side of the River Tyne, opposite the Crawcrook complex, a sequence of laminated clays and peat occurs in isolated basins at Houghton and Rudchester, and further basins occur northeast of Corbridge (Giles 1981; Mills and Holliday 1998) which might be interpreted as infilled kettleholes (Figure 6.1). The presence of a number of peat hollows (identified on the IMAU sediment maps cf. Giles 1981; Lovell 1981) along the Tyne Valley may represent kettleholes associated with *in-situ* stagnation (Figure 6.1), although only the basin at Warden has been clearly identified as a kettlehole by Lunn (1995). Further work is required to determine their origin and association with the landform-sediment assemblages.

The mound topography in the lower Tyne Valley (i.e. the Crawcrook complex) and lower South Tyne Valley has previously been interpreted as a stagnant ice contact landscape (see section 1.2.3). The BGS (Cox 1983; Richardson 1983; Mills and Holliday 1998) suggests the sands and gravels of the lower Tyne Valley complex aggraded in a predominantly subglacial environment from seasonal streams under/within stagnant or wasting ice. Subsequent collapse of the dead ice resulted in the disturbance of the surface expression of the sediments forming the mound topography. However, detailed examination of the sediments in the lower Tyne Valley in this current study points to a different conclusion. This thesis refutes earlier ideas that sand and gravel sequences were aggraded in a subglacial environment. Subglacial meltout or supraglacial flow till, which should be expected if the ice was stagnating, is absent. Where subglacially deposited sand and gravel deposits occur in Canada they are found in associated with subglacial till (cf. Munro and Shaw, 1997). The sediments are not subject to disturbance, e.g. massive faulting which would be expected if the sediments were deposited over stagnant ice and subject to let down as the ice melted out. The faults recorded in the sequence are minor and insignificant, most likely to be related to localised adjustment of water saturated sediments. The lack of major faulting is further evidence that active ice was responsible for deposition. The earlier development of the ice marginal Glacial Lake Bradley within the Crawcrook complex does not fit with later subglacial deposition of sand and gravel. Based on the current exposures and IMAU borehole logs from the Tyne Valley, the BGS interpretation cannot be supported.

Flat-topped features within the Crawcrook complex, one associated with the Glacial Lake Bradley sequence (Figure 5.31) and the other close to Barlow Burn, east of Greenside (NZ 415 562), may represent ice-contact delta or outwash fan formation in a lake impounded by active ice in the valley. Evidence for the delta is tentative because there are no exposures available, and although the flat surface at Glacial Lake Bradley is associated with a steeply dipping slope, the surface at Barlow Burn is not. Subsurface data is limited (Figure 5.6) but suggests deposition in a subaqueous environment. The flat surfaces may simply represent non-dissected remnants of sandur. Roberts *et al*. (2007) reported a similar sequence on the Isle of Man, which they attributed to ice marginal sandur and postglacial incision.

The localised occurrence of diamicton overriding part of the sands in the Crawcrook complex suggests the sequence was ice proximal, although the provenance of the till remains to be discovered. Hence there is evidence of active ice in the vicinity of the Crawcrook complex post aggradation of the sandur sequence (see section 5.2.1; Figure

5.7). There are two possible explanations for this: (1) it represents local ice margin fluctuation and so implies the persistence of active ice; or (2) it represents fluctuation of the coastal ice margin. The latter would imply that the lower Tyne Valley became ice-free very early during deglaciation and coastal ice 'bumped' into the complex as a response to regional climate oscillations before retreating and allowing Glacial Lake Wear to develop.

There is no relationship between the surface morphology of the lower Tyne Valley landforms and the underlying sediments. Whilst the sediments comprise proglacial and ice-contact fluvial and lacustrine sequences, the surface expression does not relate to those environments. Erosion of the glacigenic sediments into mounds was probably penecontemporaneous with the latter stages of deglaciation by meltwaters (down-valley directed) or even post-glacially (down-slope directed). This flow direction implies ice directed flow and the continued presence of active ice in the South Tyne Valley. The extent of small, dry valleys and misfit streams parallel to the ice flow direction that dissect some of the complexes is indicative of drainage associated with the present day network.

The laterally extensive glacigenic complex between Thornborough and Farnley Gate, mid Tyne Valley, which comprises laminated clays and a sequence of gravel and sands, has also been interpreted as an outwash plain. Some of the sediments were probably aggraded subaqueously at times, with localised meltwater ponding due to impoundment against bedrock and ice. However, the extent and thickness of the deposit is indicative of active ice in the lower South Tyne Valley continuing to deliver sediments and meltwaters, rather than stagnant ice downwasting in the valley bottom.

The presence of a large outwash plain implies a continuous supply of sediment and meltwater at the ice margin. The ice-margin was active as wasting ice would be supply limited (cf. Brown, 1993). The glaciofluvial sediments in the Crawcrook complex were not deposited over ice, but formed as a sandur plain, which became increasingly distal to an active ice mass as it retreated up the Tyne Valley.

Within the South Tyne Valley the ice continued to stagnate *in situ*. There is little evidence of sediment benches on upper valley sides to indicate high level glaciofluvial ice contact or outwash terraces in the lower South Tyne Valley. However, downwasting of ice from the higher ground, where it would have been thinner, is indicated by the absence of sand and gravel deposits on the interfluves and high valley sides. These are covered in till, and can be accounted for in the topographic control model proposed by Mulholland (1982). Whilst the landform-sediment assemblages that flank the South Tyne Valley were not directly examined due to problems of exposure, re-appraisal of subsurface data from the South Tyne Valley IMAU subsurface record (Lovell, 1981) and topography provided evidence to suggest that they formed in an ice contact environment (see sections 5.2.5; 5.6.2) as Lunn (1980) has proposed. A single kettlehole was found in association with the deposits at Warden, indicating, at least locally, the presence of dead ice (Lunn, 1995). The sequence in the South Tyne Valley is interpreted as ice-contact and the development of the deposits is envisaged as progressive retreat through stagnation. It is probable that these deposits terminate at an ice-margin to the west of the present day watershed. The sequence can be traced beyond the lower South Tyne Valley west through the Tyne Gap and into the Brampton area where an impressive complex of sand and gravel deposits outcrop but have yet to be fully investigated (although work is currently underway by the University of Durham). A similar sediment sequence in the Tweed

Valley was traced up valley and shown to terminate at an ice-margin (cf. Rhind, 1969). There is a need to investigate further the landform-sediment assemblages in the South Tyne Valley and Tyne Gap to understand fully the sequence of events during deglaciation, and to establish whether an ice margin was located in the vicinity of the North-South Tyne confluence.

Associated with the rapid retreat of ice from the lowlands, an ice-free zone developed in the mid/lower Tyne Valley. This is characteristic of deglaciation in northeastern England and led to a second phase of development. Following retreat of ice from the lower Tyne Valley the most significant glacial lake identified in northeast England developed (Glacial Lake Wear), which extended over the lowlands of the rivers Tyne and Wear from the present day coastline to ~20km inland (cf. Smith, 1994). The damming mechanism for Glacial Lake Wear was the persistence of active coastal ice at a time when the lowlands were ice free but were receiving substantial quantities of meltwaters and sediments from an active ice margin located in the lower South Tyne Valley. The persistence of coastal ice may be a reflection of the size of the ice lobe extending off the (present day) coast and that it continued to receive nourishment from the Scottish lowlands (Everest *et al*., 2005). Within the Team Valley, which is encompassed within the limits of Glacial Lake Wear, laminated clays up to ~60m thick have been recorded (Smith 1994; Mills and Holliday 1998). This suggests prolonged deposition in the distal part of the lake basin; indicating persistence of the lake over a period of time. In the lower Tyne Valley, sequences of laminated clays between 10 and 17m thick have been recorded (Land *et al*. 1974; Passmore *et al*. 1992). Borehole logs below Newcastle, recorded as coal mines were opened, show sequences of ~55m thick of intercalated sands, laminated clays and diamicton; the sequences are sandier and contain boulders at the present day coastline (Land *et al*., 1974). The occurrence of interbedded diamicton suggests either ice-rafting or marginal fluctuation; however, the sequence may simply reflect the complex development of a subglacial till (cf. Evans *et al*., 1995). Sequences of laminated clays are also recorded in the River Blyth and its tributary the Pont (Figure 2.12) (cf. Young *et al*., 2002), indicative of the extent of glacial lake development during deglaciation. Whether these deposits formed due to ponding as a consequence of coastal ice, or more localised ponding as the ice masses downwasted, remains to be disentangled, although the former is the most probable interpretation given the existence of Glacial Lake Wear, which was impounded, by coastal ice (cf. Smith, 1994). The examination of boreholes in the lower valleys along the Northumberland and Durham coast in closer detail may provide further clues on the development of glacial lakes and the behaviour of the coastal ice sheet margin during deglaciation.

6.2.1 Retreat of Tyne Valley ice and persistence of coastal ice: a possible mechanism

The exact mechanism and cause of ice sheet retreat and decay in the Tyne Valley at a time when the coastal ice was able to persist and surge down the coast is unclear. The western ice was fed from the upland areas of the Lake District and south west Scotland, and it seems unlikely that the decline in precipitation, warming temperatures and equilibrium line rise associated with the onset of deglaciation would not also affect the northern (coastal) ice fed from the Cheviots and Scottish lowlands. This thesis proposes an alternative explanation. There is geomorphological evidence for ice streaming around the Irish Sea Basin from the Lake District and southwest Scotland. At the onset of deglaciation, fast ice flow and the formation of drumlins, identified in both Ireland and the Lake District, indicate that the Irish Sea Basin was the focus of ice sheet discharging (McCabe *et al*. 1999; Roberts *et al*. 2007). Ice in the Tyne Valley is thought to have been fed by an ice stream, *via* the Tyne Gap, flowing east (Beaumont

1971; Everest *et al*., 2005; Boulton and Hagdorn 2006, 2007). Fast flow flattens the ice surface, drawing ice and basal water from slower moving streams, which would effectively have a steeper surface slope (D. Rippin, personal communication, 2007). As a consequence of fast ice flow being directed from the Lake District west to the Irish Sea Basin, the Tyne Gap ice stream could have been switched off or subject to water piracy (cf. Alley *et al*., 1994). Contemporary ice stream switch off and water piracy has been widely reported in the west Antarctic ice sheet (Alley *et al*., 1994). The downdraw of the ice sheet in the Irish Sea Basin and the associated ice flow towards this area may have resulted in the capture of east flowing water and ice by rapid flow in the west. This would then have become self propagating, as ice stream flow to the Irish Sea Basin then destabilised the whole ice sheet to the east, altering its configuration and thickness. Flow-sets in the Tyne Gap and north Cumbria area indicate this change in flow direction to the west at a later stage of glaciation (S. Livingstone, 2007, personal communication). The switch off/diversion of the Tyne Gap ice stream led to thinning, lowering of the ice surface profile below the equilibrium line, and in combination with a rise in the equilibrium line altitude due to deglacial warming, led to retreat and stagnation in the lowlands. The loss or capture of ice sources from the west effectively resulted in the Tyne Valley and North Tyne Basin being starved of ice. There is no other source that could have continued to feed the Tyne Valley ice and, thus the area was isolated from active ice.

The coastal ice was part of a larger ice lobe and was connected to the Tweed Ice Stream, draining ~3500km^2, which responded to the climatic change by rapid discharging of its ice mass, as evidenced by the extensive subglacial landforms in the Tweed Valley (cf. Everest *et al*., 2006). The coastal ice may have been less affected by initial deglacial warming due to the extent of the ice lobe modifying the climate and air circulation and dampening the impact of warming (cf. Bennett and Boulton, 1993). Thus, although the coastal ice was probably thinning and lowering, it was connected to active ice sources flowing out of central Scotland and the highlands, which were less affected by the initial climatic amelioration in Scotland, and continued to flow actively down the coast, maintaining a marginal position at Holderness/North Lincolnshire.

The development of ice-free conditions in the Tyne Valley lowlands occurring before deglaciation of the coastal zone can be explained by this model of ice thinning and ice stream capture/switch off. Thinning began as a response to the rapid climatic amelioration at the onset of deglaciation, resulting in increased ablation. Work elsewhere has shown that fast flowing ice zones are quick to respond to climatic or eustatic change (McCabe 1996). In combination with this, the emergence of the higher ground due to thinning and the associated rise in the equilibrium line, the ice retreated in a stagnant zone. The development of an ice-free zone between coastal ice and the uplands is also observed in the Isle of Man (cf. Roberts *et al*., 2007) and was attributed to thinning induced by rapid marine downdraw and fast ice flow in the Irish Sea Basin. This hypothesis proposes a plausible mechanism for retreat of Tyne Valley ice and persistence of coastal ice. The thesis suggests that stagnation that took place in the Tyne Valley is a direct response to millennial scale climate forcing or a H1 event. The idea is couched in the context of the regional evidence and responses elsewhere in the BIIS. If drumlinised bedforms in the Tweed Valley could be dated it may help to disentangle this story. The drumlins may relate to climate-induced surging events of the Tweed ice stream during deglaciation, and if so could be correlated with other dated events recorded across the BIIS. This may constrain the timing of the coastal ice lobe, and thereby the timing of the development of Glacial Lake Wear in the lower Tyne/Wear Valleys.

6.2.2 Proposed model of deglaciation in the Tyne Valley

Based on the evidence presented in this thesis, it is proposed that the pattern and mode of deglaciation is complex, involving both active retreat and stagnation; deglaciation was characterised by this contrast. There were three-phases of development in the Tyne Valley: (1) deglaciation of the main/lower Tyne Valley as the ice stagnated due to ice capture in the Irish Sea Basin; (2) persistence and/or renewed glacial advance of an ice lobe down the coast resulting in major glacial lake development in the lowlands; and (3) regional deglaciation as climate ameliorated and the BIIS finally disappeared (Figure 6.3).

The extent of the proglacial outwash suggests retreat was rapid to the confluence of the North-South Tyne, where ice proceeded to retreat in a stagnation zone retreat mode, transmitting meltwaters and sediments to the lower Tyne Valley. The sequence at Stocksfield may represent a shedding zone as there is evidence for ice contact and marginal sediments, or it simply reflects a lobe or tongue of ice remaining in the tributary valley. However, marginal sediments occur on both sides of the valley, which lends support to the former interpretation.

The first phase involved the lower Tyne Valley becoming ice free through rapid, active retreat and stagnation zone retreat as evidenced by the absence of end moraines and localised dead ice features in the Derwent Valley. The ice mass in the lower Tyne Valley separated from coastal ice probably as a consequence of both the rise in the equilibrium line altitude above the ice surface elevation (cf. Knight, 2003) and in response to the increased fast flow in the central BIIS, to the west of the region, which resulted in downdraw elsewhere. The lower Tyne Valley ice mass thinned and

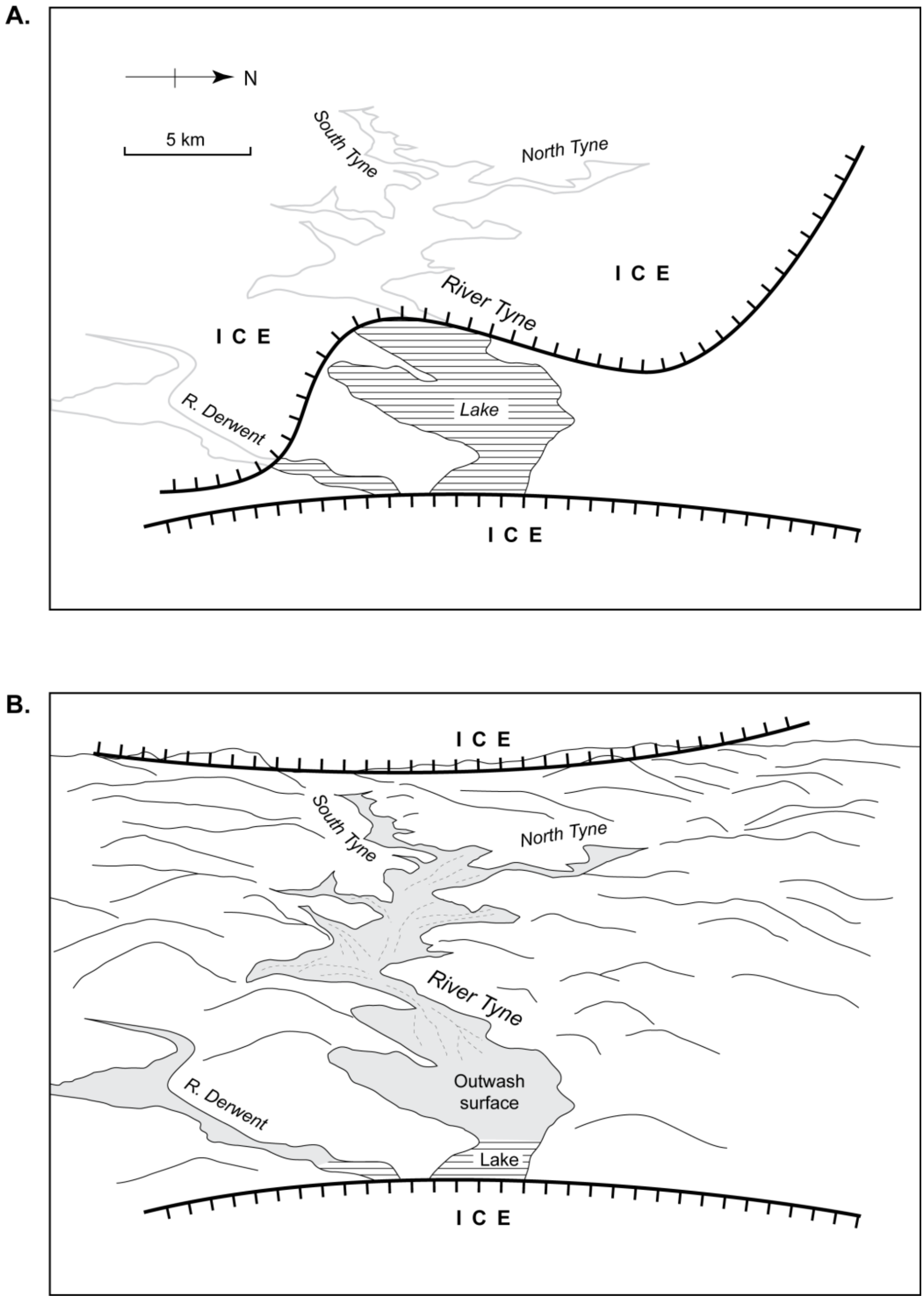


FIGURE 6.3: Cartoon depicting the development of the Tyne Valley during deglaciation. A. Represents the early behaviour of the Tyne Valley ice as is began to stagnate and disappear to the west, whilst the coastal ice (in the foreground) persisted. As the ice downwasted in-situ Glacial Lake Bradley developed in the lower Tyne in the vicinity of Crawcrook, and an ice-contact lake also developed in the Derwent Valley. B. Represents the later behaviour of the ice. The Tyne Valley is now starved of ice as the ISB ice-stream has captured the ice flow (located on the present day watershed). The Tyne Valley is infilled with sandur but the outlet remained blocked by coastal ice, resulting in the impoundment of the lower valley and development of the extensive Glacial Lake Wear.

downwasted, resulting in the formation of ice marginal lakes and fan development. Topographic control (cf. Mulholland, 1982) became more important as the interfluves emerged through the ice, and the ice mass separated into tributary and main valleys as a series of ice tongues. Evidence for nunataks has not been recorded, but there is evidence of a severe periglacial climate recorded by ice wedge features in shales above the valley floor at Slayley (cf. Clark 1970; 1971), suggesting exposure early during deglaciation. Topographic control is indicated by the occurrence of sand and gravel deposits on the valley sides and not on the interfluves (which were covered in till). As the ice margin retreated into the Derwent valley there was some esker formation associated with the temporary glacial lakes. Dead ice and supraglacial deposition is limited to the ice margin in the Derwent Valley, where features are spatially discrete and developed as the ice retreated into the valley to form an ice tongue. Extensive outwash is restricted to the main Tyne Valley, deposited when the ice in the South Tyne Valley was still active. The lack of outwash deposited in the lower South Tyne Valley suggests that by the time the South Tyne Valley became ice-free the ice was stagnant, and meltwater and sediment discharge from the ice-margin was supply limited. Development during early deglaciation of the Tyne Valley, therefore, reflects a dynamic period where ice retreated but remained active in the South Tyne Valley, as evidenced by the high and continuous supply of meltwater and sediment supply, leading to the build up of outwash in the lower Tyne Valley. It is possible that features of the stagnation zone have simply been buried or even removed by early postglacial incision and reworking. The truncation of the lake sequence by outwash (between locations 1 and 2, Figure 5.31) is indicative of this erosion of earlier sequences.

The second phase of development involved coastal ice ‘surging’ down the east coast during the later stages of deglaciation (~19-18ka cal. BP) (Lambeck, 1993). The ice-free zone, which developed in the Tyne/Wear lowlands during the first phase of deglaciation, became the focus for meltwater drainage and deposition from both the western and surging coastal ice, leading to the formation and persistence of Glacial Lake Wear. The ice margin blocked the outlet to the North Sea basin of meltwater from the South Tyne Valley ice and non-glacial runoff. Combined with this runoff, sediments and meltwater from the ice margin built up in the Tyne and Wear lowlands, leading to the development of Glacial Lake Wear. Clearly the coastal ice sheet encompassed the lowland areas of Northumberland, Durham and Teesside, as evidenced by the glaciolacustrine sediments ubiquitous in east draining rivers from the Wansbeck to the Tees (cf. Smythe 1907; Raistrick 1931). Whilst the coastal ice remained active it was able to maintain the lake. There is evidence to suggest the lake level did fall (cf. Smith, 1981; 1994), probably as the ice began to stagnate, but the timing of this is still unknown. Once the ice dam was breached, allowing Glacial Lake Wear to drain, it probably resulted in the incision of the outwash plains in the lower Tyne Valley. The final decay of ice led to meltwater erosion and (paraglacial) reworking of the glacigenic sediments upstream.

6.3 Regional deglaciation and implications

The lack of end or recessional moraines in the Tyne Valley is important in terms of reconstructing ice retreat behaviour. The end moraine at Dimlington, East Yorkshire represents a known position of the margin of the BIIS. The date obtained from the glacimarine sequence at Dimlington (Catt and Penny 1966; Penny *et al*. 1969) provides an indication of the timing of ice marginal fluctuating/surging in response to the onset of deglaciation across the BIIS. A recessional moraine sequence has been identified at Escrick, near York, related to the BIS, with others identified further north related to local valley ice (Clark *et al*. 2004). The lack of extensive moraines could imply: (i) they do not exist, thus favouring a regional stagnation interpretation; (ii) that deglaciation was rapid, i.e. without standstills so moraines did not develop; or finally (iii) the possibly that the features have not been identified because detailed mapping has yet to be carried out. From the Tyne sequence, and evidence from elsewhere in Northumberland and Durham (Smith and Francis 1967; Lunn 1980; Douglas 1991; Mills and Holliday 1998), it seems clear that retreat was rapid; although there is evidence to suggest local stagnation of the ice mass, creating ice-free lowlands bounded by the persistence of the coastal ice lobe. The presence of coastal ice after western ice had begun to downwaste and disappear from the Tyne Valley implies that there should be some deposits of ice marginal sediments, morainic dumps or ice contact landforms other than the lakes. However, the areal extent of lakes in the lowlands of the valleys suggests that the ice mass was located at or just off the present day coastline, so that any earlier morainic dumps or marginal features may already have been buried by the lake sediments.

The divide between the coastal ice lobe and the western ice mass has not been established; little evidence has been found either through this study or in earlier work. Clark (1970) proposed that the inland ice margin of the coastal ice lay along the high ground that borders the coastal plain (Figure 1.3), ~20km inland from the present day coastline. Evidence for this is still to be recovered, and requires the investigation of the tills on the coastal plain (currently underway by the University of Durham) and those which border the high ground to characterise the ice sheet signatures. The best evidence to date for the inland margin of the coastal ice comes from a map (Figure 6.4) published by Raistrick (1931): the erratic limit of Cheviot clasts (which characterise the coastal ice) was demarked based on detailed 'boulder' studies that were carried out in the early part of the last century. The diamicton that overlies sands in the Crawcrook

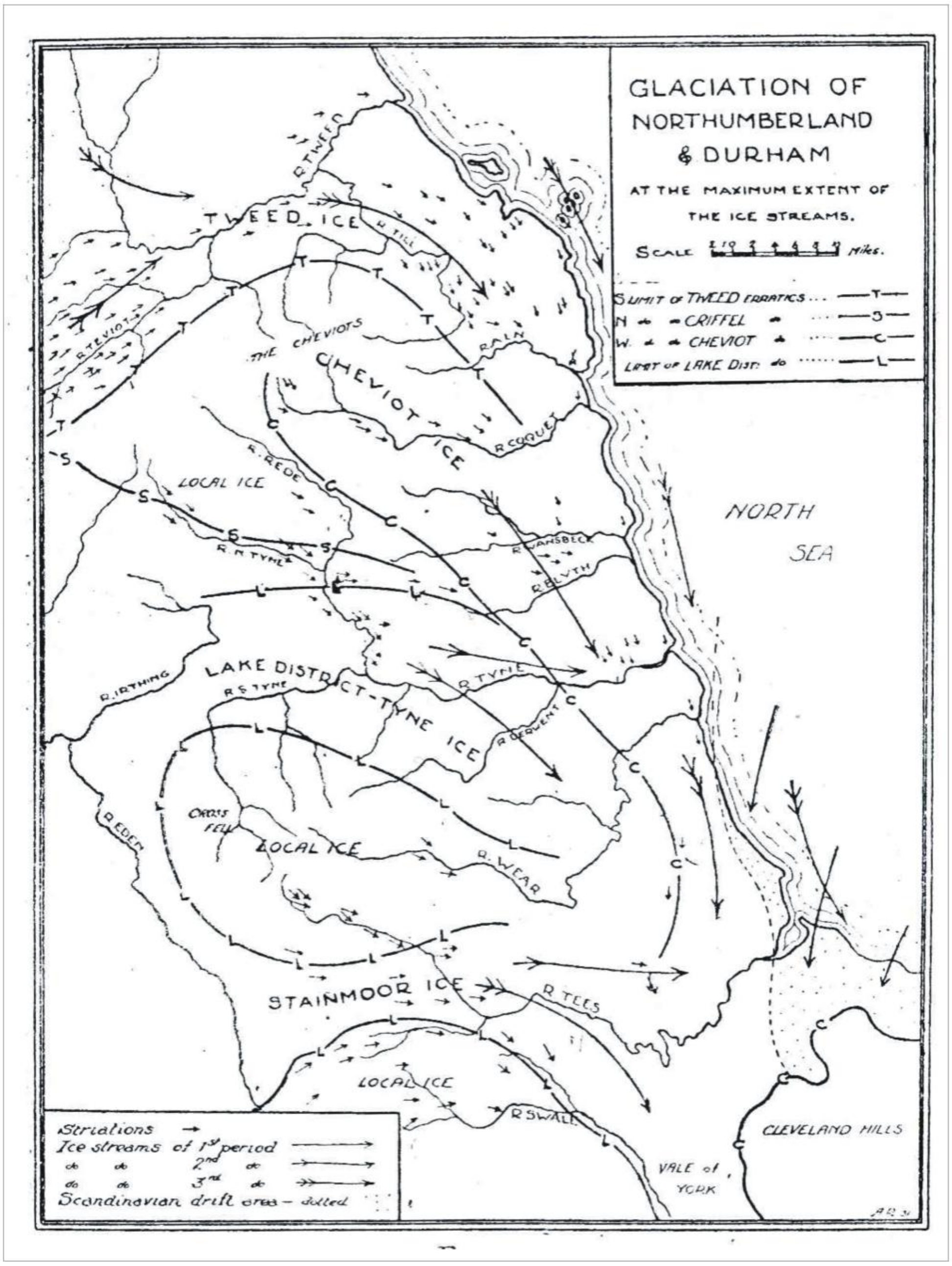


FIGURE 6.4: Raistrick's (1931) map outlining the erratic limits of the various ice streams (Cheviot, Tweed, Lake District, North Pennine, Stainmore) that flowed across northeastern England.

complex identified in this study (see section 5.2.1; Figure 5.7) may provide potential evidence for the inland extent of the ice lobe; this has yet to be examined in detail.

The mode of deglaciation by stagnation zone retreat (SZR) as proposed in the Tyne Valley may be applicable more generally across the northeast of England. Clapperton (1971) identified an assemblage of ice-contact features on the eastern flanks of the Cheviots and Milfield Basin in north Northumberland that included kettled moraine, eskers, ice contact terraces, kettleholes, lake deltas and thus the existence of temporary lakes, all of which point to stagnation zone retreat and *in-situ* downwasting. Whilst ice downwasted in the Cheviot valleys it persisted to the north in the Tweed Valley and costal zone creating an ice-free zone in which lakes formed, with the largest in the Milfield Basin (Clapperton, 1971). Chronologically this was probably synchronous with deglacial sedimentation in the Tyne Valley. With north Northumberland/Scottish lowlands under coastal ice, which was still active and surging, downdraw may have propagated deglaciation in the area. This mechanism has been suggested elsewhere (cf. Stokes and Clark 2003; Clark *et al*. 2003).

6.3.1 North Atlantic climate change and terrestrial response

Contemporary with renewed interest in the behaviour of the BIIS during deglaciation has been the identification of ice streams and the recognition of their importance during deglaciation (Clark and Stokes 2001; Stokes and Clark 2001; Ó Cofaigh *et al*. 2002). Subglacial bedform structures that indicate fast ice flow include drumlinised bedforms, streamlining and basal erosion. Fast ice flow is a symptom of high frequency discharge events when the ice is shedding its mass due to climate driven changes in the ice, and drumlin bedforms have been interpreted as events indicating pulses of extremely rapid environmental change (McCabe, 1996). McCabe (*op cit*) highlighted the poor quality of existing deglacial models in northern Britain, pointing out that while landform-sediment assemblages exists (such as cross-cutting relationships in drumlins and facies assemblages) which may help to understand fully the response, they are yet to be investigated. These types of features and evidence relate directly to climatic oscillations and if a chronology could be established they would be useful in linking events to climatic shifts that punctuated the last deglacial cycle (cf. Bond and Lotti, 1995).

Landform-sediment assemblages in Ireland, which have been dated, suggest the waxing and waning of the Irish Ice Sheet was in tempo with millennial scale climate change in the North Atlantic recorded in the GRIP and GISP ice core $\delta^{18}O$ ($^{0}/_{00}$) records (Grootes *et al*., 1993). To establish any link between the events during deglaciation and climate change in the North Atlantic, marginal terrestrial landforms or sediment sequences (e.g. drumlinised sequences and moraines) that record glacial readvance, surging or recession need to be investigated, as these may be directly linked to climatic change. Whilst McCabe and others have been successful in developing event stratigraphies from sequences in Ireland that have been correlated with millennial-scale climatic oscillations recorded in the north Atlantic marine and ice cores (McCabe 1996; McCabe and Clark 1998; 2003; McCabe *et al*. 1998), not all ice-marginal fluctuations need to be related to major climatic events to explain their behaviour (cf. Thomas and Chiverrell, 2007). In the Tyne Valley the ice-contact and proglacial sediments, which aggraded through downwasting and *in-situ* stagnation, could not be dated. The OSL dates (if successful) could have provided a comprehensive stagnation chronology for the Tyne Valley area, which may be related to a major climatic event (i.e. Heinrich event 1, Dansgaard-Oeschger, MWP-1A) in the North Atlantic. Although there is a lack of known ice-marginal landforms or glacigenic sediments that may document any ice sheet recession or advances (cf. Evans *et al*., 2005), drumlins are known to be present in the North Tyne Valley (Frost and Holliday, 1980), the Tweed Valley (Everest *et al*. 2005) and, possibly on the north Northumberland coast south of Bamburgh (Carruthers *et al*., 1927), which might be related to North Atlantic climate-induced surging events.

The Tyne Basin drumlin field was identified in the North Tyne Valley with a ‘flow-set’ indicating roughly NW-SE flow (Frost and Holliday, 1980; Figure 6.1). The drumlins have yet to be fully investigated in terms of their morphology and distribution pattern, which would give a clearer indication of their origin. They are capped by sand and gravels, which represents a post formational event, possibly as spreads of outwash formed after recession of ice, partly burying drumlins. The drumlin field might represent subglacial or ice-proximal deposition as the ice decoupled from its bed or as the ice retreated; however, without investigation it is still unclear if/how the drumlins relate to regional deglaciation. The drumlin bedforms in the Tweed Valley (cf. Everest *et al*., 2005) are probably related to climate oscillations in the North Atlantic, but as yet no event stratigraphy exists. To help disentangle and correlate deglaciation across the BIIS, this is one area that should be studied further.

6.4 Establishing a deglacial chronology and the relationship to other dates from the BIIS

The timing of the Late Devensian Glacial Maximum occurred around 21ka cal. BP (Alley and Clark, 1997); deglaciation then began and was characterised by ice retreat punctuated by a series of advances (McCabe and Clark 2003; McCabe *et al*. 2007) induced by millennial (~1500years) Dansgaard-Oeschger oscillations (Bond and Lotti 1995; Bond *et al*. 1997) and meltwater-induced Heinrich (H1 at 18ka cal. BP) events (Heinrich 1988; Bond *et al*. 1993). Based on recent cosmogenic and radiocarbon dates from the BIIS margin in Ireland and eastern Scotland, the onset of deglaciation occurred earlier that previously thought, before 17.1k ^{14}C yrs BP, and was punctuated by readvances at 14-15k ^{14}C yrs BP. The degree of synchrony across the BIIS confirms that ice retreat and advance was linked to climate. Deglaciation was terminated at 14.1k ^{14}C yrs BP with the onset of the Bølling-Allerød warm interval (McCabe and Clark, 1998). Although the location of BIIS immediately adjacent to the North Atlantic would have made it particularly sensitive to early warming (McCabe and Clark, 2003), and the rise in sea level around the margins that accompanied warming temperatures further enhance ice wastage through marine downdraw, terrestrial records of the response are needed to examine and date these events.

The precise timing of deglacial events in the Tyne Valley (and on the coastal sections) remains unknown. As discussed in section 5.3, several problems with the samples (such as weak or absent fast component) meant an OSL age could not be derived. The same problem arose in glacigenic sediments in the Swale-Ure Valley, North Yorkshire (Bridgland, 2006, personal communication), and it would appear that the local geology which comprises Carboniferous sandstones in the Tyne basin are unsuitable for OSL. Other avenues for developing a chronology need to be explored, and one possible route is cosmogenic nuclide analysis. Across the Tyne Basin and Northumberland, bare rock surfaces exist that are striated and were exposed following deglaciation. These surfaces potentially provide the opportunity to constrain the timing of deglaciation using cosmogenics, as has been applied elsewhere e.g. Scotland (cf. Everest and Kubik, 2006).

The bulk radiocarbon date obtained from the silts (glacimarine sequence) at Dimlington, East Yorkshire of 18.5ka BP (21ka cal. BP) remains the only known date for ice-marginal response in northeastern England. The sequence may provide evidence of ice surging down the east coast, probably linked to climatic events in the North Atlantic, but remains to be investigated fully. Although the Dimlington date coincides with other dates from the margins of the BIIS, there are issues with the reliability of the radiocarbon dating method used (bulk rather than AMS) and more dates are required to develop a robust chronology. Due to he lack of features and dated events in northeast England further work is required at a regional scale to disentangle fully the chronology of deglaciation in northeast England.

6.5 Paraglaciation and the sediment hiatus between deglaciation and the earliest Holocene

Following the onset of deglaciation the pattern of sedimentation is affected by a period of paraglaciation. Paraglaciation, as envisaged by Church and Ryder (1972), encompasses proglacial processes and the period following this comprising enhanced sediment yields due to glaciation until they return to ‘normal’ background levels (Figure 2.11). Ballantyne (2003) modified this slightly, indicating periods of both rejuvenated and renewed paraglaciation. It was proposed in Conceptual Model 1 (Figure 1.4) that immediately following deglaciation, slope instability would contribute to paraglaciation, but as the climate improved during the Bølling-Allerød Interstadial and vegetation colonised the slopes, there would be a decline and cessation of the initial paraglacial period (or proglacial period, see Figure 2.11A,B). The sedimentary evidence and sediment budget suggests that an initial paraglacial period continued from deglaciation into the early Holocene, and was characterised by two phases of sedimentation, with an apparent hiatus in the record during the intervening (Interglacial) period. Paraglaciation continued through the Holocene as glacigenic sediments were entrained and the river system responded to regional changes in baselevel (induced by isostatic uplift and RSL change).

Following ice retreat into the lower South Tyne Valley, proglacial rivers began aggrading the extensive sandur plains recorded in the mid and lower Tyne Valley. The aggradation of extensive sandur reflects the greatest period of paraglacial sedimentation (during the proglacial period, see Figure 2.11) in the Tyne Valley, and although there are no absolute dates for deposition, it must have taken place during early deglaciation based on the relative sequence of events. Trenching of the glacigenic sediments resulted from declining sediment supply and discharges associated with the end of the proglacial period (complete deglaciation). Incision was further enhanced by isostatic uplift. Reworking of outwash sediments through postglacial erosion and dissection resulted in a rejuvenated paraglacial response as glacigenic material were mobilised into the river.

The lack of landform-sediment assemblages identified or investigated from the Lateglacial period in upland glaciated basins in the UK represents a hiatus in the sediment record and creates difficulties when trying to elucidate the extent of continued paraglacial and alluvial response following deglaciation. However, work in the uplands of the Isle of Man, where solifluction terraces and alluvial fan development have been dated to between 18 and 10.5k radiocarbon years BP have been identified, indicate that these valleys were active and driven by paraglaciation during the Lateglacial period. Furthermore, paraglaciation must have continued into the Holocene as the rivers continued to occupy Lateglacial surfaces before incision/reworking during the late Holocene (Chiverrell *et al*., 2001). However, in general it is thought that the UK alluvial record is incomplete due to erosion of Bølling-Allerød Interstadial terraces or sediments (cf. Lewin and Macklin, 2003; Figure 2.8). This may be related to a combination of: (i) lack of mapping, i.e. features may exist but have not been identified; (ii) poorly resolved chronologies, thus features are not constrained within a recognised timeframe; (iii) a lack of response; or (iv) poor preservation and the complete excavation of Bølling-Allerød Interstadial alluvial sequences. Alternatively fluvial response in upland glaciated catchments might have been subdued, or the landscape insensitive to change during the Lateglacial, due to the greater magnitude of forces operating during the preceding glacial period i.e. landscape sensitivity (cf. Brunsden and Thornes 1979; Brunsden 2001) played a role. The propensity for change depends upon the sensitivity of the landscape; at any point this varies both spatially and temporally and is related to the ratio between the magnitudes of the resisting forces (i.e. barriers to change) versus those of the disturbing forces (i.e. high magnitude events). Thus geomorphic (internal and external) thresholds (cf. Schumm, 1973) may not have been crossed until the early Holocene because the magnitude of the previous glacial forces (i.e. landscape relaxation time) was greater than that of the subsequent fluvial processes (disturbing force) and continued to resist landscape changes.

Whilst it is unclear what was happening in terms of alluvial system paraglaciation (*sensu* Ballantyne, 2003) during the Lateglacial period, it is generally accepted that river systems were post-paraglacial at the start of the Holocene (Lewin and Macklin, 2003). Within the Tyne Basin there is evidence (suggested by a relief-shaded NEXTMap DSM of the area) of paraglacial activity (i.e. gullying of drift-mantled slopes, alluvial fans, valley fills) which probably reflects phases of paraglaciation following deglaciation. However, further work is required to accurately map, investigate and date these landform-sediment assemblages, and to disentangle the various aspects (i.e. alluvial, drift-slope, glacial foreland) of landsystem paraglaciation (cf. Ballantyne, 2003). The Tyne terrace sequence suggests both rejuvenated and renewed paraglaciation took place at the start of the Holocene and during the late Holocene as extrinsic factors e.g. climate, isostatic uplift, sea level change resulted in renewed connectivity between stored glacigenic sediments and the fluvial system. The continued entrainment of stored glacigenic sediments by the river during the early Holocene is evidenced by the coarse, glacigenic materials that comprise the terrace units. The River Tyne continued to transport glacially-derived sediments through the system until the river became entrenched within its own terraces (post 8ka BP), and lateral migration of the river was unable to access the glacigenic sediments which flank the valley sides. As the River Tyne continues to entrench its valley bottom, exploiting paraglacial and glacial valley fill sediments, renewed paraglaciation will occur, thus the influence of paraglaciation can be seen to extend to the present day and beyond.

6.6 Early postglacial alluvial response in the UK

In the context of the other findings, incision through the glacigenic sediments in the Tyne Valley may have taken place within the Younger Dryas period or even at the end of deglaciation and start of the Bølling-Allerød (~14.6-12.9ka cal. BP), rather than at the Younger Dryas-Holocene boundary as is generally assumed (Macklin and Lewin, 2003). Macklin and Lewin's (2003) paper shows there are major differences between the upland, glaciated catchments and lowland un-glaciated catchments in the UK in terms of the early Holocene alluvial record. Lowlands experienced greater valley floor stability and limited lateral reworking during this period; however, the record is sparse. The rivers responded to continued (postglacial) isostatic uplift and declining sediment supply by progressive, episodic incision through glacial/periglacial sediments during the late Devensian and early Holocene. Considerable lateral and vertical movement of upland rivers has continued throughout the Holocene. During the mid to late Holocene, incision is clearly related to both climatic change and anthropogenic disturbances in the catchment, and has been widely reported elsewhere (cf. Macklin, 1999).

6.7 Incision mechanisms and river terrace development

Macklin and Lewin (1989, 1993) and Lewin *et al*. (2005), based on reviews of dated alluvial units in British upland valleys, suggest that lateral migration tends to eliminate older sediment units as new ones are created, whereas incision leads to preferential preservation of older units beyond the incision period. It is generally accepted that in upland Britain, the model for fluvial development since deglaciation is reflected in the progressive incision of the valley floor (Macklin and Lewin, 2003). In the Tyne Valley incision would have been driven in response to the sharp decline in sediment supply and glacio-isostatic uplift at the end of the Devensian; subsequent Holocene incision is related to climatic change (i.e. large flood events), limited sediment supply and anthropogenic disturbance (Macklin and Lewin 1989; Macklin, 1999; Macklin and Lewin 2003). Lewin *et al*. (2005) suggest the valley floors of upland (glaciated) rivers developed after 6ka BP through episodic incision, due to higher sediment delivery rates from hillslopes and tributary streams, and larger floods (principally climate related) that have been superimposed on longer term incision trends, which resulted in the preservation of more alluvial units. However, in the Tyne Valley (and other valley affected by glaciation), there is a strong case for continued isostatic uplift driving incision throughout the Holocene.

6.7.1 Climate (discharge and sediment supply)

It is suggested that, during deglaciation (and the Younger Dryas), sediment supply and seasonal runoff (nival) resulted in river aggradation, burial and preservation of alluvial units, characterised by gravels (Lewin *et al*., 2005), reflecting both pro- and paraglacial response. Within the Tyne Valley, development during the deglacial period is reflected through the aggradation of proglacial outwash (e.g. Crawcrook complex) and subsequent incision of the deposits. Alluvial units have not been recognised from this period. Terrace 4 (T4) alluvial sediments truncate glacigenic sediments, indicating river down-cutting through the glacigenic sequence. Thus in the study area, the earliest floodplain must have comprised the newly exposed surface of the deglacial sediments. The river reworked the glacigenic sediments before incising through them during the Bølling-Allerød Interstadial; sediment export rates are high between the incision of outwash deposits and the formation of T4 (see Table 5.4). Macklin and others recognise that following deglaciation incision took place as sediment delivery declined, enhanced by glacio-isostatic uplift in the uplands. Terraces dating to the Bølling-Allerød are generally not found, and although alluvial units are recognised from the Younger Dryas in both the uplands and lowlands, dating control is poor in the uplands (Lewin *et al.*, 2005; see Figure 2.8). The results in this thesis confirm the lack of alluviation (or preservation) between the end of deglaciation and the Holocene. Whilst paraglaciation continued throughout this period, it may have been that little fluvial activity took place during this time as a consequence of the relaxation period associated with glaciation (cf. Brunsden and Thornes, 1979).

From the chronology developed for this study it is unclear when incision of the glacigenic valley infill took place, however OSL dates for T4 and Terrace 2 (T2) indicate that the three terraces (T4-T2) formed during the early Holocene period (see section 5.3). Approximately 10m of incision through the glacigenic terrace took place before 10ka BP. Although there is no absolute date for incision, it must precede development of the upper river terrace (T4). Alluvial sedimentation formed a veneer on the surface of the glacigenic infill and gave rise to T4 ~10ka cal. BP. Through the quantification of the valley infill, combined with the rudimentary chronology for terrace development, this suggests that sediment supply and storage was highest during the early Holocene period (see section 5.8.2.1). Early Holocene fluvial activity was quite dynamic with the most significant amount of fluvial development taking place during two incision and aggradation cycles. The sequence in the Tyne Valley is in contrast with general findings on a UK regional scale, where little evidence of activity is recorded during the early Holocene (cf. Macklin and Lewin 2003; Lewin *et al*. 2005). This may be because few or no alluvial units have been preserved due to valley floor reworking, or through lack of chronological control (Macklin and Lewin 1993; 2003; Lewin *et al*. 2005). However, there is some recent evidence to suggest early Holocene activity in the lowlands (Lewin *et al*. 2005; Macklin *et al*. 2005).

However, since publication of the Macklin and Lewin (2003) and Lewin *et al*. (2005) studies, Tipping *et al*. (2007) have identified terraces and periods of increased fluvial activity during the Lateglacial and early Holocene period in the Kelvin Valley, central Scotland. This illustrates that there may be further regional complexity in the UK alluvial record. Terraces of early Holocene age have been identified in the Cheviots (cf. Tipping, 1998), the Southern Uplands (cf. Tipping *et al*., 1999) and central Scotland (Tipping *et al*., 2007). The sequences in these areas comprise alluvial fans and terraces, underlain by reworked glacigenic sediments, inset below ice-proximal landforms. The landforms are thought to have developed during the Younger Dryas or early Holocene (Tipping 1998; Tipping *et al*. 1999). The formation of pre-Holocene valley floors suggests incision of ice-proximal landform-sediment assemblages took place in the centuries or millennia that followed deglaciation, associated with stabilisation of the landscape. Rejuvenated paraglaciation resulted in reworking of glacigenic sediments resulting in the formation of alluvial fans and terraces. The development of the Tyne Valley and other upland glaciated catchments probably developed on a similar timescale and pattern. Whilst there is no evidence for interstadial activity, valleys were most likely experiencing paraglaciation and it is probable that valley fills reflect these processes. More work needs to be carried out on these valley fill sequences which are likely to prove fruitful, shedding light on early postglacial responses.

The sedimentation style for the Tyne terraces (T4-T2) is characterised by channel and overbank stacked sequences, typical of a wandering gravel bed river. The sedimentological evidence suggests there was no significant change in fluvial style between phases of terrace development, although the sequence at Fourstones does comprise evidence of flooding and debris flow events (see section 5.2.4). Periods of quiescence are suggested in borehole logs (Figure 5.17) from T4, lower South Tyne, which reveal peat horizons. The presence of peat suggests either a decrease in overbank flooding during which the floodplain experienced stability, or may simply reflect distance from the main channel as it migrated across the valley floor leading to formerly active areas developing as 'backswamp'. This sequence requires further investigation as peat offers the potential for radiocarbon dating.

6.7.2 Baselevel (sea-level change)

Sea-level change following deglaciation of the UK was discussed in section 2.4.1; where a baselevel change takes place it may be translated upstream and recorded in river terrace long profiles. However, translation of low or rising sea-level into the terrace record depends on the persistence and magnitude of the change. Lambeck's UK glacio-hydro-isostatic model (1995) predicted a static baselevel between 18 and 14ka BP, and only after 9ka BP, when isostatic rebound outstripped eustatic rise, is an effective sea-level fall predicted. On the northeast coast of England, sea-level was predicted to lie between -50 and -75m below OD at *c*. 12ka BP, rising to -10m below OD at *c*. 7ka BP (Lambeck 1995). However, Shennan *et al*. (2006) have demonstrated that between 15 and 6.5ka BP relative sea-level continued to rise along the Northumberland coast (Figure 2.10). As sea-level rose, coastal shortening resulted in the submergence and burial of land off the present day coastline, and probable incision in the uplands. In the lower Tyne at Blaydon and Newburn, borehole logs record peat bands within the fine grained sediments marking the transition between the terrestrial sequence and the onlap of estuarine sediments and alluvial sedimentation, indicative of a positive sea level tendency (cf. Giles, 1981). At Blaydon, a RSL index point at -3.18m below OD marks a marine transgression at 7.7k ^{14}C yr BP (*c*. 8.5ka cal. BP) (Horton *et al.*, 1999a; Shennan *et al*. 2006), indicating that rising sea level reached the lower Tyne Valley during the early Holocene. The sedimentary record (e.g. IMAU borehole logs) indicates that the lower reaches continued to aggrade alluvial sediments.

Evidence of baselevel change in indicated by the formation of the terraces, which formed as glacioisostatic uplift induced a baselevel fall despite a rising sea level. The oldest terrace (T4) developed after ~11.4ka cal. BP, when relative sea-level is predicted to be ~-30m OD. As sea-level continued to rise, the River Tyne incised its valley fill in response to glacioisostatic uplift and a very low sea-level. It is likely the Tyne gorge and the characteristic ‘denes’ of other east draining rivers developed during late deglaciation, incising their lower reaches in response to baselevel change and (probably) initial isostatic rebound (see section 6.7.3). The smaller east draining streams, i.e. the ‘denes’, have steeper gorges because of the power required to incise them at a rate equal to that of uplift (cf. Merritt *et al.*, 1994). Horton *et al*. (1999b) propose that the nearby lower Tees Valley gorge formed following drainage of Glacial

Lake Tees in the lowlands, where incision was caused by the free draining river down cutting to the low sea level predicted by Lambeck (1995). Therefore, it is probable that drainage of Glacial Lake Wear and Lake Tees explains the incision of the lowlands of other east draining rivers on the northeast coastline, and the trenching of the glacigenic sediments that now flank the valley sides.

The terrace long profiles demonstrate that river terraces of the Tyne Valley remain in grade throughout the Holocene. This implies sea-level change has not been translated beyond the coastal fringe. Evidence from the lower Tyne Valley and estuary suggests that although there is evidence for rising sea-level, burial of older terrace surfaces by onlapping coastal sediments has not occurred. The surface of T3 (~8ka cal. BP) at Blaydon (cf. Passmore *et al*., 1992) still lies ~10m above present river level, and the terrace does not dip below onlapping sediments. Change in sea-level during the late Devensian and throughout the early-mid Holocene has not been translated upstream of the coastal fringe. The changes in sea-level that were taking place as the upper terraces were forming were not significant (i.e. of low magnitude) or persistent enough to register a change in the river's long profile. Relative sea-level change is recorded in the alluvial sediments that lie within the present day tidal limit, *c*. 24km upstream of Tynemouth.

6.7.3 Isostatic uplift

Differential isostatic uplift along the Northumberland coast is evidenced by sea-level index points, which illustrate that crustal uplift in north Northumberland was (and continues to be) more pronounced than in south Northumberland (Plater and Shennan 1992; Horton *et al*. 1999c). The models for north and south Northumberland (Figure 2.10) indicate relatively small isostatic rebound, recorded in the coastal estuarine/saltmarsh sequences (cf. Plater and Shennan, 1992) as a rising/falling sea level.

The height of the upper Tyne terraces above present river level suggests that a significant amount of incision has taken place during the Holocene (~20m). This incision trend is demonstrable throughout the Tyne basin and its tributaries, and in the smaller east draining rivers (e.g. Blyth, Wansbeck) on the Northumberland coastal plain where the terraces are a consistent height of ~20 and 10m above present river level and inset ~10m below glacigenic sediments. Even considering climatic instability and coastal shortening as a mechanism for rapid incision and aggradation during the early Holocene, it seems likely that isostatic rebound played some role, given that the ice was ~1km thick over much of the Tyne Valley (cf. Boulton *et al*., 1985) and the amount of incision was ~20m. The total depth of incision achieved since deglaciation is ~30m. There is no evidence of tilting in the long profiles, suggesting uniform uplift and a continuous trend of incision since deglaciation. Relative sea level records for Northumberland, south of Alnmouth, suggest isostatic rebound had ceased by the early Holocene, or was at least reduced, so that sea level continued to rise and transgress the coastal plain, inundating the lowlands. The scale of the early post-glacial incision suggests climate change alone is unlikely to be the cause of such a degree of incision. As has been proposed by Macklin and others (Macklin and Lewin 1993; Lewin *et al*. 2005), incision during the late Devensian is related to both glacio-isostatic rebound and changes in catchment hydrology induced by climatic amelioration. The mid-late Holocene incision trend is on a much smaller scale (<7m) and it is widely accepted to be related to both climatic instabilities and anthropogenic activities (cf. Passmore and Macklin 1994; 1997; 2001), perhaps indicative of the cessation of isostatic rebound. Perhaps the most compelling evidence for isostatic uplift driven incision is evidenced by the absence of any significant terrace development in those river valleys (e.g. Trent) that lie beyond the limits of the Devensian ice sheet (cf. Howard *et al*, 2007).

In the Tyne Valley, the river terrace surfaces (T4-T1) lie at consistent heights above present river level; they are in grade with the current profile and they remain the same height above present river level, above and below the North/South Tyne confluence knick point (Figure 5.40). Although knick points generally indicate the upstream propagation of base level change, the two knick points identified in the terrace long profiles are clearly not related to baselevel change. Because the location of both knick points remains stationary in all of the long profiles, they cannot reflect base level change as they would show propagation upstream in the successive profiles. Therefore, the knick point that exists at the confluence of the North and South Tyne is simply interpreted as an erosional one related to increased erosion at the confluence of the two rivers. The knick point that is located in the vicinity of Riding Mill (Figure 5.40) remains stationary over time, and the most likely interpretation is a lithological knick point related to the presence of a more resistant bed of Carboniferous sandstone.

6.7.4 Summary

The quasi-continuous incision trend in the Tyne that has preserved the upper terraces is a consequence of both isostatic rebound, change in sediment delivery and runoff related to climate instability, as well as (probably) a degree of coastal shortening in the early Holocene. The major baselevel changes that took place following deglaciation resulted in down-cutting through the glacigenic sediments by the river and formation of the steeply incised lower reaches. Although the bedrock channel of the Tyne Valley lies at least -46m below OD, it probably represents the late Devensian fall in sea level as the global ice sheets built up. The terraces of the Tyne Valley lie above the influence of any sea level change and developed after the major period of eustatic sea level rise was complete. At the start of the Holocene, terraces formed through the two cycles of incision and aggradation, which were driven by a combination of isostatic rebound and climatic instability. The early Holocene was likely to have been a period of 'dynamic equilibrium'. During the formation of the river terraces, eustatic baselevel change was not translated up valley; the influence of RSL change is only found in the lower reaches of the river, where onlapping sediments bury earlier fills. Based on the recent (i.e. mid-late Holocene) period, it would appear that isostatic uplift in the Tyne Valley is negligible, if not totally complete. However, evidence from Castle Eden Dene (Durham) suggests isostatic uplift is still continuing (cf. Evans, 1999) and thus, further work on the Tyne sequence should help disentangle isostatic uplift from climate driven incision, which is the current paradigm in Holocene valley floor development.

6.8 Thesis aims and conceptual models reviewed

Three hypotheses and two conceptual models were set out in chapter 1 that were informed by the existing literature, and were an attempt to draw together the landform-sediment assemblages and mode of deglaciation in the Tyne Valley into simple development overviews. These can now be reviewed and tested against the evidence.

6.8.1 Hypotheses

H(i): The landform-sediment assemblages in the lower Tyne valley were aggraded in a supraglacial context; they are depositional features.

EVIDENCE: The sequence in the lower Tyne Valley comprises proglacial and ice marginal sediments, which although developed in association with the complex of the ice, were largely aggraded in an ice free environment (see section 5.2.1). Their present mound-like morphology developed through erosion and dissection by meltwaters and re-establishment of drainage routes.

Hypothesis (i) is REJECTED.

H(ii): The highest terrace surfaces in the mid Tyne Valley developed as proglacial sandur were graded to the baselevel of Glacial Lake Wear and subsequently incised after the lake drained.

EVIDENCE: The upper terrace surfaces that flank both sides of the river in the vicinity of Farnley formed as proglacial outwash under both fluvial and lacustrine conditions (see section 5.2.3). However, there is insufficient data to correlate the long profile of the outwash surface with a lake baselevel.

Hypothesis (ii) is ACCEPTED WITH MODIFICATION. The highest terrace surfaces formed as proglacial outwash. However, there is no evidence to indicate they are graded to the baselevel of Glacial Lake Wear.

H(iii): The river terraces developed during deglaciation and their long profiles are graded to a baselevel below OD. They developed in response to paraglaciation and were incised due to isostatic uplift, declining sediment supply and hydrological change during the Bølling-Allerød Interstadial and Younger Dryas Stadial.

EVIDENCE: There is no evident to indicate the long profiles of terraces (T4-T1) are graded to lower postglacial sea levels as demonstrated from RSL index points recorded along the Northumberland coast. The profiles reflect the current river profile. Terrace formation was dated to the period after major global eustatic sea level rise was complete. The terraces examined suggest they formed beyond the zone where the influence of sea level change was transmitted (see section 5.7). However, the formation of a terrace staircase reflects changing baselevel. The ages of the terraces suggest they did not form during deglaciation but represent postglacial fluvial development (see section 5.6.3) in the Younger Dryas and early Holocene periods (~11.4-7.9ka cal. BP).

Hypothesis (iii) is REJECTED IN PART. The terraces developed after 11.4ka cal. BP, reflecting baselevel change but are not graded to a sea level below OD. They developed during the initial paraglacial period, which into the early Holocene, in response to isostatic uplift and changing climatic conditions.

6.8.2 Conceptual models

Based on the current evidence, neither of the two conceptual models proposed in chapter 1 (section 1.4.1) are a valid interpretation. Conceptual Model 1 (Figure 1.4) was built on the assumption that the landform-sediment assemblages in the lower Tyne Valley had developed in supraglacial or dead ice topography with the resultant landforms forming through ice melt out and let down. This thesis has demonstrated that although the sediments were proglacial fluvial, lacustrine and ice contact in origin, there is little evidence to support widespread aggradation over ice. It is acknowledged that some of the sediments that comprise the Crawcrook complex were probably aggraded over dead ice (though these have not been investigated by the author), but the evidence for widespread stagnation and ice melt out is not present in the sequences examined here (Figure 6.5).

Conceptual Model 2 (Figure 1.4) was built on two phases of development - the retreat of ice along the South Tyne Valley and the formation of 'kame' deposits, and the subsequent infilling by proglacial outwash and dissection to form the river terraces that characterise the South Tyne Valley. Because of the lack of accessible exposures in the 'kame' deposits, interpretation relied upon the limited IMAU borehole data. However, these suggested the sequence formed in an ice contact environment and in the light of the stagnation zone retreat mode of deglaciation, the sequence is accepted to have formed by active retreat of ice as the ice downwasted into the South Tyne Valley. The terraces (T4-T1), however, do not comprise extensive sedimentary sequences of proglacial outwash although there is some evidence of localised outwash sediments outcropping at the base of the sequence at Fourstones, South Tyne Valley, but clearly formed in a wandering gravel bed river (Figure 6.6). The valley bottom contains fluvial gravels directly overlying diamicton, although tentative clues in the borehole logs point to glaciofluvial depositional environments. This requires further investigation. There are no extensive outwash complexes like those in the mid and lower Tyne Valley. Rather, the lower South Tyne Valley is characterised by discrete mounds situated above the presently incised valley bottom, and discrete associations occur buried beneath the fluvial sequence. The lack of extensive outwash in the South Tyne Valley suggests both extensive erosion and removal, or, when the lower South Tyne Valley became ice-free (timing is unknown), the ice-front was stagnant and therefore supply limited so that extensive sandur development was not possible. Whilst the lower contact is glacigenic sediments (till, outwash), above which the river has deposited alluvium, it represents the extent of incision and uplift achieved by the river in the postglacial period. The chronology confirms that the upper river terraces (T4-T2) developed in the early Holocene and reflect the continuation of the paraglacial period into the Holocene.

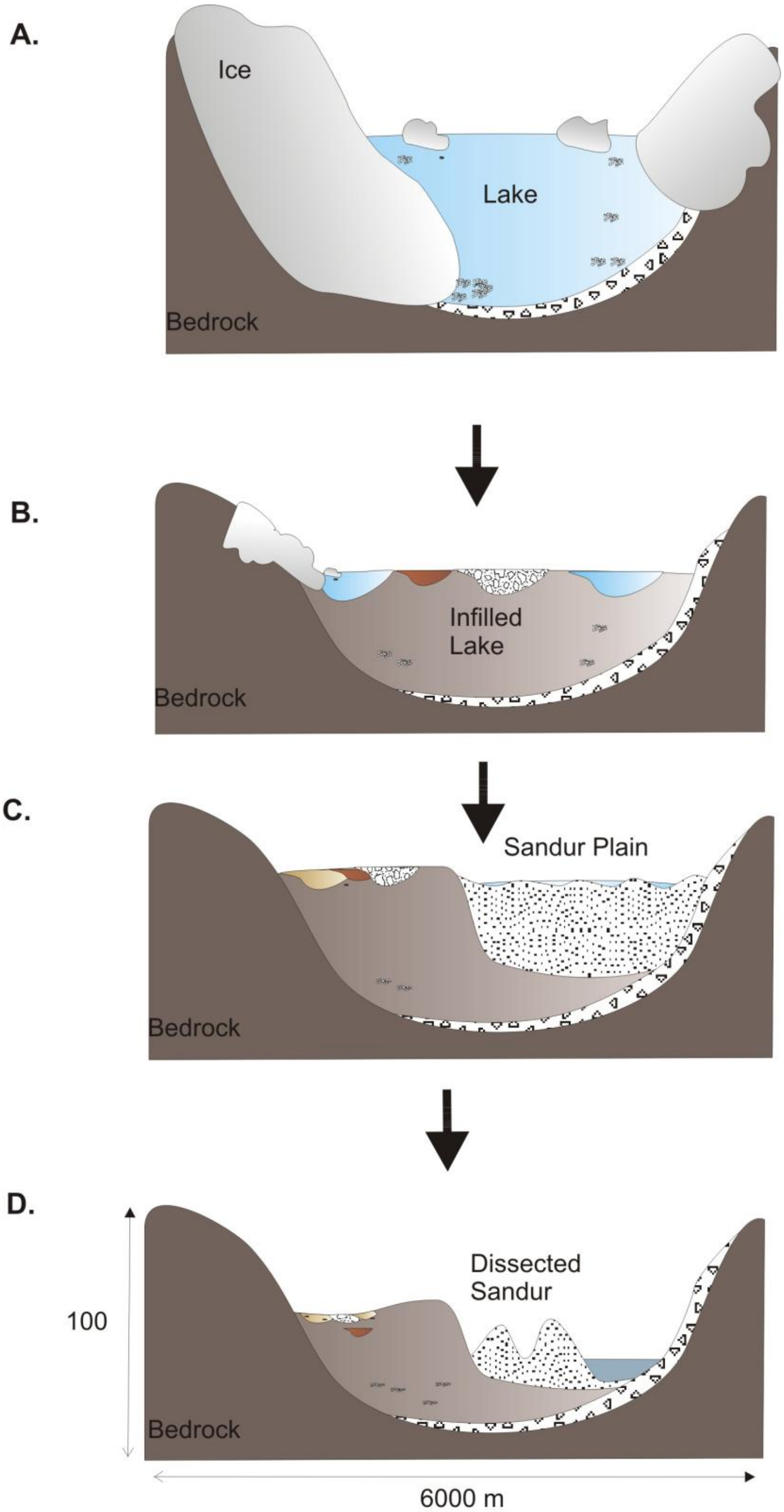


FIGURE 6.5: Cartoon depicting the development of the Crawcrook complex, lower Tyne Valley. A. During early deglaciation an ice proximal glacial lake forms, Glacial Lake Bradley. B. The lake gradual infilled. Later, small lake basins and channels form on the surface as proglacial streams flow from the stagnating ice in the vicinity of the complex. C. The lower Tyne is now substantially ice-free, with active ice located in the lower South Tyne Valley. Proglacial rivers flowing from the active ice front incised the earlier (lacustrine) deposits and a large sandur plain develops. D. During late deglaciation, dissection and erosion of sandur plain takes place as water flowing down the valley sides in small channels degrades the surface and leads to development of the mounded landforms, which form part of the present day landscape.

Based on the work carried out for this thesis, a new model of development can be proposed for the deglacial and paraglacial periods. Whilst the timing of events remains unclear, two phases of glaciation (i.e. ice advance and retreat in the Tyne Valley followed by ice advance and retreat down the coastal zone) best explains the complexity of the evidence. This idea remains to be explored. The evidence from this thesis suggests deglaciation in the Tyne Valley took place as: (1) active retreat (stagnation zone retreat mode) in the lower Tyne Valley resulting in the development of an ice free zone, but the continued presence of active ice in the South Tyne Valley continuing to deliver vast quantities of sediment charged meltwaters which built up as extensive proglacial sequences in the mid and lower Tyne Valley; (2) surging coastal ice (~19ka cal. BP) along the Northumberland/Durham coastline; and (3) persistence of active ice along the coastal zone resulted in the development of Glacial Lake Wear in the Tyne/Wear lowlands and other east draining rivers along the northeast coast. These lakes have been a major component of later deglaciation in the lowlands.

Incision and reworking of the glacigenic valley infill probably took place following deglaciation and continued into the Bølling-Allerød Interstadial and Younger Dryas Stadial, reflecting the initial 'paraglaciation' period. Floodplain development probably began in the Younger Dryas through reworking of readily available sediment. Two cycles of incision and aggradation took place during the early Holocene (~11-7ka cal. BP), resulting in the development of three major river terraces whose surfaces lie at ~20, 10 and 8m above present river level, and Holocene cut-and-fill terraces (here classified as a single unit, T1) which lie between 4 and 1m above present river level. Their development is linked to glacio-isostatic uplift and climate change (combined with readily available coarse sediments, unstable slopes, high discharge), which was driving incision across the region.

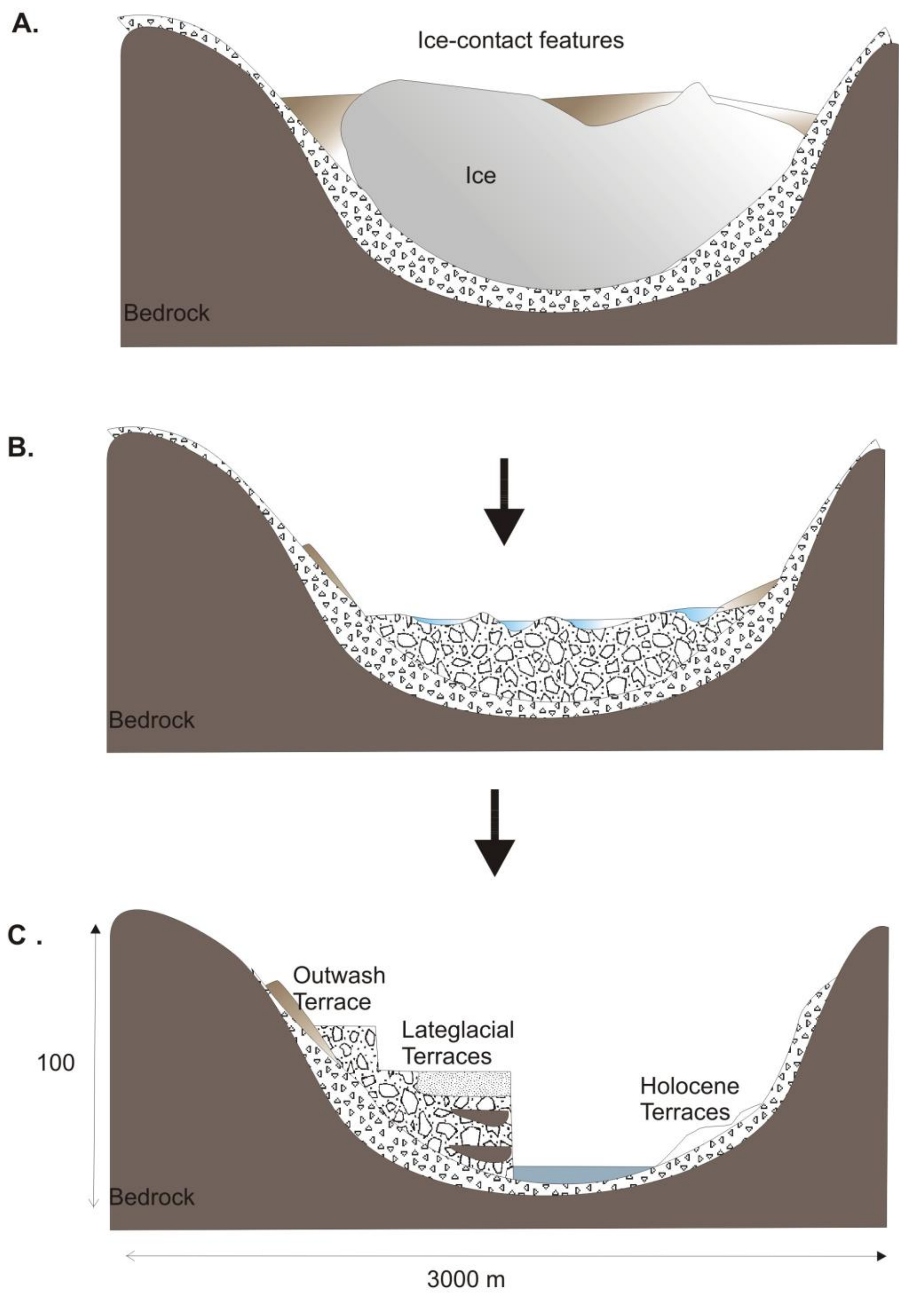


FIGURE 6.6: Cartoon depicting the development of the terrace sequence in the South Tyne Valley. A. During early deglaciation, the Tyne Valley ice recedes up the valley rapidly as the ice is captured by the ISB ice stream. Ice-contact features develop alongside the lateral ice margins in the South Tyne. B. Proglacial discharges from the stagnating ice results in the development of sandur plain and infilling of the valley floor. C. Continued paraglaciation results in reworking of glacigenic sediments. Glacio-isostatic uplift, and climate change leads to incision and downcutting through the glacigenic sediments. Substantial incision took place during the early Holocene leads to the formation of terraces, and through the re-establishment of a wandering gravel bed river, coarse bedload continues to build up the floodplain.

Chapter Seven

Conclusions

7.1 Thesis overview

This thesis has investigated a number of sites in very close detail, and the sedimentological analyses provide a much better understanding of their formational environment than morphological studies alone can do. When this data is combined with the growing body of work that has been gathered through the national scale morphological mapping and modelling (i.e. BRITICE) programs run by the universities of Sheffield and Durham, it is building a comprehensive picture of deglaciation in Britain. With increasing interest in morphological studies of the British-Irish ice sheet, the need for sediment-landform studies and stratigraphy is as critical as ever to ground remotely-sensed mapping exercises.

Through a combination of new investigation, evaluation and integration of existing datasets, a new model of response during deglaciation has been proposed. The sequence in the Tyne Valley records the stagnation signature of the Tyne Gap ice stream as it was diverted/captured/switched off during early deglaciation. Without the ice to the west, the Tyne Valley had no other source of supply and became starved of ice and deglaciation ensued. Currently the most likely candidate for this action was the ISB ice stream which drained southwest Scotland and the lakes (the main contributing area to the Tyne). The mechanism driving disintegration and stagnation of the ice in the Tyne Valley could be either millennial scale climatic change in the North Atlantic, though currently we only have H1 as a known event, or simply internal ice-sheet dynamics operating throughout a large scale deglaciation.

The landform-sediment assemblages confirm initial deglaciation was by active ice, followed by later *in-situ* stagnation, and this thesis considers that both climate change and internal ice sheet dynamics were probably the drivers of change resulting in the assemblages that outcrop along the Tyne Valley. Previous work by Clarke (1970), Francis (1975) and Lunn (1980) provided some insight into the development of the Tyne Valley sequence but it was clear that little was known about landform-sediment assemblages on the ground. Existing ideas about deglaciation were couched in the paradigms of the time, which favoured regional stagnation, downwasting and *in situ* stagnation driven by climate change (cf. Clark 1970; Lunn 1995; Mills and Holliday 1998). Whilst Clark's (1970) ideas concerning the persistence of coastal ice, development of a lower valley sediment trap, and glacial lakes all remain valid, this thesis has put forward a new mechanism for explaining the disappearance of the ice and development of the landform-sediment assemblages. The combination of both new (detailed sedimentology and stratigraphy, morphology) and existing data (i.e. sediment logs, landform maps) have helped disentangle existing ideas on deglaciation and explain the behaviour of the ice during deglaciation in the Tyne Valley.

This research contributes to the wider debate and interest in the response of the BIIS during deglaciation, further adding complexity to the story. It illustrates that we are far from understanding the behaviour of the ice, and certainly there is not a one model fits all solution. Current (on-going) work by Livingstone and others at Durham (in press) has suggested ice starvation for the Cumbria lowlands and for the Tyne gap zone, with the sediment-landform story developed through this thesis, I have the evidence to support their hypothesis. Critically, the sandar and ice disintegration evidence uncovered here can provide a testing ground for further work on these ideas as a phase of changing ice sheet dynamics is well-evidence. As well as this, with improvements in OSL dating protocols for glacigenic sediments (Thrasher, 2008, personal communication) the sequence offers the opportunity to secure a chronology for this time frame. The next phase for this research should be focused on dating the earliest ice wastage signature in the Tyne Valley and extending the research remit to provide a greater focus on similar decoupling zones in the eastwards draining Pennines.

7.2 Testing of hypotheses and conceptual models

Based on the available data at the start of the research, two conceptual models and three hypotheses were proposed to capture the complexity and development of the landform-sediment assemblages in the Tyne Valley. The evidence in this thesis has shown that existing models, and the mode of deglaciation is more complex than previously thought. A single conceptual model cannot represent accurately the development of the Tyne Valley sequence. It is proposed that an initial phase of deglaciation (stagnation) in the main Tyne Valley due to ice stream capture was followed by renewed glacial advance down the coast impounding meltwaters in the lower reaches and final deglaciation of the BIIS and the onset of the Holocene resulting in the formation of the terrace sequence. Based on the evidence presented here, this is the most plausible sequence of events that resulted in the landform-sediment assemblages present in the Tyne Valley.

7.2 Deglacial history

The landform-sediment assemblages investigated in the Tyne Valley indicate there is complexity and confusion in the evidence and there is not a single mode of deglaciation. The suite of geomorphic features relate to both active and stagnant ice operating in the valley over a prolonged period, and represents the stagnation pattern of the ice once deglaciation had begun. Two-phases of glacial activity might better explain the complexity and dichotomy of the landform-sediment assemblages in the region.

Initially, the main Tyne Valley became ice-free but active ice remained in the lower South Tyne Valley. Active ice retreat in the main Tyne Valley may have been driven by ice stream capture of the Tyne Gap ice stream, with ice flow being diverted into the Irish Sea Basin (as suggested by flow-sets in the Tyne Gap area indicating a east-west flow direction; Livingstone, personal communication, 2007) leaving the Tyne Valley starved of ice. This idea remains speculative, further work being required to establish the behaviour and timing of ice stream activity in the Tyne basin. Rapid retreat of lowland ice was by downwasting and stagnation, thus explaining the lack of end moraines. As the ice began to downwaste, small ice-marginal lakes developed. Glacial Lake Bradley (which forms part of the Crawcrook complex) represents the earliest observed sedimentation during deglaciation. The sedimentary evidence confirms the development of an extensive outwash plain in the main Tyne Valley, supplied by active ice located in the lower South Tyne Valley. There is no evidence to suggest the Crawcrook complex developed above, or within stagnant ice, as previous interpretations suggest. However, there is geomorphic evidence (e.g. eskers, ice-contact landforms, undulating topography, dead-ice features) for stagnation zone retreat and downwasting of ice in the southeastern area (in the Derwent Valley) of the Crawcrook complex.

In the lower South Tyne Valley, ice-contact features lie above the presently incised valley floor. Sequences could not be examined, but borehole logs provided tentative evidence for ice-proximal deposition. Further work is required to understand these landforms and their place in the deglaciation of the Tyne Valley.

Latterly, coastal ice impounded the lowlands of all the major east-draining rivers in northeast England. Meltwater drainage from coastal ice, active ice in the lower South Tyne Valley, and non-glacial runoff, led to the development of Glacial Lake Wear in the Tyne/Wear lowlands. The coastal ice-margin (probably) lay at the eastern extent of the Crawcrook complex, as evidenced by the basal till overriding the proglacial outwash.

A chronology for the Tyne Valley glaciagenic sequence could not be developed from the suite of sediments investigated. Thus, the deglacial chronology of the region remains to be disentangled. In the lower Tyne Valley, the problems encountered with OSL dating of glacigenic sediment, could relate to poor bleaching and poor quartz behaviour (i.e. low sensitivity). The lack of a fast component, due to both low dosing and sensitivity, or more likely due to the quartz from the bedrock lithology (Carboniferous Coal Measure sandstones) prevented any dates from being measured. Presently, the absence of a fast component in the quartz grains from this lithology provides barriers to determining the timing of stagnation in the Tyne Valley and establishing any synchronicity between regional climatic drivers, i.e. H1, the formation of ice-marginal and proglacial deposition and of regional deglaciation. However, future developments in OSL dating protocols and the application of cosmogenics to exposed bedrock surfaces may help resolve the problem. The bulk ^{14}C date from Dimlington, East Yorkshire (Catt and Penny, 1969), remains the only known date for deglacial activity in northern England.

7.3 Early postglacial development

In the Tyne Valley, there was ~10m of down-cutting from the Devensian glacigenic surface. The incision occurred prior to the start of the Holocene, and is demonstrable elsewhere in northern Britain (e.g. Kelvin Valley, central Scotland; Tipping *et al*. 2007). The timing of incision is unknown, but an OSL age from the overbank sediments of the oldest river terrace (T4) at Farnley (Table 5.1; X2734) indicates incision took place prior to the start of the Holocene.

There is no evidence to suggest fluvial development took place during the Bølling-Allerød Interstadial. It is probable that the period was characterised by down-cutting through the proglacial landform-sediment assemblages, and the associated reworking and export of glacially derived sediments. Preservation potential is low and fluvial deposits may have been removed during the Younger Dryas period (Lewin *et al*., 2005). During the Younger Dryas Stadial and early Holocene, valley floor incision and widening continued, resulting in the removal of glacigenic landform-sediment assemblages.

The initial paraglacial period continued into the early Holocene and activity in the Tyne Valley is reflected in the development of extensive proglacial sediments during deglaciation, the reworking of glacially derived material and the formation of river terraces. Within the lower and mid Tyne Valley, and the lower South Tyne Valley, four terrace surfaces were identified (T4-T1). The terrace surfaces lie at ~20(T4), 10(T3), 8(T2) and 4(T1) metres above present river level. Sedimentation within the terrace units is typically wandering gravel bed river style, characterised by gravel bars and channel floor deposits, overbank sediments and occasional flood sequences.

Valley floor refilling following the start of the Holocene was punctuated by two cycles of incision and aggradation driven by glacio-isostatic uplift, climatic change (sediment supply and discharge), readily available (glacial) sediments and continued paraglaciation. The high rates of incision and reworking culminated in the development of two major river terraces (T4 and T3) before ~8ka cal. BP. During the period between 11 and 7ka cal. BP the River Tyne responded to isostatic recovery and changes in climate (discharge and sediment supply) by incising through valley floor proglacial sediments and re-filling with glacially derived sediments. The volume of sediment reworking and loss was highest during the early Holocene, representing the most dynamic period of river valley development. Since the early Holocene, the modern valley floor has become confined within the upper terraces (T4-T2). During the mid/late Holocene the rate and amount of incision declined in response to reduced (but continued) isostatic uplift. The impact of Holocene climatic (and anthropogenic) change on the fluvial system was significant (cf. Macklin and Lewin, 2003) but less dramatic than the previous deglacial regime, and has continued to respond to paraglaciation.

River response during the Lateglacial period and prior to the Holocene remains poorly understood. The lack of sediment-landform assemblages means that evidence cannot be examined, thus whatever the river systems were doing, be that incising, eroding or removing glacigenic deposits, it cannot be ascertained. A lack of evidence is no proof at all for supporting any case.

The findings from the Tyne Valley suggest pre-Holocene incision of the glacigenic surfaces, and development of fluvial terraces between deglaciation and the early Holocene. Incision and erosion took place during the Younger Dryas rather that at the Younger Dryas/Holocene boundary or during the early Holocene. This suggests existing models are not applicable (Macklin and Lewin 2003). There is possible correlation with fluvial development in the Cheviots and central Scotland (Tipping 1998; Tipping *et al*. 2007), but more dates are required from the Tyne Valley terraces to investigate this further. A new regional complexity can be added into existing models. In the Tyne Valley, the early Holocene was a period of dynamic instability and provides support that (at least in the uplands of northern Britain) fluvial development was not characterised by quiescence or by incision, erosion and removal of earlier deposits. It is increasingly apparent that fluvial response across formerly glaciated catchments is complex and non-linear, and current models should be reconsidered.

7.4 Directions for future work

The idea of multi- phase glaciation during the Late Devensian remains to be explored. Future work should concentrate on re-investigating the wealth of glacigenic landform-sediment assemblages across Northumberland. New work is imperative to clarify their genesis and to investigate patterns and modes of deglaciation. This thesis has shown that re-investigation can provide new insights into deglaciation. An audit of the landform-sediment assemblages will generate an event stratigraphy that can help disentangle the respective periods of development, and components of glacial advance/retreat/deglaciation can be integrated to generate a fuller picture of ice-sheet response in northern England. With the recent country-wide remote mapping of glacigenic landforms (carried out at University of Sheffield) this is a timely opportunity to work on the sedimentary sequences that underlie the landforms, which will provide a clearer interpretation and understanding of these landforms/landscapes.

Investigating the sedimentary sequence along the coast, that was deposited by the late surge of an ice lobe down the northeast coast, would provide possible data on ice advance/retreat and offer correlations with the streamline bedforms within the Tweed Valley. Furthermore, the (basal) till exposed at Crawcrook (location 3) should be investigated to determine its origin and explore its relationship (if any) to the coastal ice lobe. This exposure could provide evidence of the inland limit of the coastal ice, or may represent a subsequent advance of local ice. Either interpretation would provide further evidence of complexity during the last deglacial period.

Within the Tyne Valley, and others in northern England, it is clear further work should be carried out to understand the full suite of sediments and landforms pertaining to paraglaciation. It is not simply a matter of looking for secondary sediment fluxes in the alluvial record to elucidate the paraglacial response.

The sites in the Tyne provide a record of glacial, deglacial and fluvial history for the last ~26ka. Potentially there is a rich record of palaeoenvironmental data in the early Holocene terraces. The presence of terraces themselves provides good evidence of glacioisostasy, and it would be wise to extent the investigation of terraces across the northeast region to include the Rivers Wear and Tees, and the smaller rivers of the Blyth, Coquet and Wansbeck, where BGS studies have hinted at the terrace sequences therein.

Finally, it is imperative that sequences are dated and event stratigraphies developed in order to place them in the wider context of climate change and responses across the BIIS. Whilst OSL dating has proved unsuccessful in this thesis, the technique is continuing to be refined (cf. Duller, 2006) and future developments may result in new opportunities to date quartz with low sensitivity from proglacial sequences. Cosmogenic dating is another emerging technique that may help constrain the timing of glacial and deglacial events in northern England.

References

Adediran, A. O., Parcharidis, I., Poscolieri, M. and Pavlopoulos, K. (2004) Computer-assisted discrimination of morphological units on north-central Crete (Greece) by applying multivariate statistics to local relief gradients. Geomorphology, **58**, 357-370.

Aitken, J. F. (1995) Lithofacies and depositional history of a Late Devensian ice-contact deltaic complex, northeast Scotland. Sedimentary Geology, **99**, 111-130.

Aitken, J. F. (1998) Sedimentology of Late Devensian glaciofluvial outwash in the Don Valley, Grampian region. Scottish Journal of Geology, **34**(2), 97-117.

Aitken, M. J. (1985) **Thermoluminescence dating**. London, Academic Press.

Alexanderson, H. and Murray, A. S. (2007) Was southern Sweden ice free at 19-25 ka, or were the post LGM glacifluvial sediments incompletely bleached? Quaternary Geochronology, **2**, 229-236.

Allen, C. T. (1995) Interferometric Synthetic Aperture Radar. IEEE Geoscience and Remote Sensing Society Newsletter, **96**, 6-13.

Allen, J. R. L. (1982) Late Pleistocene (Devensian) glaciofluvial outwash at Banc-y-Warren, near Cardigan (West Wales). Geological Journal, **17**, 31-47.

Allen, J. R. L. (1985) **Principles of physical sedimentology**. Allen & Unwin.

Allen, P. and Rose, J. (1986) A glacial meltwater drainage system between Whittonstall and Ebchester, Northumberland. In: Macklin, M. G. and Rose, J. (eds) **Quaternary River Landforms and Sediments in the Northern Pennines. Field Guide.** Oxford, British Geomorphological Research Group/Quaternary Research Association, pp.88.

Alley, R. B., Anandakrishnan, S., Bentley, C. R. and Lord, N. (1994) A water-piracy hypothesis for the stagnation of Ice Stream C, Antarctica. Annals of Glaciology, **20**, 187-194.

Alley, R. B. and Clark, P. U. (1999) The deglaciation of the Northern Hemisphere: A global perspective. Annual Review of Earth and Planetary Sciences, **27**, 149-182.

Anderson, E., Harrison, S. and Passmore, D. G. (1997) Geomorphological Mapping and Morphostratigraphic Relationships: reconstructing the glacial history of the Macgillycuddy's Reeks, south west Ireland. Society of Cartographers Bulletin, **30**(2), 9-20.

Anderson, W. (1940) Buried-Valleys and Lateglacial Drainage Systems in North West Durham. Proceedings of the Geologists' Association, **51**, 274-281.

Andrews, J. T. (1971) **Techniques of Till Fabric analysis**. Technical Bulletin 6. London, British Geomorphological Research Group.

Aniya, M. and Welch, R. (1981) Morphological analyses of glacial valleys and estimates of sediment thickness on the valley floor: Victoria Valley system, Antarctica. Antarctic Research, **71**, 76-95.

Anson, W. W. and Sharp, J. K. (1960) **Surface and Rockhead Relief Features in the Northern Part of the Northumberland Coalfield**. Department of Geography Research Series 2. University of Durham.

Armstrong, G. and Kell, J. (1951) The Tyne Tunnel. Geological Magazine, **88**(5), 357-361.

Ashley, G. M. (1975) Rhythmic sedimentation in glacial Lake Hitchcock, Massachusetts-Connecticut. In: Jopling, A. V. and McDonald, B. C. (eds) **Glaciofluvial and Glaciolacustrine Sedimentation**. Tulsa, Ok, Society of Economic Paleontologists and Mineralogists, pp.304-320.

Ashley, G. M. (1988) Classification of glaciolacustrine sediments. Genetic Classification of Glacigenic Deposits. AA Balkema, Rotterdam, **234**, 243-260.

Ashley, G. M. (1995) Glaciolacustrine environments. In: Menzies, J. (ed.) **Modern Glacial Environments: Processes, Dynamics and Sediments**. Oxford, Butterworth-Heinman, pp.417-427.

Ashley, G. M. (2002) Glaciolacustrine Environments. In: Menzies, J. (ed.) **Modern and Past Glacial Environments**. Oxford, Butterworth Heinman, pp.336-359.

Ashley, G. M., Southard, J. B. and Boothroyd, J. C. (1982) Deposition of climbing ripple beds: a flume simulation. Sedimentology, **29**, 67-79.

Ashley, G. M. and Warren, W. P. (1997) The Ice-Contact Environment. Quaternary Science Reviews, **16**, 629-634.

Aspinall, R. J., Macklin, M. G. and Brewis, T. (1986) Metal mining and floodplain sedimentation at Garrigill, and their influence on terrace and floodplain soil development. In: Macklin, M. G. and Rose, J. (eds) **Quaternary River Landforms and Sediments in the Northern Pennines, England. QRA/BGRG Field Guide**. Newcastle upon Tyne, QRA/BGRG, pp.35-45.

Bailey, R. M., Smith, B. W. and Rhodes, E. J. (1997) Partial bleaching and the decay form characteristics of quartz OSL. Radiation Measurements, **27**, 123-136.

Baker, V. R. (1978) Paleohydraulics and hydrodynamics of scabland floods. In: Baker, V. R. and Nummedal, D. (eds) **The Channeled Scabland, a guide to the geomorphology of the Columbia Basin, Washington**. Washington, D. C., NASA, pp.59-79.

Baker, V. R. (2007) Introduction: Regional Landform Analysis. **Geomorphology** [Internet]. Available from: <http://disc.gsfc.nasa.gov/geomorphology/GEO_1/GEO_CHAPTER_1.shtml> [Accessed 25 July 2007].

Ballantyne, C. K. (1991) Late Holocene Erosion in Upland Britain: Climatic Deterrioration or Human Influence? The Holocene, **1**(1), 81-85.

Ballantyne, C. K. (2002a) A general model of paraglacial landscape response. The Holocene, **12**(3), 371-376.

Ballantyne, C. K. (2002b) Paraglacial geomorphology. Quaternary Science Reviews, **21**, 1935-2017.

Ballantyne, C. K. (2003) Paraglacial Landsytems. In: Evans, D. J. A. (ed) **Glacial Landsystems**. London, Hodder Arnold, pp.432-461.

Ballantyne, C. K. and Whittington, G. (1999) Late Holocene floodplain incision and alluvial fan formation in the central Grampian highlands Scotland: chronology environment and implications. Journal of Quaternary Science, **14**(7), 651-671.

Banerjee, I. and McDonald, B. C. (1975) Nature of Esker Sedimentation. In: Jopling, A. V. and McDonald, B. C. (eds) **Glaciofluvial and Glaciolacustrine Sedimentation**. Tulsa, OK, Society of Economic Paleontologists and Mineralogists, pp.132-154.

Barber, D. C., Dyke, A., Hillaire-Marcel, C., Jennings, A. E., Andrews, J. T., Kerwin, M. W., Bilodeau, G., McNeely, R., Southon, J., Morehea, M. D. and Gagnon, J.-M. (1999) Forcing of the cold event of 8200 years ago by catastrophic drainage of Laurentide lakes. Nature, **400**, 344–348.

Bard, E., Hamelin, B., Arnold, M., Montaggioni, L., Cabioch, G., Faure, G. and Rougerie, F. (1996) Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. Nature, **382**, 241-244.

Bartley, D. D. (1966) Pollen analysis of some lake deposits near Bamburgh in Northumberland. New Phytologist, **65**(2), 141-156.

Beaumont, P. (1968) **A history of glacial research in northern England**. Department of Geography Occasional Paper 9. University of Durham, pp.21.

Beaumont, P. (1971) Stone orientation and stone count data from the lower till sheet, eastern Durham. Proceedings of the Yorkshire Geological Society, **3**(16), 343-360.

Beckett, S. C. (1977) **The Bog, Roos**. X INQUA Congress Excursion Guide: Yorkshire and Lincolnshire. Norwich, Geo Abstracts. 42-46pp.

Bell, M. (1982) The effects of land-use and climate on valley sedimentation. In: Harding, A. F. (ed.) **Climate Change in later Prehistory**. Edinburgh, Edinburgh University Press.

Bell, M. and Walker, M. J. C. (1992) **Late Quaternary Enironmental Change: Physical and Human Perspectives**. Essex, Longman.

Benn, D. I. (1996) Subglacial and subaqueous processes near a glacier grounding line: sedimentological evidence from a former ice-dammed lake, Achnasheen, Scotland. Boreas, **25**, 23-36.

Benn, D. I. and Evans, D. J. A. (1998) **Glaciers and Glaciation**. London, Arnold. 734pp.

Bennett, M. R. and Boulton, G. S. (1993) Deglaciation of the Younger Dryas or Loch Lomond Stadial ice-field in the northern Highlands, Scotland. Journal of Quaternary Science, **8**, 133-145.

Bennett, M. R. and Glasser, N. F. (1996) **Glacial Geology: Ice Sheets and Landforms**. Chichester, Wiley.

Birks, H. J. B. (1989) Holocene isochrone maps and patterns of tree-spreading in the British Isles. Journal of Biogeography, **18**, 103-115.

Bitinas, A., Karmaziene, D. and Jusiene, A. (2004) Glaciolacustrine kame terraces as an indicator of conditions of deglaciation in Lithuania during the Last Glaciation. Sedimentary Geology, **165**(3-4), 285-294.http://www.sciencedirect.com/science/article/B6V6X-4BJX1VM-2/2/3135be07f8fc277a6ea562b558c9f56b.

Blacknell, C. (1981) Sandy gravel accumulation in a fluvial environment. Geological Journal, **16**, 287-97.

Blacknell, C. (1982) Morphology and surface sedimentary features of point bars in Welsh gravel-bed rivers. Geological Magazine, **119**(2), 181-192.

Blott, S. J. and K., P. (2001) GRADISTAT: a grain size distribution and statistics package for the analysis of unconsolidated sediments. Earth Surface Processes and Landforms, **26**, 1237-1248.

Bluck, B. J. (1974) Structure and directional properties of some valley sandur deposits in southern Iceland. Sedimentology, **21**, 533-554.

Bluck, B. J. (1979) Structure of coarse grained braided stream alluvium. Transactions of the Royal Society of Edinburgh, **70**, 181-221.

Blum, M. D. and Tornqvist, T. E. (2000) Fluvial responses to climate and sea-level change: a review and look forward. Sedimentology, **47**(s 1), 2-48.

Boe, A., Murray, A. S. and Dahl, S. O. (2007) Resetting of sediments mobilised by the LGM ice-sheet in southern Norway. Quaternary Geochronology, **2**, 222-228.

Bolongaro-Crevenna, A., Torres-Rodriguez, V., Sorani, V., Frame, D. and Ortiz, M. A. (2005) Geomorphometric analysis for characterising landforms in Morelos State, Mexico. Geomorphology, **67**, 407-422.

Bond, G., Broecker, W., Johnsen, S., McManus, J., Labeyrie, L., Jouzel, J. and Bonani, G. (1993) Correlations between climate records from North Atlantic sediments and Greenland ice. Nature, **365**, 143-147.

Bond, G. and Lotti, R. (1995) Iceberg discharges into the North Atlantic on millennial time scales during the last deglaciation. Science, (267), 1005-1010.

Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., deMenocal, P., Priore, P., Cullen, H., Hajdas, I. and Bonani, G. (1997) A Pervasive Millennial-Scale Cycle in North Atlantic Holocene and Glacial Climates. Science, **278**, 1257-1266.

Boothroyd, J. C. (1972) **Coarse-grained sedimentation on a braided outwash fan, northeast Gulf of Alaska**. Coastal Research Division Technical Report. Columbia, University of South Carolina, pp.127.

Boothroyd, J. C. and Ashley, G. M. (1975) Processes, bar morphology, and sedimentary structures on braided outwash fans, northeastern gulf of Alaska. In: Jopling, A. V. and McDonald, B. C. (eds) **Glaciofluvial and Glaciolacustrine Sedimentation**. Tulsa, Oklahoma, Society of Economic Paleontologists and MineralogistsSpecial Publication 23, pp.193-222.

Boothroyd, J. C. and Nummedal, D. (1978) Proglacial braided outwash: A model for humid alluvial-fan deposits. In: Miall, A. D. (ed.) **Fluvial Sedimentology**. Canadian Society for Petroleum Geologists, pp.641-688.

Botter-Jensen, L., Bulur, E., Murray, A. S. and Poolton, N. R. J. (2002) Enhancements in luminescence measurement techniques. Radiation Protection Dosimetry, **101**, 119-124.

Botter-Jensen, L., McKeever, S. W. S. and Wintle, A. G. (2003) **Optically Stimulated Luminescence Dosimetry**. Amsterdam, Elsevier Science.

Boulton, G. and Hagdorn, M. (2006) Glaciology of the British Isles Ice Sheet during the last glacial cycle: form, flow, streams and lobes. Quaternary Science Reviews, **25**(23-24), 3359-3390.

Boulton, G. S. and Hagdorn, M. (2007) **Glaciology of the British Isles Ice Sheet during the last glacial cycle: form, flow, streams and lobes: abstract**. **QRA Annual Discussion Meeting: The Growth, Maximum Extent and Deglaciation of the Last British Ice Sheet**, **4-5 January 2007**, St Andrews.

Boulton, G. S., Jones, A. S., Clayton, K. M. and Kenning, M. J. (1977) A British ice-sheet model and patterns of glacial erosion and deposition in Britain. In: Shotton, F. W. (ed.) **British Quaternary Studies, recent advances**. Oxford, Clarendon Press, pp.231-246.

Boulton, G. S., Smith, G. D., Jones, A. S. and Newsome, J. (1985) Glacial geology and glaciology of the last mid-latitude ice sheets. Journal of the Geological Society, **142**, 447-474.

Bowen, D. Q. (1994) The Pleistocene history of North West Europe. Science Progress, **14**, 107-112.

Bowen, D. Q., Phillips, F. M., McCabe, A. M., Knutz, P. C. and Sykes, G. A. (2002) New data for the Last Glacial Maximum in Great Britain and Ireland. Quaternary Science Reviews, **21**, 89-101.

Bowen, D. Q., Sykes, G. A., Maddy, D., Bridgland, D. R. and Lewis, S. G. (1995) Aminostratigraphy and amino acid geochronology of English lowland valleys: the Lower Thames in context. In: Bridgland, D. R., Allen, P. and Haggert, B. A. (eds) **The Quaternary of the lower reaches of the Thames, Field Guide**. Durham, Quaternary Research Association, pp.61-63.

Brazier, V., Kirkbride, M. P. and Gordon, J. E. (1998) Active ice-sheet deglaciation and ice-dammed lakes in the northern Cairngorm Mountains, Scotland. Boreas, **27**, 297-310.

Brennand, T. A. (1994) Macroforms, large bedforms and rythmic sedimentary sequences in subglacial eskers, south-central Ontario: implications for esker genesis and meltwater regime. Sedimentary Geology, **91**, 9-55.

Bridgland, D. R. (1983) The Quaternary fluvial deposits of north Kent and eastern Essex. City of London Polytechnic, London.

Bridgland, D. R. (1986) **Clast Lithological Analysis, Technical Guide 3**. London, Quaternary Research Association.

Bridgland, D. R. (1994) **Quaternary of the Thames**. London, Chapman & Hall. 441pp.

Bridgland, D. R., Maddy, D. and Bates, M. (2004) River terraces sequences: templates for Quaternary geochronology and marine-terrestrial correlation. Journal of Quaternary Science, **19**(2), 203-218.

Bridgland, D. R. and Schreve, D. C. (2002) River terrace formation in synchrony with long-term climatic fluctuations: examples from southern Britain. In: Maddy, D., Macklin, M. G. and Woodward, J. C. (eds) **River Basin Sediment Systems: Archives of Environmental Change**. Rotterdam, Balkema.

Bridgland, D. R. and Schreve, D. C. (2004) Quaternary lithostratigraphy and mammalian biostratigraphy of the lower Thames terrace system, South-east England. Quaternaire, **15**(1-2), 29-40.

Brown, A. G. (1991) Hydrogeomorphology and palaeoecology of the Severn basin during the last 18,000 years: orders of change in a maritime catchment. In: Gregory, K. J., Starkel, L. and Thornes, J. B. (eds) **Fluvial Processes in the Temperate Zone During the Last 15,000 Years**. Chichester, Wiley, pp.147–169.

Brown, A. G. (1997) **Alluvial Geoarchaeology Floodplain Archaeology and environmental Change**. Cambridge Manuals in Archaeology. Cambridge, Cambridge University Press.

Brown, D. G., Lusch, D. P. and Duda, K. A. (1998) Supervised classification of types of glaciated landscapes using digital elevation data. Geomorphology, **21**, 233-250.

Brown, I. M. (1992) Deglaciation of the Dee Valley, NE Scotland. Unpublished PhD, University of Aberdeen.

Brown, I. M. (1993) The pattern of deglaciation of the last (Late Devensian) Scottish ice-sheet: evidence from the Dee Valley, NE Scotland. Journal of Quaternary Science, **8**, 235-50.

Brown, I. M. (1994) Former glacial lakes in the Dee Valley: origin, drainage and significance. Scottish Journal of Geology, **30**(2), 147-158.

Brunsden, D. (1993) Barriers to Geomorphological change. In: Thomas, D. S. G. and Allison, R. J. (eds) **Landscape Sensitivity**. Chichester, Wiley.

Brunsden, D. (2001) A critical assessment of the sensitivity concept in geomorphology. Catena, **42**(2-4), 99-123.

Brunsden, D. and Thornes, J. B. (1979) Landscape sensitivity and change. Transactions of the Institute of British Geographers, **4**, 463-484.

Bryant, D. (1983) Facies sequences associated with some braided river deposits of Late Pleistocene age from southern Britain. In: Collinson, J. D. and Lewin, J. (eds) **Modern and Ancient Fluvial Systems, Special Publication of the International Association of Sedimentologists**. pp.267-275.

Bryant, I. D. (1991) Sedimentology of glaciofluvial deposits. In: Ehlers, J., Gibbard, P. L. and Rose, J. (eds) **Glacial Deposits in Great Britain and Ireland**. Rotterdam, Balkema.

Burns, D. (1875) **Annual Report of the Union of Mechanics Institute of Northumberland**. Newcastle upon Tyne.

Burrough, P. A. (1986) **Principles of geographical information systems for land resources assessment**. Clarendon Press.

Burrough, P. A. and McDonnell, R. A. (1998) **Principals of Geographical Information Systems**. Oxford, Oxford University Press. 365pp.

Candy, I., Black, S. and Sellwood, B. W. (2005) U-series isochron dating of immature and mature calcretes as a basis for constructing Quaternary landform chronologies for the Sorbas basin, southeast Spain. Quaternary Research, **64**(1), 100-111.

Cant, D. J. (1978) Development of a facies model for sandy braided river sedimentation: comparison of the South Saskatchewan River and the Battery Point Formation. In: Miall, A. D. (ed.) **Fluvial Sedimentology**. Canadian Society of Petroleum Geologists, pp.627-639.

Cant, D. J. and Walker, R. G. (1976) Development of a facies model for sandy braided-fluvial facies for the Devonian Battery Point Sandstone, Quebec. Canadian Journal of Earth Sciences, **13**, 102-119.

Carling, P. (1996) Morphology, sedimentology and palaeohydraulic significance of large gravel dunes, Altai Mts, Siberia. Sedimentology, **43**, 647-664.

Carruthers, R. G., Dinham, C. H., Burnett, G. A. and Maden, J. (1927) **The geology of Belford, Holy Island and the Fame Islands**. Memoirs of the British Geological Survey of Great Britain Sheet 4. London, HMSO.

Carter, R. G. (1992) Sea-level changes: past, present and future. Quaternary Proceedings Journal of Quaternary Science, **2**, 111-132.

Catt, J. A. (1991a) The Quaternary history and glacial deposits of East Yorkshire. In: Ehlers, J., Gibbard, P. L. and Rose, J. (eds) **Glacial deposits in Great Britain and Ireland**. Rotterdam, Balkema, pp.185-191.

Catt, J. A. (1991b) Late Devensian glacial deposits and glaciations in eastern England and the adjoining offshore region. In: Ehlers, J., Gibbard, P. L. and Rose, J. (eds) **Glacial deposits of Great Britain and Ireland**. Rotterdam, Balkeme, pp.61-68.

Catt, J. A. and Penny, L. F. (1966) The Pleistocene deposits of Holderness, East Yorkshire. Proceedings of the Yorkshire Geological Society, **35**, 375–420.

Challis, K. and Howard, A. J. (2003) GIS-based modelling of sub-surface deposits for archaeological prospection in alluvial landscapes. In: Howard, A. J., Macklin, M. G. and Passmore, D. G. (eds) **Alluvial Archaeology of Europe**. Rotterdam, Balkema, pp.263-275.

Chappell, A. (1995) The spatial variability of net soil flux in the south-west Niger, West Africa. Unpublished PhD thesis, Department of Geography, University College London.

Chappell, A., Heritage, G. L., Fuller, I. C., Large, A. R. G. and Milan, D. J. (2003) Geostatistical analysis of ground-survey elevation data to elucidate spatial and temporal river channel change. Earth Surface Processes and Landforms, **28**, 349-370.

Cheel, R. J. and Rust, B. R. (1986) A sequence of soft-sediment deformation (dewatering) structures in Late Quaternary subaqueous outwash near Ottawa, Canada. Sedimentary Geology, **47**, 77-93.

Church, M. (1972) **Baffin Island sandurs: a study of Arctic fluvial processes**. Bulletin of the Geological Survey of Canada. pp.208.

Church, M. (1983) Pattern of instability in a wandering gravel bed channel. In: Collinson, J. D. and Lewin, J. (eds) **Modern and Ancient Fluvial Systems, Special Publication of the International Association of Sedimentologists No. 6**. Oxford, Blackwell, pp.169-180.

Church, M. and Gilbert, R. (1975) Proglacial fluvial and lacustrine environments. In: Jopling, A. V. and McDonald, B. C. (eds) **Glaciofluvial and Glaciolacustrine Sedimentation**. Society of Economic Paleontologists and Mineralogists Special Publication, pp.22-100.

Church, M., Moore, D. and Rood, K. (1981) **Catalogue of Alluvial River Channel Regime Data**. Department of Geography, University of British Columbia.

Church, M. and Ryder, J. M. (1972) Paraglacial sedimentation: a consideration of fluvial processes conditioned by glaciation. Geological Society of America Bulletin, **83**, 3059-3072.

Church, M. and Slaymaker, O. (1989) Disequilibrium of Holocene sediment yield in glaciated British Columbia. Nature, **337**, 452-454.

Clapperton, C. M. (1967) The Deglaciation of the East Cheviot Area, Northumberland. Unpublished PhD, University of Edinburgh, Edinburgh.

Clapperton, C. M. (1970) The evidence for a Cheviot ice cap. Transactions of the Institute of British Geographers, **50**, 115-126.

Clapperton, C. M. (1971) The pattern of deglaciation in part of north Northumberland. Transactions of the Institute of British Geographers, **53**, 67-78.

Clark, C. D. (1997) Reconstructing the evolutionary dynamics of palaeo-ice sheets using multi-temporal evidence, remote sensing and GIS. Quaternary Science Reviews, **16**, 1067-1092.

Clark, C. D., Evans, D. J. A., Khatwa, A., Bradwell, T., Jordan, C. J., Marsh, S. H., Mitchell, W. A. and Bateman, M. D. (2004a) Map and GIS database of glacial landforms and features related to the last British Ice Sheet. Boreas, **33**, 359-375.

Clark, C. D., Evans, D. J. A. and Piotrowski, J. A. (2003) Palaeo-ice streams: an introduction. Boreas, **32**, 261-262.

Clark, C. D., Gibbard, P. L. and Rose, J. (2004b) Pleistocene glacial limits in England, Scotland and Wales. In: Ehlers, J. and Gibbard, P. L. (eds) **Quaternary Glaciations - Extent and Chronology: Part 1 Europe**. Amsterdam, Elsevier B. V., pp.488.

Clark, C. D. and Stokes, C. R. (2001) Extent and basal characteristics of the M'Clintock Channel Ice Stream. Quaternary International, **86**(1), 81-101.

Clark, R. (1970) Aspects of glaciation in Northumberland. Proceedings of theCumberland Geological Society, **6**, 277-280.

Clark, R. (1971) Periglacial landforms and landscapes in Northumberland. Proceedings of the Cumberland Geological Society, **3**, 5-20.

Clarke, B. G, Hughes, D. B. And Hashemi, S. (2008) Physical characteristics of subglacial tills. Géotechnique, **58**(1), 67-76.

Clarke, M. L. (1996) IRSL dating of sands: bleaching characteristics at deposition inferred from the use of single aliquots. Radiation Measurements, **26**, 611-620.

Clarke, M. L., Rendell, H. M. and Wintle, A. G. (1999) Quality Assurance in luminescence dating. Geomorphology, **29**, 173-185.

Clemmensen, L. B. and Houmark-Nielsen, M. (1981) Sedimentary features of a Weichselian glaciolacustrine delta. Boreas, **10**, 229-245.

Collins, P. E. F., Fenwick, I. M., Keith-Lucas, D. M. and Worsley, P. (1996) Late Devensian river and floodplain dynamics and related environmental change in northwest Europe, with particular reference to a site at Woolhampton, Berkshire, England. Journal of Quaternary Science, **11**(5), 357-375.

Colls, A. E., Stokes, S., Blum, M. D. and Straffin, E. (2001) Age limits on the Late Quaternary evolution of the upper Loire river. Quaternary Science Reviews, **20**, 743-750.

Common, R. (1954) The Geomorphology of the east Cheviot area. Scottish Geographical Magazine, **70**, 124-138.

Cooke, R. U. and Doornkamp, J. C. (1990) **Geomorphology in environmental management: a new introduction**. Clarendon Press.

Coope, G. R. (1977) Fossil coleopteran assemblages as sensitive indicators of climatic changes during the Devensian (Last) cold stage. Philosophical Transactions of the Royal Society of London, **B280**, 313-340.

Coope, G. R. (1986) Coleoptera analysis. In: Berglund, B. E. (ed.) **Handbook of Holocene Palaeoecology and Palaeohydrology**. Chichester, John Wiley & Sons, pp.703–713.

Coope, G. R. and Brophy, J. A. (1972) Lateglacial environmental changes indicated by a coleopteran successions from North Wales. Boreas, **1**, 97-142.

Coope, G. R., Lemdahl, G., Lowe, J. J. and Walkling, A. (1998) Temperature gradients in northern Europe during the last glacial–Holocene transition (14–9 14 C kyr BP) interpreted from coleopteran assemblages. Journal of Quaternary Science, **13**(5), 419-433.

Costa, J. E. (1983) Paleohydraulic reconstructions of flash-flood peaks from boulder deposits in the Colorado Front Ranges. Geological Society of America Bulletin, **94**, 986-1004.

Costello, W. R. and Walker, R. G. (1972) Pleistocene sedimentology, Credit River, Southern Ontario: A new component of the braided river model. Journal of Sedimentary Petrology, **42**, 389-400.

Cotton, J. A., Heritage, G. L., Large, A. R. G. and Passmore, D. G. (1999) Biotic response to Late Holocene floodplain evolution in the River Irthing catchment Cumbria. In: Marriot, S. B. and Alexander, J. ed. **Floodplains: Interdisciplinary Approaches**. London, Geological Society, pp.63-178.

Cox, F. C. (1983) **Geological notes and local details for 1:10 000 sheets NZ25NW, NE, SW and SE, Kibblesworth, Birtley, Craghead and Chester-le-Street, part of 1:50 000 sheet 20 (Newcastle upon Tyne) and sheet 21 (Sunderland)**. Institute of Geological Sciences, Geological Notes and Local Details for 1:10 000 Sheets. Newcastle upon Tyne, HMSO.

Cullingford, R. A. and Smith, D. E. (1980) Late Devensian raised shorelines in Angus and Kincardineshire. Boreas, **9**, 21-38.

Cullingford, R. A., Smith, D. E. and Firth, C. R. (1991) The altitude and age of the Main Postglacial Shoreline in eastern Scotland. Quaternary International, **9**, 39-52.

Cumming, J. S. (1971) **Rockhead relief of southeast Northumberland and the lower Tyne Valley**. Department of Geography Seminar Papers No. 18. University of Newcastle upon Tyne, pp.12.

Cumming, J. S. (1977) A descriptive account of the buried rockhead topography of Tyneside. In: Conference Transactions. **Institute of British Geographers Annual Conference**, **4-7th January 1977, Newcastle upon Tyne**, pp.57-65.

Currier, L. W. (1941) Disappearance of the last ice sheet in Massachusetts by Stagnation Zone Retreat [abs]. Geological Society of America Bulletin, **52**, 1895.

Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahljensen, D., Gundestrup, N. S., Hammer, C. U., Hvidberg, C. S., Steffensen, J. P., Sveinbjornsdottir, A. E., Jouzel, J. and Bond, G. (1993) Evidence for general instability of past climate from a 250-kyr ice-core record. Nature, **364**, 218-220.

Dawson, A. G. (1989) Distribution and development of the Main Rock Platform: reply. Scottish Journal of Geology, **25** (2), 233-238.

Dawson, M. R. (1985) Environmental reconstruction of a late Devensian terrace sequence: some preliminary findings. Earth Surface Processes and Landforms, **10**, 237-246.

Dawson, M. R. and Bryant, I. D. (1987) Three-Dimensional Facies Geometry in Pleistocene Outwash Sediments in Worcestershire, UK. In: Etheridge, F. G., Flores, R. M. and Harvey, A. D. (eds) **Recent Developments in Fluvial Sedimentology**. Tulsa, Oklahoma, The Society of Economic Paleontologists and Mineralogists, pp.191-196.

Dawson, M. R. and Gardiner, V. (1987) River Terraces: The General Models and a Palaeoenvironmental and Sedimentological Interpretation of the Terraces of the Lower Severn. In: Gregory, K. J., Lewin, J. and Thornes, J. B. (eds) **Palaeohydrology in Practice**. Chichester, Wiley & Sons, pp.269-305.

Delaney, C. (2002) Sedimentology of a glaciofluvial landsystem, Lough Ree area, Central Ireland: implications for ice margin characteristics during Devensian deglaciation. Sedimentary Geology, **149**, 111-126.

Demoulin, A., Bovy, B., Rixhon, G. and Cornet, Y. (2007) An automated method to extract fluvial terraces from digital elevation models: The Vesdre valley, a case study in eastern Belgium. Geomorphology, **91(**1-2), 51-64.

Desloges, J. R. and Church, M. (1987) Channel and floodplain facies in a wandering gravel-bed river. In: Ethridge, F. G., Flores, R. M. and Harvey, M. D. (eds) **Recent Developments in Fluvial Sedimentology**. Tulsa, OK, Society of Economic Paleontologists and Mineralogists, Special Publication 39, pp.99-109.

Donnelly, R. and Harris, C. (1989) Sedimentology and origin of deposits from a small ice-dammed Lake, Leirbreen, Norway. Sedimentology, **36**, 581-600.

Douglas, T. D. (1991) Glacial deposits of Northumbria. In: Ehlers, J., Gibbard, P. L. and Rose, J. (eds) **Glacial Deposits in Great Britain and Ireland**. Rotterdam, Balkema, pp.169-174.

Douglas, T. D. and Harrison, S. (1985) Periglacial landforms and sediments in the Cheviots. In: Boardman, J. (ed.) **Field Guide to the Periglacial Landforms of Northern England**. Cambridge, Quaternary Research Association, pp.68-75.

Douglas, T. D. and Harrison, S. (1987) Late Devensian Periglacial slope deposits in the Cheviot Hills. In: Boardman, J. (ed.) **Periglacial Processes and Landforms in Britain and Ireland**. Cambridge, Cambridge University Press, pp.237-244.

Dowman, I. (2004) Integration of LiDAR and Ifsar for mapping. In: Altan, O. (ed.) **Proceedings of the XXth ISPRS Congress Geo-Imagery Bridging Continents**, **12-23 July 2004, Istanbul, Turkey**, pp.90-100.

Dowman, I., Balan, P., Renner, K. and Fischer, P. (2003) **An Evaluation of Nextmap Terrain Data in the Context of UK National Datasets for Getmapping**. University College London, Department of Geomatic Engineering, pp.18.

Duller, G. A. T. (1991) Equivalent dose determination using single aliquots. Nuclear Tracks and Radiation Measurements, **18**, 371-378.

Duller, G. A. T. (1994a) Luminescence dating of poorly bleached sediments from Scotland. Quaternary Geochronology, **13**, 521-524.

Duller, G. A. T. (1994b) Luminescence dating of sediments using single aliquots-new procedures. Quaternary Science Reviews, **13**, 149-156.

Duller, G. A. T. (1995) Luminescence dating using single aliquots: methods and applications. Radiation Measurements, **24**, 217-226.

Duller, G. A. T. (2004) Luminescence dating of Quaternary sediments: recent advances. Journal of Quaternary Science, **19**(2), 183-192.

Duller, G. A. T. (2006) Single grain optical dating of glacigenic deposits. Quaternary Geochronology, **1**, 296-304.

Duller, G. A. T., Bøtter-Jensen, L., Kohsiek, P. and Murray, A. S. (1999) A High-Sensitivity Optically Stimulated Luminescence Scanning System for Measurement of Single Sand-Sized Grains. Radiation Protection Dosimetry, **84**(1-4), 325-330.

Dwerryhouse, A. R. (1902) The Glaciation of Teesdale, Weardale and the Tyne Valley, and their Tributary Valleys. Quaterley Journal of the Geological Society of London, **58**, 572-608.

Environmental Protection Agency U.S. (2004) Developing Spatially Interpolated Surfaces and Estimating Uncertainty. **Environmental Protection Agency** [Internet]. Available from: <http://epa.gov/air/oaqps/pm25/docs/dsisurfaces.pdf> [Accessed 16 October 2006].

ESRI. (2004) **Using ArcGIS Geostatistical Analyst: ArcGIS 9**. ESRI Press. 300pp.

Etzelmuller, B. (2000) Quantification of thermo-erosion in proglacial areas - examples from Svalbard. Zeitschrift fur Geomorpholgie, **Supplementband 44**, 343-361.

Evans, D. J. A., Clark, C. D. and Mitchell, W. A. (2005) The last British Ice Sheet: A review of the evidence utilised in the compliation of the Glacial Map of Britain. Earth Science Reviews, **70**, 253-312.

Evans, D. J. A. and Cofaigh, C. (2003) Depositional evidence for marginal oscillations of the Irish Sea ice stream in southeast Ireland during the last glaciation. Boreas, **32**(1), 76-101.

Evans, D. J. A., Owen, L. A. and Roberts, D. (1995) Stratigraphy and sedimentology of Devensian (Dimlington Stadial) glacial depsits, east Yorkshire, England. Journal of Quaternary Science, **10**(3), 241-266.

Evans, I. S. (1999) Castle Eden Dene and Blunts Dene. In: Bridgland, D. R., Horton, B. P. and Innes, J. B. (eds) **The Quaternary of North-East England: Field Guide**. Quaternary Research Association, pp.57-64.

Everest, J., Bradwell, T. and Golledge, N. (2005) Subglacial Landforms of the Tweed Palaeo-Ice Stream. Scottish Geographical Journal, **121**(2), 163-173.

Everest, J. D. and Kubik, P. W. (2006) The deglaciation of eastern Scotland: cosmogenic 10 Be evidence for a Lateglacial stillstand. Journal of Quaternary Science, **21**, 95–104.

Eyles, N. (1987) Late Pleistocene debris-flow deposits in large glacial lakes in British Columbia and Alaska. Sedimentary Geology, **53**, 33–71.

Eyles, N., Clark, B. M. and Clague, J. J. (1987) Coarse-grained sediment gravity flow facies in a large supraglacial lake. Sedimentology, **34**(2), 193-216.

Eynon, G. and Walker, R. G. (1974) Facies relationships in Pleistocene outwash gravels southern Ontario: a model for bar growth in braided rivers. Sedimentology, **21**, 43-70.

Fairbanks, R. G. (1989) A 17,000 year glacio-eustatic sea level record: influence of glacial melting on the Younger Dryas event and deep-ocean circulation. Nature, **352**, 637-642.

Fard, A. M. M., Gruszka, B., Brunnberg, L. and Ringberg, B. (1997) Sedimentology of a glaciofluvial deposit at Ekeby, East Central Sweden. Quaternary Science Reviews, **16**, 755-765.

Fehrentz, M. and Radtke, U. (2001) Luminescence dating of Pleistocene outwash sediments of the Senne area (Eastern Munsterland) Germany. Quaternary Science Reviews, **20**, 725-729.

Ferguson, R. I. and Werritty, A. (1983) Bar development and channel changes in the gravelly River Feshie, Scotland. In: Collinson, J. D. and Lewin, J. (eds) **Modern and Ancient Fluvial Systems, Special Publication of the International Association of Sedimentologists No. 6**. Oxford, Blackwell, pp.181-193.

Fisher, T. G. (2004) River Warren boulders, Minnesota, USA: catastrophic paleoflow indicators in the southern spillway of glacial Lake Agassiz. Boreas, **33**, 349-358.

Forbes, D. L. (1983) Morphology and sedimentology of a sinuous gravel-bed channel system: lower Babbage River, Yukon coastal plain, Canada. In: Collinson, J. D. and Lewin, J. (eds) **Modern and Ancient Fluvial Systems, Special Publication of the International Association of Sedimentologists No. 6**. Oxford, Blackwell, pp.195-206.

Francis, E. A. (1970) The Quaternary. The Geology of Durham County. Transactions of the Natural History Society of Northumberland, **41**, 134–152.

Francis, E. A. (1975) Glacial Sediments: a selected review. In: Wright, A. E. and Moseley, F. (eds) **Ice Ages: Ancient & Modern**. Liverpool, Seel House Press, pp.43-68.

Fraser, G. S. (1993) Sedimentation in an interlobate outwash stream. Sedimentary Geology, **83**, 53-70.

Fraser, G. S. and Cobb, J. C. (1982) Late Wisconsinan proglacial sedimentation along the west Chicago moraine in northeastern Ilinois. Journal of Sedimentary Petrology, **52**(2), 473-491.

Frost, D. V. and Holliday, D. W. (1980) **Geology of the country around Bellingham.** Memoir for 1:50000 geological sheet 13. London, HMSO.

Fuller, I. C., Macklin, M. G., Passmore, D. G., Brewer, P. A., Lewin, J. and Wintle, A. G. (1996) Geochronologies and environmental records of Quaternary fluvial sequences in the Guadalope basin north-east Spain based on luminescence dating. In: Branson, J., Brown, A. G. and Gregory, K. (eds) **Global Continental Changes: the Context of Palaeohydrology**. London, Geological Society, pp.99-120.

Fulton, R. J. and Pullen, M. (1969) Sedimentation in Upper Arrow Lake. BC Can. J. Earth Sci, **6**, 785-791.

Gao, C., Keen, D. H., Boreham, S., Coope, G. R., Pettit, M. E., Stuart, A. J. and Gibbard, P. L. (2000) Last Interglacial and Devensian deposits of the River Great Ouse at Wollpack Farm, Fenstanton, Cambridgshire, UK. Quaternary Science Reviews, **19**, 787-810.

Gardiner, V. and Dackombe, R. V. (1983) **Geomorphological field manual**. London, Allen and Unwin. 254pp.

Gaunt, G. D. (1976) The Devensian maximum ice limit in the Vale of York. Proceedings of the Yorkshire Geological Society, **40**, 631-637.

Gemmell, A. M. D. (1999) IRSL from fine-grained glacifluvial sediment. Quaternary Geochronology, **19**, 207-215.

Gemmell, A. M. D., Murray, A. S. and Connell, E. R. (2007) Devensian glacial events in Buchan (N E Scotland): A progress report on new OSL dates and their implications. Quaternary Geochronology, **2**, 237-242.

Gibbard, P. L. (1977) Pleistocene history of the Vale of St Albans. Philosophical Transactions of the Royal Society, **B280**, 445-483.

Gibbard, P. L. (1979) Middle Pleistocene drainage in the Lower Thames Valley. Geological Magazine, **116**, 35-44.

Gibbard, P. L. (1988) The history of the great north-west European rivers during the past three million years. Philosophical Transactions of the Royal Society, **Series B**(318), 559-602.

Gilbert, G. K. (1890) **Lake Bonneville**. US Geological Survey Monograph 1. USGS. 438pp.

Giles, J. R. A. (1981) **The sand and gravel resources of the country around Blaydon, Tyne and Wear: description of the 1:25 000 resource sheet NZ 06**. London, HMSO.

Glanville, C. (1997) Glaciolacustrine and glaciofluvial deposits defining the margins of uncoupling ice lobes in the southeastern midlands of Ireland. Quaternary Science Reviews, **16**, 685-703.

Glen, J. W. (1954) The stability of ice-dammed lakes and other water-filled holes in glaciers. Journal of Glaciology, **2**, 316-318.

Goldthwait, R. P. and Matsch, C. eds. (1989) **Genetic Classification of Glacigenic Deposits**. Columbus, Ohio State University Press.

Gonzalez, S. and Cowell, R. (2004) Formby Point Foreshore, Merseyside: Palaeoenvironment and Archaeology. In: Chiverrell, R. C., Plater, A. and Thomas, G. (eds) **Quaternary of the Isle of Man & North West England, Field Guide**. Cambridge, Quaternary Research Association, pp.206-216.

Graf, W. L. (1970) The geomorphology of the glacial valley cross-section. Arctic and Alpine Research, **2**, 303-312.

Graham, D. F. and Grant, D. R. (1991) A test of airborne, side-looking synthetic aperture radar in central Newfoundland for geological reconnaissance. Canadian Journal of Earth Sciences, **28**, 257-265.

Gray, J. M. (1991) Glaciofluvial Landforms. In: Ehlers, J., Gibbard, P. L. and Rose, J. (eds) **Glacial Deposits in Great Britain and Ireland**. Rotterdam, Balkema.

Gray, J. M. and Sutherland, D. G. (1977) The 'Oban-Ford Moraine': a reappraisal. In: Gray, J. M. and Lowe, J. J. (eds) **Studies in the Scottish Lateglacial Environment**. Oxford, Pergamon, pp.33-44.

Greenwood, M. C. and Humphrey, N. (2002) Glaciated valley profiles: an application of nonlinear regression. In: **Proceedings of Symposia on Interfaces**, **Canada**.

Grootes, P. M., Stuiver, M., White, J. W. C., Johnsen, S. and Jouzel, J. (1993) Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. Nature, **366**, 552-554.

Gustavson, T. C. (1975) Bathymetry and sediment distribution in proglacial Malaspina Lake, Alaska. Journal of Sedimentary Research, **45**(2), 451-461.

Gustavson, T. C., Ashley, G. M. and Boothroyd, J. C. (1975) Depositional sequences in glaciolacustrine deltas. In: Jopling, A. V. and McDonald, B. C. (eds) **Glaciofluvial and Glaciolacustrine Sedimentation**. Tulsa, OK, Society of Economic Paleontologists and Mineralogists, pp.264-280.

Gustavson, T. C. and Boothroyd, J. C. (1987) A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analogue of the southeastern margin of the Laurentide Ice Sheet. Geological Society of America Bulletin, **99**, 187-200.

Hall, A. M. and Jarvis, R. A. (1989) A preliminary report on the Late Devensian glaciomarine deposits around St Fergus, Grampian Region. Quaternary Newsletter, **59**, 5-7.

Hambrey, M. J., Davies, J. R., Glasser, N. F., Waters, R. A., Dowdeswell, J. A., Wilby, P. R., Wilson, D. and Etienne, J. L. (2001) Devensian glacigenic sedimentation and landscape evolution in the Cardigan area of southwest Wales. Journal of Quaternary Science, **16**(5), 455-482.

Hancock, G. S., Anderson, R. S., Chadwick, O. A. and Finkel, R. C. (1999) Dating fluvial terraces with 10Be and 26Al Profiles: applications to the Wind River, Wyoming. Geomorphology, **27**, 41-60.

Hanson, P. R., Mason, J. A. and Goble, R. J. (2006) Fluvial terrace formation along Wyoming's Laramie Range as a response to increased late Pleistocene flood magnitudes. Geomorphology, **76**(1-2), 12-25.

Harbor, J. (1990) A discussion of Hirano and Anyia's (1988, 1989) explanation of glacial-valley cross profile development. Earth Surface Processes and Landforms, **15**, 369-377.

Harbor, J. and Wheeler, D. A. (1992) On the mathematical description of glaciated valley cross-sections. Earth Surface Processes and Landforms, **17**, 477-485.

Harrison, S. (1989) Late Devensian Solifluction Sheets in the Cheviot Hills. Unpublished PhD, Newcastle upon Tyne Polytechnic, Newcastle upon Tyne.

Harrison, S. (2002) Lithological variability of Quaternary slope deposits in the Cheviot Hills, UK. Proceedings of the Geologists' Association, **113**, 121-138.

Harvey, A. M., Oldfield, F., Baron, A. F. and Pearson, G. W. (1981) Dating of post-glacial landforms in the central Howgills. Earth Surface Processes and Landforms, **6**, 401-412.

Harvey, A. M. and Renwick, W. H. (1987) Holocene alluvial fan and terrace formation in the Bowland Fells Northwest England. Earth Surface Processes and Landforms, **12**, 249-257.

Hein, F. J. (1982) Depositional mechanisms of deep-sea coarse clastic sediments. Cap Enrage Formation, Quebec: Canadian Journal of Earth Sciences, **19**(2), 267–287.

Heinrich, H. (1988) Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years. Quaternary Research, **29**, 143-152.

Labyrinth. (1986) Directed by Jim Henson. TriStar Pictures LucasFilm Henson Associates.

Herdman, T. (1909) The Glacial Phenomena of the Vale of Derwent. Proceedings of the University of Durham Philosophical Society, **3**(3), 109-118.

Hinderer, M. (2001) Late Quaternary denudation of the Alps, valley and lake fillings and modern river loads. Geodinamica Acta, **14**, 231-263.

Hirano, M. and Aniya, M. (1988) A rational explanation of cross-profile morphology for glacial valleys and of glacial valley development. Earth Surface Processes and Landforms, **13**, 707-716.

Hirano, M. and Aniya, M. (1989) A rational explanation of cross-profile morphology for glacial valleys and of glacial valley development: a further note. Earth Surface Processes and Landforms, **14**, 173-174.

Hodkins, R., Cooper, R., Wadham, J. and Tranter, M. (2003) Suspended sediment fluxes in a high-Arctic glacierised catchment: implications for fluvial sediment storage. Sedimentary Geology, **162**, 105-117.

Hooke, J. M., Harvey, A. M., Miller, S. Y. and Redmond, C. E. (1990) The chronology and stratigraphy of the alluvial terraces of the River Dane valley Cheshire N W England. Earth Surface Processes and Landforms, **15**, 717-737.

Horton, B. P. (1999) The contemporary distribution of intertidal foraminifera of Cowpen Marsh, Tees Estuary, UK: implications for studies of Holocene sea-level changes. Palaeogeography, Palaeoclimatology, Palaeoecology Special Issue, **149** (1-4), 127-149.

Horton, B. P., Edwards, R. J. and Lloyd, J. M. (2000) Implications of a microfossil transfer function in Holocene sea-level studies. In: Shennan, I. and Andrews, J. E. (eds) **Holocene land-ocean interaction and environmental change around the western North Sea. Geological Society Special Publication**. London, Geological Society, pp.41-54.

Horton, B. P., Innes, J. B., Plater, A. J., Tooley, M. J. and Wright, M. R. (1999b) Post-glacial evolution and relative sea-level changes in Hartlepool Bay and the Tees estuary. In: Bridgland, D. R., Horton, B. P. and Innes, J. B. (eds) **The Quaternary of North-East England: Field Guide**. London, QRA, pp.65-86.

Horton, B. P., Innes, J. B. and Shennan, I. (1999a) Late Devensian and Holocene relative sea-level changes in Northumberland, England. In: Bridgland, D. R., Horton, B. P. and Innes, J. B. (eds) **The Quaternary of North-East England: Field Guide**. London, QRA, pp.35-47.

Horton, B. P., Innes, J. B., Shennan, I., Gehrels, W. R., Lloyd, J., McArthur, J. and Rutherford, M. (1999c) The Northumberland Coast. In: Bridgland, D. R., Horton, B. P. and Innes, J. B. (eds) **The Quaternary of North-East England: Field Guide**. London, QRA, pp.147-165.

Howard, A. J., Keen, D. H. and Hollin, J. T. (1999a) Amino acid dating of a molluscan fauna from Bassingham Fen, Lincolnshire: implications for the chronology of the Trent Terraces. Proceedings of the Geologists' Association, **110**, 233-239.

Howard, A. J., Keen, D. H., Mighall, T. M., Coope, G. R., Griffiths, H. I. and Macklin, M. G. (2000) Early Holocene environments of the River Ure near Ripon, North Yorkshire. Proceedings of the Yorkshire Geological Society, **53**(1), 31-42.

Howard, A. J. and Macklin, M. G. (1999) A generic geomorphological approach archaeological interpretation and prospection in British river valleys: A guide for archaeologists investigating Holocene landscapes. ANTIQUITY, **73**, 527-41.

Howard, A. J., Macklin, M. G., Black, S. and Hudson-Edwards, K. A. (1999b) Holocene river development and environmental change in Upper Wharfedale Yorkshire Dales England. Journal of Quaternary Science, **15**(3), 239-252.

Howse, R. (1863) Notes on the fossil remains of some recent and extinct Mammalia found in the counties of Northumberland and Durham. Transactions of the Tyneside Naturalists Field Club, **V**, 111-121.

Hubbard, A., Bradwell, T. and Stoker, M. (2007) **Modelling the dynamic response of the Last British Ice Sheet: abstract**. **QRA Annual Discussion Meeting: The**

Growth, Maximum Extent and Deglaciation of the Last British Ice Sheet, St Andrews.

Huddart, D. (1999) Supraglacial trough fills, southern Scotland: origins and implications for deglacial processes. **Glacial Geology and Geomorphology**, Available from: <http://boris.qub.ac.uk/ggg/papers/full/1999/rp041999/rp04.html> [Accessed 30 September 2004].

Huddart, D. and Bennett, M. R. (1997) The Carstairs Kames (Lanarkshire, Scotland): morphology, sedimentology and formation. Journal of Quaternary Science, **12**, 467-484.

Hughes, D. and Teasdale, D. (1999) Herrington Colliery opencast coal site. In: Bridgland, D. R., Horton, B. P. and Innes, J. B. (eds) **The Quaternary of north-east England**. London, QRA, pp.137-145.

Hughes, M. L., McDowell, P. F. and Marcus, W. A. (2006) Accuracy assessment of georectified aerial photographs: Implications for measuring lateral channel movement in a GIS. Geomorphology, **74**, 1-16.

Huisink, M. (1999) Lateglacial river sediment budgets in the Maas Valley, The Netherlands. Earth Surface Processes and Landforms, **24**, 93-109.

Huntley, D. J., Godfrey-Smith, D. I. and Thewalt, M. L. W. (1985) Optical dating of sediiments. Nature, **313**, 105-107.

Huntley, D. J. and Lian, O. B. (1999) Using optical dating to determine when a sediment was last exposed to sunlight. Holocene Climate and Environmental Change in the Palliser Triangle: A Geoscientific Context for Evaluating the Impacts of Climate Change on the Southern Canadian Prairies. Geological Survey of Canada, Ottawa, Bulletin, **534**(295), 211-222.

Jacobs, Z., Duller, G. A. T. and Wintle, A. G. (2003) Optical dating of dune sands from Blombos Cave, South Africa: II-single grain data. Journal of Human Evolution, **44**, 613-625.

James, L. A. (1996) Polynomial and power functions for glacial valley cross-section morphology. Earth Surface Processes and Landforms, **21**, 413-432.

Jarvis, R. A. (1977) **Soils of the Hexham district Sheet 19**, 1:63360. Silsoe, National Soil Resources Institute.

Johannsson, C. E. (1976) Structural studies of frictional sediments. Geografiska Annaler, **58A**(4), 201-301.

Johnson, G. A. L. (1997) **Geology of Hadrian's Wall. Geologists' Association Guide No. 59**. London, GA.

Johnson, W. H. and Menzies, J. (2002) Supraglacial and ice-marginal deposits and landforms. In: Menzies, J. (ed.) **Modern and Past Glacial Environments**. Oxford, Butterworth Heinman, pp.317-333.

Johnstone, E., Macklin, M. G. and Lewin, J. (2006) The development and application of a database of radiocarbon-dated Holocene fluvial deposits in Great Britain. Catena, **66**, 14-23.

Jones, A. F., Brewer, P. A., Johnstone, E. and Macklin, M. G. (2007) High-resolution interpretative geomorphological mapping of river valley environments using airborne LiDAR data. EARTH SURFACE PROCESSES AND LANDFORMS, **32**(10), 1574.

Jones, A. P., Tucker, M. E. and Hart, J. K. (eds.) (1999) **The Description and Analysis of Quaternary Stratigraphic Field Sections. Technical Guide 7**. London, Quaternary Research Association.

Jones, R. L. (1977) Late Devensian deposits from Kildale, northeast Yorkshire. Proceedings of the Yorkshire Geological Society, **41**, 185-188.

Jopling, A. V. and Walker, R. G. (1968) Morphology and origin of ripple-drift cross-laminations, with examples from the Pleistocene of Massachusetts. Journal of Sedimentary Petrology, **38**, 971-984.

Jordan, G., Mejininger, B. M. L., van Hinsbergen, D. J. J., E., M. J. and van Dijk, P. M. (2005) Extraction of morphotectonic features from DEMs: Development and applications for study areas in Hungary and NW Greece. International Journal of Applied Earth Observation and Geoinformation, **7**, 163-182.

Keen, D. H. (1990) Significance of the record provided by Pleistocene fluvial deposits and their included molluscan faunas for palaeoenvironmental reconstruction and stratigraphy: case studies from the English Midlands. Palaeogeography, Palaeoclimatology, Palaeoecology, **80**, 25-34.

Keen, D. H. (2001) Towards a late Middle Pleistocene non-marine molluscan biostratigraphy for the British Isles. Quaternary Science Reviews, **20**, 1657-1665.

Keen, D. H., Jones, R. L. and Robinson, J. E. (1984) A Late Devensian fauna and flora from Kildale north-east Yorkshire. Proceedings of the Yorkshire Geological Society, **44**, 385-397.

Kendall, P. F. (1902) A system of glacier lakes in the Cleveland Hills. Quaterley Journal of the Geological Society of London, **58**, 471-571.

King, C. A. M. (1976) **The Geomorphology of the British Isles: Northern England**. London, Methuen.

Kjær, K. H., Sultan, L., Krüger, J. and Schomacker, A. (2004) Architecture and sedimentation of outwash fans in front of the Myrdalsjokull ice cap, Iceland. Sedimentary Geology, **172**, 139-163.

Knight, J. (2003) Geomorphic and sedimentary evidence for patterns of late Midlandian ice retreat in the Tempo Valley, north-central Ireland. Irish Geography, **36**(2), 127-144.

Knight, J. (2006) Geomorphic evidence for active and inactive phases of Late Devensian ice in north-central Ireland. Geomorphology, **75**, 4-19.

Knighton, D. (1998) **Fluvial Forms and Processes: a new perspective**. 2nd edition. London, Arnold. 383pp.

Knox, J. C. (1983) Responses of river systems to Holocene climates. In: Wright Jr, H. E. (ed.) **Late-Quaternary Environment of the United States: Vol. 2 the Holocene**. Minneapolis, University of Minnesota Press, pp.26–41.

Knuepfer, P. L. K. (2004) Extracting tectonic and climatic signals from river and terrace long profiles in active orogens, Taiwan and New Zealand. 2004 Denver Annual Meeting.

Koster, E. H. (1978) Traverse ribs: their characteristics, origin and paleohydraulic significance. In: Miall, A. D. (ed.) **Fluvial Sedimentology**. Calgary, Canadian Society of Petroleum Geologists, pp.161-186.

Koutsoukos, P. G. and Kontoyannis, C. G. (1984) Precipitation of Calcium Carbonate in Aqueous Solutions. Journal of the Chemical Society, Faraday Transactions, **1**(80), 1181-1192.

Krainer, K. and Poscher, G. (1990) Ice-rich, redeposited diamicton blocks and associated structures in Quaternary outwash sediments of the Inn Valley near Insbruck, Austria. Geografiska Annaler, **72A**(3-4), 249-254.

Krigstrom, A. (1962) Geomorphological studies of sandur plains and their braided rivers in Iceland. Geografiska Annaler, **44**, 328-346.

Krivoruchko, K. and Gotway, A. C. (2004) Creating Exposure Maps Using Kriging. [Internet]. Available from: <http://www.esri.com/software/arcgis/arcgisextensions/geostatistical/research_paprs.html> [Accessed 25 July 2007].

Lambeck, K. (1993) Glacial Rebound of the British Isles. II: a high resolution, high-precision model. Geophysial Journal International, **115**, 960-990.

Lambeck, K. (1995) Late Devensian and Holocene shorelines of the British Isles and North Sea from models of glacio-hydro-isostatic rebound. Journal of the Geological Society, **152**, 437-448.

Land, D. H., Calver, M. A. and Harrison, R. K. (1974) **Geology of the Tynemouth District.** Memoir (Sheet) Geological Survey of GB (England & Wales) - New Series (15). London, HMSO.

Lang, A. and Nolte, S. (1999) The Chronology of Holocene alluvial sediments from the Wetterau, Germany, provided by optical and 14C dating. The Holocene, **9**(2), 207-214.

Laslett, G. (1984) Kriging and splines: an empirical comparison of their predictive performance in some applications. Journal of the American Statistical Association, **89**(426), 391-405.

Laslett, G. and McBratney, A. B. (1990) Estimation and implications of instrumental drift, measurement error and nugget variance of spatial soil attributes – a case study for soil pH. Journal of Soil Science, **41,** 451-471.

Laslett, G. M., McBratney, A. B., Pahl, P. J. and Hutchinson, M. F. (1987) Comparison of several spatial prediction methods for soil pH. European Journal of Soil Science, **38**(2), 325-341.

Lawrence, D. J. D. and Jackson, I. (1986) **Geology of the Ponteland-Morpeth district: 1:10000 sheets NZ17NE, SE and NZ18NE, SE parts of 1:50000 sheets 9 (Rothbury) and 14 (Morpeth)**. British Geological Survey, Geological Notes and Local Details for 1:10000 Sheets 1986 - 38: ill, maps. London, HMSO.

Lebour, G. A. L. (1878) **Outlines of the Geology of Northumberland**. Newcastle upon Tyne, Lambert & Co.

Lebour, G. A. L. (1884) Brief notes on the geology of Corbridge, Northumberland. Proceedings of the Berwickshire Naturalists' Club 1882-1884, **10**, 121-122.

Lebour, G. A. L. (1889) Recent and Sub-Recent deposits. In: Lebour, G. A. (ed.) **Handbook of Geology and Natural History**. Newcastle upon Tyne, Lambert and Co.

Lebour, G. A. L. (1893) On certain surface features of the glacial deposits of the Tyne Valley. Transactions of the Natural History Society of Northumberland, Durham and Newcastle upon Tyne, **9**, 191-195.

Lepper, K. and McKeever, S. W. S. (2002) An objective methodology for dose distribution analysis. Radiation Measurements, **23**, 555-561.

Levson, V. M. and Rutter, N. W. (1988) A lithofacies analysis and interpretation of depositional environments of montane glacial diamictons, Jasper, Alberta, Canada. Genetic Classification of Glacigenic Deposits: Rotterdam, Balkema, 117–140.

Lewin, J. and Macklin, M. G. (2003) Preservation potential for Late Quaternary river alluvium. Journal of Quaternary Science, **18**(2), 107-120.

Lewin, J., Macklin, M. G. and Johnstone, E. (2005) Interpreting alluvial archives: sedimentological factors in the British Holocene fluvial record. Quaternary Science Reviews, **24**(16-17), 1873-1889.

Lewis, S. G. and Maddy, D. (1999) Description and analysis of Quaternary fluvial sediments: a case study from the upper River Thames, UK. In: Jones, A. P., Tucker, M. E. and Hart, J. K. (eds) **The description and analysis of Quaternary stratigraphic field sections. Technical Guide 7**. London, Quaternary Research Association.

Lewis, S. G., Maddy, D. and Scaife, R. (2001) The fluvial system response to abrupt climate change during the last cold stage: the Upper Pleistocene River Thames fluvial succession at Ashton Keynes, UK. Global and Planetary Change, **28**, 341-359.

Li, J., Taylor, G. and Kidner, D. B. (2005) Accuracy and reliability of map-matched GPS coordinates: the dependance on terrain model resolution and interpolation algorithm. Computers and Geosciences, **31**, 241-251.

Li, X., Tennant, K. and Lawrence, G. (2004) Three-dimensional mapping with airborne IFSAR based Star Technology Intermap's experiences. Proceedings of, 12-23.

Li, Y., Liu, G. and Cui, Z. (2001) Glacial valley cross-profile morphology, Tien Shan Mountains, China. Geomorphology, **38**(1-2), 153-166.

Lian, O. V. and Roberts, R. G. (2006) Dating the Quaternary: progress in luminescence dating of sediments. Quaternary Science Reviews, **25**, 2449-2468.

Long, A. J. and Shennan, I. (1993) Holocene relative sea-level and crustal movements in Southeast and Northeast England. In: Owen, L. A., Stewart, I. and Vita-Finzi, C. (eds) **Quaternary Proceedings**. Cambridge, Quaternary Research Association, pp.15-19.

Lønne, I. (1993) Physical signatures of ice advance in a younger dryas ice-contact delta, Troms, Northern Norway: implications for glacier-terminus history. Boreas, **22**, 59-70.

Lovell, J. H. (1981) **The sand and gravel resources of the country around Hexham, Northumberland: description of the 1:25 000 resource sheet NY 86 and 96**. London, HMSO.

Lowe, D. R. (1982) Sediment gravity flows; II, Depositional models with special reference to the deposits of high-density turbidity currents. Journal of Sedimentary Research, **52**(1), 279-297.

Lowe, J. J. and Walker, M. J. C. (1997) **Reconstructing Quaternary Environments**. 2nd edition. Essex, Longman.

Lukas, S., Spencer, J. Q. G., Robinson, R. A. J. and Benn, D. I. (2007) Problems associated with luminescence dating of Late Quaternary glacial sediments in NW Scottish Highlands. Quaternary Geochronology, **2**, 243-248.

Lunkka, J. P. and Gibbard, P. L. (1996) Ice-marginal sedimentation and its implications for ice-lobe deglaciation patterns in the Baltic region: Pohjankangas, western Finland. Journal of Quaternary Science, **11**(5), 377-388.

Lunn, A. (1980) Quaternary. In: Robson, D. A. (ed.) **The Geology of north east England**. Newcastle upon Tyne, The Natural History Society of Northumbria, pp.48-60.

Lunn, A. (1995) The Quaternary of South Tynedale. In: Scrutton, C. T. (ed.) **Northumbrian Rocks and Landscapes**. Leeds, Ellenbank Press, pp.137-145.

MacKay, H. (2005) Mapping Great Britain at 1-Metre Z Values. **Intermap Technologies White Papers** [Internet]. Available from: <http://www.intermap.com/customer/papers.cfm> [Accessed 21 April 2005].

Macklin, M. G. (1997) Fluvial Geomorphology of north-east England. In: Gregory, K. J. (ed.) **Fluvial Geomorphology of Great Britain**. London, Chapman & Hall, pp.208-238.

Macklin, M. G. (1999) Holocene river environments in prehistoric Britain: Human interaction and impact. Journal of Quaternary Science, **14**(6), 521-530.

Macklin, M. G. and Aspinall, R. J. (1986) Historic floodplain sedimentation in the River West Allan Northumberland: a case study of channel change in an upland gravel-bed river in the northern Pennines. In: Macklin, M. G. and Rose, J. (eds) **Quaternary River Landforms and sediments in the Northern Pennines**. Oxford, British Geomorphological Research Group/Quaternary Research Association.

Macklin, M. G., Johnstone, E. and Lewin, J. (2005) Pervasive and long-term forcing of Holocene river instability and flooding in Great Britain by centennial-scale climate change. The Holocene, **15**(7), 937.

Macklin, M. G. and Lewin, J. (1986) Terrace fills of Pleistocene and Holocene age in the Rheidol Valley Wales. Journal of Quaternary Science, **1**, 21-34.

Macklin, M. G. and Lewin, J. (1989) Sediment Transfer and transformation of an alluvial valley floor: the river South Tyne Northumbria UK. Earth Surface Processes and Landforms, **14**, 233-246.

Macklin, M. G. and Lewin, J. (1993) Holocene river alluviation in Britain. Zeitschrift fur Geomorphologie, **47**, 109-122.

Macklin, M. G. and Lewin, J. (1997) Channel, floodplain and drainage basin response to environmental change. In: Thorne, C. R., Hey, R. D. and Newson, M. D. (eds) **Applied Fluvial Geomorphology for River Engineering and Management**. Chichester, Wiley, pp.15-45.

Macklin, M. G. and Lewin, J. (2003) River sediments, great floods and centennial-scale Holocene climate change. Journal of Quaternary Science, **18**(2), 101-105.

Macklin, M. G., Passmore, D. G. and Newson, M. D. (1998) Controls of short and long term river instability: processes and patterns in gravel-bed rivers the Tyne Basin northern England. In: Klingman, P., Beschta, R. L., Komar, P. D. and Bradley, J. B. (eds) **Proceedings of the 4th International Gravel Bed Rivers Conference.**

Macklin, M. G., Passmore, D. G. and Rumsby, B. T. (1992a) Climatic and cultural signals in Holocene alluvial sequences: the Tyne basin, northern England. In: Needham, S. and Macklin, M. G. (eds) **Alluvial Archaeology in Britain**. Oxford, Oxbow, pp.123-140.

Macklin, M. G., Rumsby, B. T. and Heap, T. (1992b) Flood Alluviation and entrenchment: Holocene valley floor development and transformation in the British Uplands. Geological Society of America Bulletin, **104**, 631-643.

Macklin, M. G., Rumsby, B. T., Heap, T. and Passmore, D. G. (1994) Thinhope Burn Northumberland. In: Boardman, J. and Walden, J. (eds) **The Quaternary of Cumbria: Field Guide**. Oxford, Quaternary Research Association, pp.50-57.

Maddy, D. (1989) The Middle Pleistocene development of the rivers Severn and Avon. Unpublished PhD Thesis, University of London, London.

Maddy, D. (1997) Uplift driven valley incision and river terrace formation in southern England. Journal of Quaternary Science, **12**, 539-545.

Maddy, D. (2002) An evaluation of climate, crustal movement and base level controls on the Middle-Late Pleistocene development of the River Severn, UK. Geologie en Mijnbouw/Netherlands Journal of Geoscience, **81**, 329-338.

Maddy, D. and Bridgland, D. R. (2000) Accelerated uplift resulting from Anglian glacioisostatic rebound in the Middle Thames Valley, UK? evidence from the river terrace record. Quaternary Science Reviews, **19**, 1581-1588.

Maddy, D., Bridgland, D. R. and Green, C. P. (2000) Crustal uplift in southern England: evidence from the river terrace records. Geomorphology, **33**, 167-181.

Maddy, D., Bridgland, D. R. and Westaway, R. (2001) Uplift-driven valley incision and cliamte-controlled river terrace development in the Thames valley, UK. Quaternary International, **79**), 23-36.

Maddy, D., Green, C. P., Lewis, S. G. and Bowen, D. Q. (1995) Pleistocene Geology of the Lower Severn Valley, UK. Quaternary Science Reviews, **14**, 209-222.

Maddy, D., Keen, D. H., Bridgland, D. R. and Green, C. P. (1991) A revised model for the Pleistocene development of the River Avon, Warwickshire. Journal of the Geological Society of London, **148**, 473 484.

Maizels, J. (1989) Sedimentology, paleoflow dynamics and flood history of jökulhlaup deposits: paleohydrology of Holocene sediment sequences in southern Iceland sandur deposits. Journal of Sedimentary Petrology, **59**, 204-223.

Maizels, J. K. (1982) Channel Changes, Palaeohydrology and Deglaciation: Evidence from some Lateglacial Sandur deposits of Northeast Scotland. Quaternary Studies in Poland, **4**, 171-187.

Maizels, J. K. (2002) Sediment and Landforms of modern proglacial terrestrial environments. In: Menzies, J. (ed.) **Modern and Past Glacial Environments**. Oxford, Butterworth Heinman, pp.279-316.

Maizels, J. K. and Aitken, J. F. (1991) Palaeohydrological change during Deglaciationn in Upland Britain: A case study from northeast Scotland. In: Starkel, L., Gregory, K. J. and Thornes, J. B. (ed.) **Temperate Palaeohydrology**. West Sussex, Wiley, pp.105-145.

Marren, P. M. (2001) Sedimentology of proglacial rivers in eastern Scotland during the Late Devensian. Transactions of the Royal Society of Edinburgh, **92**, 149-171.

Marren, P. M. (2002) Fluvial-lacustrine interaction on Skeioararsandur, Iceland: implications for sandur evolution. Sedimentary Geology, **149**, 43-58.

McCabe, A. M. (1996) Dating and rhythmicity from the last deglacial cycle in the British Isles. Journal of the Geological Society, London, **153**, 499-502.

McCabe, A. M. and Clark, P. U. (1998) Ice sheet variability around the North Atlantic Ocean during the last deglaciation. Nature, **392**, 373-377.

McCabe, A. M. and Clark, P. U. (2003) Deglacial chronology from County Donegal, Ireland: implications for the deglaciation of the British-Irish ice sheet. Journal of the Geological Society, London, **160**, 847-855.

McCabe, A. M., Clark, P. U., Smith, D. E. and Dunlop, P. (2007) A revised model for the last deglaciation of eastern Scotland. Journal of the Geological Society, London, **164**, 313-316.

McCabe, A. M., Dardis, G. F. and Hanvey, P. M. (1987) Sedimentology at the margins of Late-Pleistocene ice-lobe terminating in shallow marine environments, Dundalk Bay, eastern Ireland. Sedimentology, **34**, 473-493.

McCabe, A. M. and Eyles, N. (1988) Sedimcntology of an ice-contact glaciomarine delta, Carey Valley, Northern Ireland. Sedimentary Geology, **29**, 1-14.

McCabe, A. M., Knight, J. and McCarron, S. G. (1999) Ice-flow stages and glacial bedforms in north central Ireland: a record of rapid environmental change during the last glacial terminiation. Journal of the Geological Society, London, **156**, 63-72.

McCarroll, D. and Rijsdijk, K. F. (2003) Deformation styles as a key for interpreting glacial depositional environments. Journal of Quaternary Science, **18**, 473-489.

McCoy, D. G., Prescott, J. R. and Nation, R. J. (2000) Some aspects of single-grain luminescence dating. Radiation Measurements, **32**, 859-864.

McDonald, B. C. and Banerjee, I. (1971) Sediments and Bed Forms on a Braided Outwash Plain. Canadian Journal of Earth Sciences, **8**, 1282-1301.

McDonald, B. C. and Shilts, W. W. (1975) Interpretation of faults in glaciofluvial sediments. In: Jopling, A. V. and McDonald, B. C. (eds) **Glaciofluvial and Glaciolacustrine Sedimentation**. Tulsa, Oklahoma, Society of Economic Paleontologists and Mineralogists, pp.123-131.

McKee, E. D., Crosby, E. J. and Berryhill, H. L. (1967) Flood deposits, Bijou Creek, Colorado, June 1965. Journal of Sedimentary Petrology, **37**, 829-851.

Menzies, J. (2002) **Modern and Past Glacial Environments**. Oxford, Butterworth Heinemann. 580pp.

Mercer, B. (2004) DEMs created from Airborbe IFSAR - An Update. In: Altan, O. (ed.) **Proceedings of the XXth ISPRS Congress Geo-Imagery Bridging Continents**, **12-23 July 2004, Istanbul, Turkey**, pp.841-847.

Merritts, D. J., Vincent, K. R. and Wohl, E. E. (1994) Long river profiles, tectonism, and eustasy: A guide to interpreting fluvial sequences. Journal of Geophysical Research, **99**(B7), 14031-14050.

Miall, A. D. (1977) A review of Braided-River depositional environment. Earth Science Reviews, **13**, 1-62.

Miall, A. D. (1978) Lithofacies types and vertical profile models in braided river deposits: a summary. In: Miall, A. D. (ed.) **Fluvial Sedimentology**. Canadian Society of Petroleum Geologists, pp.597-604.

Middleton, G. V. and Hampton, M. A. (1973) Sediment gravity flows: mechanics of flow and deposition. Turbidites and Deep-Water Sedimentation, 1–38.

Mills, D. A. C. and Holliday, D. W. (1998) **Geology of the district around Newcastle upon Tyne, Gateshead and Consett**. Memoirs of the British Geological Society Sheet 20. London, HMSO.

Mills, H. H. (1991) Three-dimensional clast orientation in glacial and mass-movement sediments: a compilation and preliminary analysis. US Geological Survey, Open File Report, 90-128.

Mitchell, W. A. (1991) Geomorphological Mapping: An Introduction. In: Mitchell, W. A. (ed.) **Western Pennines. Field Guide**. Cambridge, Quaternary Research Association, pp.19-23.

Moore, P. D., Webb, J. A. and Collinson, M. (1991) **Pollen Analysis**. 2nd. Oxford, Blackwell. 216pp.

Moores, A. J., Passmore, D. G. and Stevenson, A. C. (1999) High resolution Palaeochannel records of Holocene valley floor environments in the North Tyne Basin, northern England. In: Brown, A. G. and Quine, T. A. (eds) **Fluvial Processes and Environmental Change**. Chichester, Wiley, pp.283-310.

Morison, S. R. and Hein, F. J. (1987) Sedimentology of the White Channel gravels, Klondike area, Yukon Territory: Fluvial deposits of a confined valley. In: Ethridge, F. G., Flores, R. M. and Harvey, M. D. (eds) **Recent Developments in Fluvial Sedimentology.** Tulsa, Oklahoma, USA, The Society of Economic Paleontologists and Mineralogists, pp.205 - 216.

Mulholland, J. W. (1982) Glacial stagnation-zone retreat in New England: Bedrock control. Geology, **10**, 567-571.

Munro, M. and Shaw, J. (1997) Erosional origin of hummocky terrain in south-central Alberta, Canada. Geology, **25**(11), 1027-1030.

Murray, A. S. and Mejdahl, V. (1999) Comparison of regenerative-dose single-aliquot and multiple-aliquot (SARA) protocols using heated quartz from archaeological sites. Quaternary Science Reviews, **18**(2), 223.http://www.sciencedirect.com/science/article/B6VBC-3WC3R1G-6/2/61a25bf6e07aa80dcee6815c09d453f1.

Murray, A. S. and Olley, J. M. (2002) Precision and accuracy in the opticallystimulated luminescence dating of sedimentary quartz: a status review. Geochronometria, **21**, 1-16.

Murray, A. S. and Wintle, A. G. (2000) Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. Radiation Measurements, **32**, 57-73.

Murton, J. B., Baker, A., Bowen, D. Q., Caseldine, C. J., Coope, G. R., A.P., C., Evans, J. G., Field, M. H., Green, C. P., Hatton, J., Ito, M., Jones, R. L., Keen, D. H., Kerney, M. P., R., M., McGregor, D. F. M., Parish, D., J.E., R., Schreve, D. C. and Smart, P. L. (2001) A late Middle Pleistocene temperate-periglacial-termperate sequence (Oxygen Isotope Stages 7-5e) near Marsworth, Buckinghamshire, UK. Quaternary Science Reviews, **20**, 1787-1825.

Murton, J. B. and Kolstrup, E. (2003) Ice-wedge casts as indicators of palaeotemperatures: precise proxy or wishful thinking? Progress in Physical Geogrpahy, **27**(2), 155-170.

Napieralski, J., Harbor, J. and Yingkui, L. (2007) Glacial geomorphology and geographic information systems. Earth Science Reviews, **85**, 1-22.

National Air Photo Library. (2004) Air Photo Mosaics. **Natural Resources Canada** [Internet]. Available from: <http://airphotos.nrcan.gc.ca/mos_e.php> [Accessed 20 April 2007].

Nemec, W. and Steel, R. J. (1984) Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits. In: Koster, E. H. and Steel, R. J. (eds) **Sedimentology of Gravels and Conglomerates**. Canadian Society of Petroleum Geologists Memoir No. 10, pp.1-31.

North of England Institute of Mining Engineers. (1878-1910) **An account of the strata of Northumberland and Durham as proved by borings and sinkings**. Newcastle upon Tyne.

Northumberland County Council. (2000) **Northumberland Minerals Local Plan**. Morpeth, Northumberland County Council, pp.106.

O'Cofaigh, C., Pudsey, C. J. and Dowdeswell, J. (2002) Evolution of subglacial bedforms along a paleo-ice stream, Antarctic Peninsula continental shelf. Geophysical Research Letters, **29**(8), 41-1. http://www.agu.org/pubs/sample_articles/sp/2001GL014488/2001GL014488.pdf.

Oguchi, T., Aoki, T. and Matsuta, N. (2003) Identification of an active fault in the Japanese Alps from DEM-based hill shading. Computers and Geosciences, **29**, 885-891.

Olley, J. M., de Deckker, P., Roberts, R. G., Fifield, L. K., Yoshida, H. and Hancock, G. S. (2004) Optical dating of deep-sea sediments using single grains of quartz: a comparison with radiocarbon. Sedimentary Geology, **169**, 175-189.

Onorati, G., Poscolieri, M., Ventura, R., Chiarini, V. and Crucilla, U. (1992) The digital elevation model of Italy for geomorphology and structural geology. Catena, **19**, 147-178.

Owen, L. A., Kamp, U., Spencer, J. Q. and Haserodt, K. (2002) Timing and style of Late Quaternary glaciation in the eastern Hind Kush, Chitral, northern Pakistan: a review and revision of glacial chronology based on new optically stimulated luminescence dating. Quaternary International, **97-98**, 41-55.

Passmore, D. G. and Macklin, M. G. (1994) Provenance of fine-grained alluvium and late Holocene land-use change in the Tyne. Geomorphology, **9**, 127-142.

Passmore, D. G. and Macklin, M. G. (1997) Geoarchaeology of the Tyne Basin: Holocene river valley environments and the archaeological record. In: Tolan-Smith, C. (ed.) **Landscape Archaeology in Tynedale**. Department of Archaeology University of Newcastle upon Tyne.

Passmore, D. G. and Macklin, M. G. (2000) Late Holocene channel and floodplain development in a wandering gravel-bed river: the River South Tyne at Lambley, Northern England. Earth Surface Processes and Landforms, **25**, 1237-1256.

Passmore, D. G. and Macklin, M. G. (2001) Holocene sediment budgets in an upland gravel bed river: the South Tyne, northern England. River Basin Sediments Systems: Archives of Environmental Change, Maddy D, Macklin M, Woodward J (eds). Balkema: Rotterdam, 423–444.

Passmore, D. G., Macklin, M. G., Brewer, P. A., Lewin, J., Rumsby, B. T. and Newson, M. D. (1993) Variability of late Holocene braiding in Britain. In: Best, J. L. and Bristow, C. S. (eds) **Braided Rivers. Geological Society Special Publication No. 75**. London, Geological Society, pp.205-229.

Passmore, D. G., Macklin, M. G., Stevenson, A. C. and O'Brien, C. F. (1992) A Holocene alluvial sequence in the lower Tyne valley northern Britain: a record of river response to environmental change. The Holocene, **2**, 138-147.

Paterson, J. T. and Cheel, R. J. (1997) The depositional history of the Bloomington complex, an ice-contact deposit in the Oak Ridges moraine, Southern Ontario, Canada. Quaternary Science Reviews, **16**, 705-719.

Pattyn, F. and Van Huele, W. (1998) Power Law or Power Flaw? Earth Surface Processes and Landforms, **23**, 761-767.

Pebesma, E. J. and Wesseling, C. G. (1998) Gstat: a program for geostatistical modelling, prediction and simulation. Computers and Geosciences, **24**(1), 17-31.

Peel, R. F. (1941) The North Tyne Valley. The Geographical Journal, **98**, 5-19.

Peel, R. F. (1951) A study of two Northumbrian Spillways. Transactions of the Institute of British Geographers, **15**, 73-89.

Peel, R. F. (1956) The Profiles of Glacial Drainage Channels. Geological Journal, **122**, 483-487.

Penny, L. F., Coope, G. R. and Catt, J. A. (1969) Age and insect fauna of the Dimlington silts, East Yorkshire. Nature, **224**, 65-67.

Petts, G. E. and Foster, I. D. L. (1985) **Rivers and Landscape**. London, Arnold. 274pp.

Phillips, E. R., Evans, D. J. A. and Auton, C. A. (2002) Polyphase deformation at an oscillating ice margin following the Loch Lomond Readvance, central Scotland, UK. Sedimentary Geology, **149**, 157-182.

Phillips, F. M., Bowen, D. Q. and Elmore, D. (1996) Surface exposure dating of glacial features in Great Britain using cosmogenic chlorine-36: preliminary results. Mineralogical Magazine 58A, 722–723.

Plater, A. J. and Shennan, I. (1992) Evidence of Holocene sea-level change from the Northumberland coast, eastern England. Proceedings of the Geologists' Association, **103**, 201-216.

Postma, G., Nemec, W. and Kleinspehn, K. L. (1988) Large floating clasts in turbidites: a mechanism for their emplacement. Sedimentary Geology, **58**, 47-61.

Postma, G., Roep, T. B. and Ruegg, G. H. J. (1983) Sandy-gravelly mass-flow deposits in an ice-marginal lake (Saalian, Leuvenumsche Beek Valley, Veluwe, the Netherlands), with emphasis on plug-flow deposits. Sedimentary Geology, **34**, 59-82.

Price, R. J. (1973) **Glacial and fluvioglacial landforms**. Edinburgh, Oliver and Boyd. 242pp.

Prima, O. D. A., Echigo, A., Yokoyama, R. and Yoshida, T. (2006) Supervised landform classification of Northeast Honshu from DEM-derived thematic maps. Geomorphology, **78** (3-4), 77-89.

QRA. (2007) The growth, maximum extent and deglaciation of the last British and Irish Ice Sheets. **QRA Annual Discussion Meeting**, **St Andrews, Scotland**, pp.82.

Raistrick, A. (1931) Glaciation. Proceedings of the Geologists' Association, **42**, 281-291.

Rasmussen, S. O., Vinther, B. M., Clausen, H. B. and Andersen, K. K. (2007) Early Holocene climate oscillations recorded in three Greenland ice cores. Quaternary Science Reviews, **26**(15-16), 1907-1914.

Repka, J. L., Anderson, R. S. and Finkel, R. C. (1997) Cosmogenic dating of fluvial terraces, Fremont River, Utah. Earth and Planetary Science Letters, **152**, 59-73.

Rhind, D. W. (1969) The Terraces of the Tweed Valley. Unpublished PhD, University of Edinburgh.

Richardson, G. (1983) **Geological notes and local details for 1:10 000 sheets NZ26NW, NE, SW and SE, Newcastle upon Tyne and Gateshead: part of 1:50 000 sheets 14 (Morpeth), 15 (Tynemouth), 20 (Newcastle upon Tyne) and 21 (Sunderland)**. Institute of Geological Sciences, Geological Notes and Local Details for 1:10 000 Sheets. Keyworth, HMSO.

Ripley, B. D. (1981) **Spatial Statistics**. New York, Wiley & Sons. 252pp.

Rittenour, T. A., Goble, R. J. and Blum, M. D. (2005) Development of an OSL chronology for Late Pleistocene channel belts in the lower Mississippi valley, USA. Quaternary Science Reviews, **24**(23-24), 2539-2554.

Roberts, D. H., Dackombe, R. V. and Thomas, G. S. P. (2007) Palaeo-ice streaming in the central sector of the British-Irish Ice Sheet during the Last Glacial Maximum: evidence from the northern Irish Sea Basin. Boreas, **36**(2), 115-129.

Roberts, N. (1998) **The Holocene: An environmental history**. 2nd. Oxford, Blackwell.

Roberts, R. G., Bird, M., Olley, J. M., Galbraith, R., Lawson, E., Laslett, G., Yoshida, H., Jones, R., Fullager, R., Jacobsen, G. and Hua, Q. (1998) Optical and radiocarbon dating at Jinmium rock shelter in northern Australia. Nature, **393**, 358-362.

Roberts, R. G., Galbraith, R., Olley, J. M., Yoshida, H. and Laslett, G. (1999) Optical dating of single and multiple grains of quartz from Jinmium rock shelter, northern Australia. Part II, results ad implications. Archaeometry, **41**, 365-395.

Robertson-Rintoul, M. S. E. (1986) A quantitative soil-stratigraphic approach to the correlation and dating of post-glacial river terraces in Glen Feshie, Western Cairngorms. Earth Surface Processes and Landforms, **11**, 605-617.

Rohling, E. J. and Pälike, H. (2005)Centennial-scale climate cooling with a sudden cold event around 8,200 years ago. Nature, **434**, 975-979.

Rose, J. (1985) The Dimlington Stadial/Dimlington Chronozone: a proposal for naming the main glacial episode of the Late Devensian in Britain. Boreas, **14**, 225-230.

Rose, J., Turner, J., Coope, G. R. and Bryan, M. D. (1980) Channel changes in a lowland river catchment over the last 13,000 years. In: Cullingford, R. A., Davidson, D. A. and Lewin, J. (eds) **Timescales in Geomorphology**. Chichester, Wiley & Sons, pp.159-175.

Rumsby, B. T. (1991) Flood frequency and magnitude estimates based on valley floor morphology and floodplain sedimentary sequences: the Tyne Basin, N.E. England. Unpublished PhD, University of Newcastle upon Tyne.

Rumsby, B. T. and Macklin, M. G. (1994) Channel and floodplain response to recent abrupt climate change: the Tyne Basin northern England. Earth Surface Processes and Landforms, **191**, 499-515.

Rumsby, B. T. and Macklin, M. G. (1996) River response to the last neoglacial (the "Little Ice Age") in northern western and central Europe. In: Branson, J., Brown, A. G. and Gregory, K. (eds) **Global Continental Changes: the Context of Palaeohydrology**. London, Geological Society Special Publication 115, pp.217-233.

Russell, A. J., Knudsen, O., Fay, H., Marren, P. M., Heinz, J. and Tronicke, J. (2001) Morphology and sedimentology of a giant supraglacial ice-walled, jokulhlaup channel, Skeioararjokull, Iceland: implications for esker genesis. Global and Planetary Change, **28**, 193-216.

Rust, B. R. (1972) Structure and process in a braided river. Sedimentology, **19**, 221-245.

Rust, B. R. (1975) Fabric and structure in glaciofluvial gravels. In: Jopling, A. V. and McDonald, B. C. (eds) **Glaciofluvial and glaciolacustrine sedimentation**. Tulsa, OK, Society of Economic Paleontologists and Mineralogists 3, pp.238-248.

Rust, B. R. (1977) Mass flow deposits in a Quaternary succession near Ottawa, Canada: diagnostic criteria for subaqueous outwash. Canadian Journal of Earth Sciences, **14**, 175-184.

Rust, B. R. (1978) Depositional models for braided alluvium. In: Miall, A. D. (ed.) **Fluvial Sedimentology**. Canadian Society of Petroleum Geologists, pp.605-625.

SALT. (2004) Prehistoric finds from Tynemouth. [Internet]. Available from: <http://historical-studies.ncl.ac.uk/SALT/> [Accessed 26 May 2004].

Schildgen, T., Dethier, D.P., Bierman, P. and Caffee, M. (2002) ^{26}Al and ^{10}Be Dating of Late Pleistocene and Holocene fill terraces: A record of fluvial deposition and incision, Colorado Front Range. Earth Surface Processes and Landforms, **27**, 773-787.

Schreve, D. C. (1997) Mammalian biostratigraphy of the later Middle Pleistocene in Britain. Unpublished Ph.D., University of London.

Schrott, L., Hufschmidt, G., Hankammer, M., Hoffmann, T. and Dikau, R. (2003) Spatial distribution of sediment storage types and quantification of valley fill deposits in an alpine basin, Reintal, Bavarian Alps, Germany. Geomorphology, **55**, 45-63.

Schumm, S. A. (1973) Geomorphic thresholds and complex response of drainage systems. In: Morisawa, M. (ed.) **Fluvial Geomorphology**. New York State Univ. Pub. Geomorph, Binghampton, pp.299-310.

Schumm, S. A. (1993) River Response to Baselevel Change: Implications for Sequence Stratigraphy. The Journal of Geology, **10**, 279-294.

Scrutton, C. T. (1995) **Northumbria Rocks and Landscape**. Leeds, Ellenbank Press. 216pp.

Seddon, M. B. and Holyoak, D. T. (1985) Evidence of sustained regional permafrost during deposition of fossiliferous Late Pleistocene river sediments at Stanton Harcouty (Ocfordshire, England). Proceedings of the Geologists' Association, **96**(1), 53-71.

Selby, K. A. and Smith, D. E. (2007) Late Devensian and Holocene relative sea-level changes on the Isle of Skye, Scotland, UK. Journal of Quaternary Science, **22**(2), 119-139.http://dx.doi.org/10.1002/jqs.1012.

Selby, K. A., Smith, D. E., Dawson, A. G. and Mighall, T. M. (2000) Late Devensian and Holocene relative sea-level and environmental changes from an isolation basin in southern Skye. Scottish Journal of Geology, **36**, 73–86.

Sharpe, D. E. (1988) Glacimarine fan deposition in the Champlain Sea. In: Gadd, N. R. (ed.) **The Late Quaternary Development of the Champlain Sea Basin**. Geological Association of Canada, pp.63-82.

Shaw, J. (1972) Sedimentation in the Ice-Contact Environment, with examples from Shropshire (England). Sedimentology, **18**, 23-62.

Shaw, J., Kvill, D. and Rains, B. (1989) Drumlins and catastrophic subglacial floods. Sedimentary Geology, **62**, 177–202.

Shennan, I. (1992) Late Quaternary sea-level changes and crustal movements in Eastern England and Eastern Scotland: an assessment of models of coastal evolution. Quaternary International, **15/16**, 161-173.

Shennan, I. (1999) Global meltwater discharge and the deglacial sea-level record from northwest Scotland. Journal of Quaternary Science, **14**, 715-719.

Shennan, I., Bradley, S., Milne, G., Brooks, A., Bassett, S. and Hamilton, S. (2006) Relative sea-level changes, glacial isostatic modelling and ice-sheet reconstructions from the British Isles since the Last Glacial Maximum. Journal of Quaternary Science, **21**(6), 585-599.

Shennan, I., Innes, J. B., Long, A. J. and Zong, Y. (1993) Late Devensian and Holocene relative sea-level changes at Rumach, near Arisaig, northwest Scotland. Norsk Geologisk Tiddskrift, 161–174.

Shennan, I., Innes, J. B., Long, A. J. and Zong, Y. (1994) Late Devensian and Holocene relative sea-level changes at Loch nan Eala, near Arisaig, northwest Scotland. Journal of Quaternary Science, **9**, 261-283.

Shennan, I., Innes, J. B., Long, A. J. and Zong, Y. (1995) Late Devensian and Holocene relative sea-level changes in Northwestern Scotland: new data to test existing models. Quaternary International, **26**, 97-123.

Shennan, I., Lambeck, K., Horton, B. P., Innes, J. B., Lloyd, J., McArthur, J. and Rutherford, M. (2000) Holocene isostasy and relative sea-level changes on the east coast of England. In: Shennan, I. and Andrews, J. (eds) **Holocene Land-Ocean Interaction and Environmental Change around the North Sea**. London, The Geological Society of London, pp.275-298.

Sissons, J. B. (1958) Sub-glacial Stream Erosion in Southern Northumberland. The Scottish Geographical Magazine, **74**, 163-174.

Sissons, J. B. (1960) Some Aspects of Glacial Drainage Channels in Britain Part I. The Scottish Geographical Magazine, **76**(3), 131-146.

Sissons, J. B. (1961) Some Aspects of Glacial Drainage Channels in Britain Part II. The Scottish Geographical Magazine, **77**, 15-36.

Sissons, J. B. (1964) The Glacial Period. In: Watson, J. W. and Sissons, J. B. (eds) **The British Isles: a systematic geography**. Edinburgh, Nelson and Sons, pp.131-149.

Sissons, J. B. (1967) **The evolution of Scotland's scenery**. Edinburgh, Oliver and Boyd. 259pp.

Sissons, J. B. (1976) **The Geomorphology of the British Isles: Scotland**. London, Methuen.

Sissons, J. B. (1979) The later lakes and associated fluvial terraces of Glen Roy, Glen Spean and vicinity. Transactions of the Institute of British Geographers, **4**, 12-29.

Sissons, J. B. (1981) A former ice-dammed lake and associated glacier limits in the Achnasheen area, central Ross-shire. Transactions of the Institute of British Geographers, **N.S. 7**, 98-116.

Sissons, J. B. and Cornish, R. (1983) Fluvial landforms associated with ice-dammed lake drainage in upper Glen Roy, Scotland. Proceedings of the Geologists' Association, **94**(1), 45-52.

Smith, B. W. and Rhodes, E. J. (1994) Charge movements in quartz and their relevance to optical dating. Radiation Measurements, **23**, 329-333.

Smith, D. B. (1981) The Quaternary Geology of the Sunderland District, North East England. In: Neale, J. and Flenley, J. (eds) **The Quaternary in Britain**. Oxford, Pergamon Press, pp.146-167.

Smith, D. B. (1994) **Geology of the district around Sunderland, sheet 21**. Memoir of the British Geological Survey. London, HMSO.

Smith, D. B. and Francis, E. A. (1967) **Geology of the Country between Durham and West Hartlepool**. Explanation of one-inch Geological Sheet 27, new series. London, HMSO. 354pp.

Smith, M. J. and Clark, C. D. (2005) Methods for visualization of digital elevation models for landform mapping. Earth Surface Processes and Landforms, **30**, 885-900.

Smith, M. J., Rose, J. and Booth, S. (2006) Geomorphological mapping of glacial landforms from remotely sensed data: An evaluation of the principal data sources and an assessment of their quality. Geomorphology, **76**, 148-165.

Smith, N. D. (1970) The braided stream depositional environment; comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians. Geological Society of America Bulletin, **91**, 2993-3014.

Smith, N. D. (1974) Sedimentology and bar formation in the Upper Kicking Horse River, a braided outwash stream. Journal of Geology, **82**, 205-223.

Smith, N. D. (1985) Proglacial Fluvial Environment. In: Ashley, G. M., Shaw, J. and Smith, N. D. (eds) **Glacial Sedimentary Environments SEPM Short Course No. 16**. Tulsa, Oklahoma, Society of Economic Palaentologists and Mineralogists Special Publication 23, pp.85-134.

Smith, N. D. and Ashley, G. M. (1985) Proglacial Lacustrine Environment. In: Ashley, G. M., Shaw, J. and Smith, N. D. (eds) **Glacial Sedimentary Environments SEPM Short Course No. 16**. Tulsa, Oklahoma, Society of Economic Paleontologists and Mineralogists Special Publication 23, pp.135-215.

Smith, S. A. (1990) The sedimentology and accretionary styles of an acient gravel bed stream: the Budleigh Salterton Pebble beds (Lower Triassic), southwest England. Sedimentary Geology, **67**, 199–219.

Smythe, J. A. (1907) The Glacial Phenomena of the Country between the Tyne and the Wansbeck. Transactions of the Natural History Society of Northumberland, Durham and Newcastle upon Tyne, **N.S. 8**, 79-109.

Smythe, J. A. (1912) The Glacial Geology of Northumberland. Transactions of the Natural History Society of Northumberland, Durham and Newcastle upon Tyne, **NS 4**, 86-116.

Spencer, J. Q. and Owen, L. A. (2004) Optically stimulated luminescence dating of Late Quaternary glaciogenic sediments in the upper Hunza valley: validating the timing of glaciation and assessing dating methods. Quaternary Science Reviews, **23**(1-2), 175-191.

Spooner, I. S. and Osborn, G. D. (2000) Geomorphology and Late Wisconsinan sedimentation in the Stikine River Valley, northern British Columbia. Quaternary International, **68-71**, 285-296.

Stein, M. L. (1988) Asymtotically efficient prediction of a random field with a misspecified covariance function. Annals of Statistics, **16**, 55-63.

Stokes, C. R. and Clark, C. D. (2001) Palaeo-ice streams. Quaternary Science Reviews, **20**(13), 1437-1457.

Stokes, C. R. and Clark, C. D. (2003) The Dubawnt Lake palaeo-ice stream: evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding the controls on ice-stream location and vigour. Boreas, **32**(1), 263-279.

Stokes, S. (1999) Luminescence dating applications in geomorphological research. Geomorphology, (29), 153 – 171.

Stokes, S., Bray, H. E. and Blum, M. D. (2001) Optical resetting in large drainage basins: tests of zeroing assumptions using single-aliquot procedures. Quaternary Science Reviews, **20**, 879-885.

Strickertsson, K., Murray, A. S. and Lykke-Andersen, H. (2001) Optically stimulated luminescence dates for Late Pleistocene sediments from Stensnaes, Northern Jutland, Denmark. Quaternary Science Reviews, **20**, 755-759.

Sugden, D. E. (1970) Landforms of deglaciation in the Cairngorms, Scotland. Transactions of the Institute of British Geographers, **51**, 201-219.

Sutherland, D. G. (1984) The Quaternary deposits and landforms of Scotland and the neighbouring shelves: a review. Quaternary science reviews, **3**(2-3), 157-254.

Syvitski, J. P. M. (1991) **Principles, Methods and Application of Particle Size Analysis**. Cambridge, New York, Cambridge University Press. 368pp.

Tatalovich, Z. (2005) A comparison of Thiessen-polygon, Kriging and Spline Models of UV exposure. **University Consortium of Geographical Information Science** [Internet]. Available from: <http://www.ucgis.org/summer2005/studentpapers/tatalovich.pdf> [Accessed 16 October 2006].

Taylor, B. J., Burgess, I. C., Land, D. H., Mills, D. A. C., Smith, D. B. and Warren, P. T. (1971) **British Regional Geology: Northern England**. 4th Ed. London, HMSO.

Taylor, M. P. and Lewin, J. (1996) River behaviour and Holocene alluvial sedimentation: the River Severn at Welshpool, mid-Wales, UK. Earth Surface Processes and Landforms, **21**, 77-91.

Taylor, M. P. and Lewin, J. (1997) Non-synchronous response of adjacent upland floodplain systems to Holocene environmental change. Geomorphology, **18**, 251-264.

Taylor, M. P. and Macklin, M. G. (1997) Holocene alluvial sedimentation and valley floor development: the River Swale, Catterick, North Yorkshire, UK. Proceedings of the Yorkshire Geological Society, **51**, 317-327.

Taylor, M. P., Macklin, M. G. and Hudson-Edwards, K. A. (2000) River sedimentation and fluvial response to Holocene environmental change in the Yorkshire Ouse Basin, northern England. The Holocene, **10**(2), 201-212.

Teasdale, D. and Hughes, D. (1999) The Glacial history of North East England. In: Bridgland, D. R., Horton, B. P. and Innes, J. B. (eds) **The Quaternary of North East England**. Cambridge, Quaternary Research Association, pp.10-17.

Thomas, E. R., Wolff, E. W., Mulvaney, R., Steffensen, J. P., Johnsen, S. J., Arrowsmith, C., White, J. W. C., Vaughn, B. and Popp, T. (2007) The 8.2 ka event from Greenland ice cores. Quaternary Science Reviews, **26**(1-2), 70-81.

Thomas, G. S. P. (1984) A Late Devensian glaciolacustrine fan-delta at Rhosesmor, Clywd, North Wales. Geological Journal, **19**, 125-141.

Thomas, G. S. P. (1985) The Quaternary of the northern Irish Sea Basin. In: Johnson, R. H. (ed.) **The Geomorphology of North West England**. Manchester, Manchester University Press.

Thomas, G. S. P. (1989) The late Devensian glaciation along the western margin of the Cheshire-Shropshire lowland. JQS. Journal of quaternary science, **4**(2), 167-181.

Thomas, G. S. P., Chiverrell, R. C. and Huddart, D. (2004) Ice-marginal depositional responses to readvance episodes in the Late Devensian deglaciation of the Isle of Man. Quaternary Science Reviews, **23**(1-2), 85-106.

Thomas, G. S. P. and Chiverrell, R. C. (2007) Structural and depositional evidence of repeated ice-marginal oscillation along the eastern margin of the Late Devensian Irish Sea Ice Stream. Quaternary Science Reviews, **25**, 2375-2405.

Thomas, G. S. P., Connaughton, M. and Dackombe, R. V. (1985) Facies variation in a Late Pleistocene supraglacial outwash sandur from the Isle of Man. Geological Journal, **20**, 193-213.

Thomas, G. S. P. and Montague, E. (1997) The morphology, stratigraphy and sedimentology of the Carstairs Esker, Scotland, UK. Quaternary Science Reviews, **16**, 661-674.

Thomas, G. S. P., Summers, A. J. and Dackombe, R. V. (1982) The Late-Quaternary deposits of the middle Dyfi Valley, Wales. Geological Journal, **17**, 297-309.

Thrasher, I. M., Chiverrell, R. C., Thomas, G. S. P., Mauz, B. and Lang, A. (2007) **A lithofacies based approach to optically dated luminescencee (OSL) dating of glaciofluvial sediments: abstract**. **QRA Annual Discussion Meeting: The Growth, Maximum Extent and Deglaciation of the Last British Ice Sheet**, **4-5 January 2007**, St Andrews.

Tipping, R. M. (1992) The determiniation of cause in the generation of major prehistoric valley fills in the Cheviot Hills, Anglo-Scottish border. In: Needham, S. and Macklin, M. G. (eds) **Alluvial archaeology in Britain**. Oxford, Oxbow, pp.111-122.

Tipping, R. M. (1994) Fluvial Chronology and Valley Floor Evolution of the Upper Bowmont Valley, Borders Region, Scotland. Earth Surface Processes and Landforms, **19**(7), 641-657.

Tipping, R. M. (1995) Holocene Evolution of a Lowland Scottish Landscape - Kirkpatrick-Fleming.3. Fluvial History. Holocene, **5**(2), 184-195.

Tipping, R. M. (1998) The chronology of late Quaternary fluvial activity in part of the Milfield basin, north-east England. Earth Surface Processes and Landforms, **23**, 845-856.

Tipping, R. M. (1999) Fluvial Processes, Land Use and Climate Change 2000 Years Ago in Upper Annandale, Southern Scotland. In: Brown, A. G. and Quine, T. A. (eds) **Fluvial Processes and Environmental Change**. Chichester, Wiley, pp.311-327.

Tipping, R.M., Jones, A. P., Carter, S., Holden, T. and Cressey, M. (2008) The chronology and long term dynamics of a low energy river system: the Kelvin Valley, central Scotland. Earth Surface Processes and Landforms, **33**(6), 910-922.

Törnqvist, T. E. (1998) Longitudinal profile evolution of the Rhine-Meuse system during the last deglaciation: interplay of climate change and glacio-eustasy? Terra Nova, **10**(1), 11-15.

Toutin, T. and Rivard, B. (1995) A new tool for depth perception of multi-source data. Photogrammetric engineering and remote sensing, **61**(10), 1209-1211.

Tucker, M. E. (2003) **Sedimentary Rocks in the Field**. The Geological Field Guide Series, 3rd. Chichester, Wiley. 234pp.

Turner, J. and Kershaw, A. P. (1973) A late- and post-glacial pollen diagram from Cranberry bog, near Beamish, County Durham. New Phytologist, **72**, 915-928.

Van Den Eeckhaut, M., Poesen, J., Verstraeten, G., Vanacker, V., Moeyersons, J., Nyssen, J. and van Beek, L. P. H. (2005) The effectiveness of hillshade maps and expert knowledge in mapping old deep-seated landslides. Geomorphology, **67**, 351-363.

Van der Meer, J. J. M. and Warren, W. P. (1997) Sedimentology of Lateglacial clays in lacustrine basins, central Ireland. Quaternary Science Reviews, **16**, 779-791.

Vandenberghe, J. (1995) Timescales, climate and river development. Quaternary Science Reviews, **14**, 631-638.

Vandenberghe, J. (2002) The relation between climate and river processes, landforms and deposits during the Quaternary. Quaternary International, **91**, 17-23.

Vandenberghe, J. and Maddy, D. (2001) The response of river systems to climate change. Quaternary International, **79**(1), 1-3.

Wackernagel, H. (2003) **Multivariate geostatistics: an introduction with applications**. 3rd edition. Berlin, Springer-Verlag.

Wallinga, J. (2002) On the detection of OSL age overestimation using single-aliquot techniques. Geochronometria, **21**, 17-26.

Ward, B. and Rutter, N. (2000) Deglacial valley fill sedimentation, Pelly river, Yukon territory, Canada. Quaternary International, **68-71**, 309-328.

Ward, B. C., Bond, J. D. and Gosse, J. C. (2007) Evidence for a 55-50 ka (early Wisconsin) glaciation of the Cordilleran ice sheet, Yukon Territory, Canada. Quaternary Research, **68**(1), 141.http://www.sciencedirect.com/science/article/B6WPN-4NSV161-1/2/6587c0dc2930ff40d6439cd0d04059cb.

Ward, B. C. and Thomson, B. (2004) Late Pleistocene stratigraphy and chronology of lower Chehalis River valley, southwestern British Columbia: evidence for a restricted Coquitlam Stade. Canadian Journal of Earth Sciences, **41**(7), 881-895.

Warren, P. W. and Ashley, G. M. (1994) Origins of ice-contact stratified ridges (eskers) of Ireland. Journal of Sedimentary Research, **A64**, 433-449.

West, R. G. (1963) Problems of the British Quaternary. Proceedings of the Geologists' Association, **72**(2), 147-183.

West, R. G. (1967) The Quaternary of the British Isles. In: Rankama, K. (ed.) **The Quaternary. Vol. 2. (Britain, France, Germany and The Netherlands)**. London, Interscience, pp.1-87.

Whiting, P. J., Dietrich, W. E., Leopold, L. B., Drake, T. G. and Shreve, R. L. (1988) Bedload sheets in heterogeneous sediment. Geology, **16**, 105-108.

Williams, G. P. (1983) Paleohydrological methods and some examples from Swedish fluvial environments. Geografiska Annaler, **65A**, 243-277.

Williams, P. F. and Rust, B. R. (1969) The sedimentology of a braided river. Journal of Sedimentary Research, **39**(2), 649-679.

Wintle, A. G. and Murray, A. S. (2006) A review of quartz optically stimulated luminescence characteristics and their relevance in single-aliquot regeneration dating protocols. Radiation Measurements, **41**, 369-391.

Wolman, M. G. and Miller, W. P. (1960) Magnitude and frequency of forces in geomorphic processes. Journal of Geology, **68**, 54-74.

Wood, N. and Boyd, E. F. (1863) On a Wash or Drift through a Portion of the Coal-Field of Durham. Trans. North of Eng, Inst. Min. Eng, 69-85.

Woolacott, D. (1905) The Superficial Deposits and Pre-Glacial Valleys of the Northumberland and Durham Coalfield. Quaterley Journal of the Geological Society of London, **61**, 64-96.

Worsley, P. (1996) On ice-wedge casts and soft sediment deformations. Quaternary Newsletter, **78**, 1-7.

Wymer, J. (1968) **Lower Palaeolithic Archaeology in Britain as Represented by the Thames Valley**. J. Baker.

Young, B., Lawrence, D. J. D. and Woodhall, D. G. (2002) **Geology of the Morpeth district: a brief explanation of the geological map sheet 14 Morpeth (2002)**. London, HMSO.

Young, J. A. T. (1976) The terraces of Glen Feshie, Inverness-shire. Transactions of the Royal Society of Edinburgh, **69**, 501-512.

Zieliński, T. and Goździk, J. (2001) Palaeoenvironmental interpretation of a Pleistocene alluvial succession in central Poland: sedimentary facies analysis as a tool for palaeoclimatic inferences. Boreas, **30**(3), 240-253.

Zuuring, H. R. (2002) **Introduction to Spatial Interpolation**. **Intermountain GIS Users Conference**, **April 9-11, 2002**, Big Sky, MT.

Zuuring, H. R. (2004) GIS: Methods and Applications I Notes - Part 7. **The University of Montana** [Internet]. Available from: <http://www.forestry.umt.edu/academics/courses/for503/spatial_stats.htm> [Accessed 16 October 2006].

Guide to Appendices

The appendices to this thesis are located on the CD held in the pocket inside the back cover. The guide to the three appendixes are listed below:

Appendix One: Map and GIS database of the landform-sediment assemblages and features related to deglaciation and post-glacial development of the Tyne Valley.

Appendix Two: Full sediment logs and descriptions.

Appendix Three: Laboratory report detailing the methods and results of OSL dating.