The glacial geomorphology of the tweed valley and surrounding area, eastern British Isles.

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THE GLACIAL GEOMORPHOLOGY OF THE TWEED VALLEY AND SURROUNDING AREA, EASTERN BRITISH ISLES

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Geography MSc by Research

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28,400 words

DECLARATION

This thesis is the result of my own work and has not been submitted for consideration in any other examination. Material from the published or unpublished work of other, which is referred to in the thesis, is credited to the author in question in the text.

Kate Staines

May 2009

ABSTRACT

This project focuses on reconstructing the glacial dynamics of the Tweed Valley and surrounding area in the Scottish Borders and northeast Northumberland from the glacial geomorphological record. Previous investigations in the region are heavily descriptive and there has been little focus on reconstructing the regional ice configuration, dynamics and style of retreat of this sector of the British Ice Sheet. Geomorphological mapping from digital elevation models conducted in this study has identified three main landform assemblages. Landform Assemblage A is comprised of highly attenuated longitudinal subglacial bedforms, situated in an arc to the north of the Cheviot Massif. Landform Assemblage B consists of glaciofluvial and glaciolacustrine complexes found at the northeast edge of the Cheviot Massif and on the North Northumberland Coastal Plain. Landform Assemblage C is comprised of meltwater channels cut into the lower flanks of the Cheviot and Lammermuir Hills. The geomorphological data was supplemented by sedimentological surveys at selected sites.

From the geomorphological and sedimentological characteristics of these landforms, and also their spatial relationship with each other, inferences have been made on regional glacial dynamics. The organisation of meltwater channels, streamlined bedforms and eskers have been used to reconstruct regional ice flow trajectories. It has been shown that a polythermal ice sheet existed, formed of a warm-based, fast-flowing Tweed Ice Stream that was coalescent with the cold-based Cheviot Ice Cap. It is proposed that the Tweed Ice Stream was diverted southwards along the North Sea coast by the North Sea Lobe during the Late Devensian. During deglaciation, which is inferred to have been relatively rapid, the Bradford Interlobate Complex formed in the suture zone between these two ice masses. Following ice streaming, the Cornihill-Wooler Glaciofluvial Complex formed and meltwater channels were cut as the Tweed Ice Stream and Cheviot Ice Cap separated along the ice stream lateral shear margin. This study has revealed that this sector of the British Ice Sheet was more dynamic than originally thought and has highlighted the importance of the geomorphological record for ice sheet reconstructions.

ACKNOWLEDGEMENTS

I wish to thank Professor Colm Ó Cofaigh and Dr David Evans for their guidance, support and tireless encouragement throughout the year. Thanks also go to Dr Chris Stokes for his invaluable advice and help with GIS and Dr David Bridgland for assisting with sedimentology. I am particularly grateful to Vicky Brown, Ashley Woods and Nat Young for help with field work, all the landowners who kindly allowed me access to their farmland and the Wooler YHA staff for the generous loan of maps. Thanks also to the laboratory staff and technicians in the Department of Geography, particularly Neil Tunstall, Kathryn Melvin, Martin West, Frank Davies and Mervyn Brown.

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INTRODUCTION

This thesis aims to reconstruct the glacial history and regional ice dynamics of the Tweed Valley and surrounding area from the glacial geomorphological record. Through combining geomorphological mapping and sedimentological surveys, the glacial landform assemblages of this sector of the British Ice Sheet have been identified. The rationale for undertaking this research is outlined in this chapter.

1.1 Background: The importance of the geomorphologic record to ice sheet research

Large ice sheets have the capacity to dramatically alter (and be altered by) climatic, atmospheric and oceanic systems (Siegert, 2001), as well as extensively and intensively modifying the land over which they flow. The growth and decay of the vast ice sheets over Europe, North America, Greenland and Antarctica during the Quaternary brought changes in global sea levels (c.f. Bell *et al.*, 1998; Dyke *et al.*, 2002), oceanic chemistry and circulation (c.f. Clark *et al.*, 2000; Alley *et al.*, 2005), atmospheric circulation and regional and global climates (c.f. Bigg *et al.*, 2003). It is therefore essential to understand processes occurring within ice sheets and the mechanisms controlling ice sheet dynamics and stability. It is, however, remarkably difficult and hazardous to directly observe contemporary ice sheets (Schoof 2004), and consequently, little is understood about their internal organisation (De Angelis and Kleman 2007). In particular, little is understood about contemporary subglacial environments owing to the difficulty in accessing the bed (Stokes and Clark, 2001; Schoof, 2004). Much of what is understood about both contemporary and palaeo-ice sheets is obtained from the geomorphological record, which is invaluable to glaciology. Information that is unattainable at contemporary glaciers (for example, data on subglacial processes) can be provided through the study of the geomorphological imprint left by palaeo-ice sheets (Hubbard and Glasser, 2005). The distribution and organisation of landforms in formerly glaciated areas provides information on ice sheet dimensions, internal configuration, flow directions and velocities, basal conditions and glacial histories (Glasser and Bennett, 2004). The geomorphological imprint left by Pleistocene ice sheets is therefore invaluable to glaciology, as information on glacial processes that is unattainable at modern glaciers, can be provided (Hubbard and Glasser, 2005). Geomorphological criteria used in the identification of palaeo-ice sheet limits include moraines, drift limits, trimlines (e.g. McCarroll and Ballantyne, 2000; Ballantyne *et al.*, 2006) and glaciolacustrine and glaciofluvial deposits (e.g. Clark *et al*, 2004). Flow trajectories can be reconstructed from erratics (e.g. Sissons, 1967) streamlined bedforms (e.g. Stokes and Clark, 2001; Briner, 2007; Stokes *et al*, 2006), striae and megagrooves (e.g. Bradwell *et al*, 2008a) and clast macrofabrics (e.g. Benn, 1995; Hart and Rose, 2001). Features such as meltwater channels and differential zones of erosion allow inferences to be made on ice thicknesses and the thermal regime of an ice sheet (e.g. Knight 2002; Hall and Glasser, 2003; Kleman and Glasser, 2007). Sedimentological data provides information on glacigenic depositional environments (e.g. Brodzikowski and van Loon, 1991; Davis *et al.*, 2006), glacial thermal regime, and subglacial processes (e.g. Mokhtari Fard and Gruszka, 2007).

An example of the value of the geomorphological record for ice sheet reconstructions has been the identification of palaeo-ice stream tracks. The identification of palaeo-ice streams within formerly glaciated regions yields information on the configuration and thermal regime of past ice sheets (Stokes *et al*, 2006) and can have far-reaching implications for not only palaeo-environmental reconstructions, but also for our understanding of contemporary ice streams. Ice streams currently drain large sectors of the Greenland and Antarctic Ice Sheets; indeed, discharge from the marine-based West Antarctic Ice Sheet (WAIS) is dominated by flow from ice streams (Jacobel *et al.*, 1996). It is argued the most important controls on ice sheet mass balance and stability are the configuration and dynamics of its ice streams (Stokes and Clark, 2001; Stokes *et al.*, 2006; Stokes et al., 2007). A collapse of the WAIS ice sheet would equate approximately to a sea level rise of 6 m (Bell et al., 1998). For these reasons, there has recently been increasing scientific interest in ice streams and palaeo-ice stream reconstructions.

Palaeo-ice streams of the Laurentide Ice Sheet (LIS) have been hypothesised at several locations, such as around the Hudson Strait (Andrews *et al.,* 1985; Boulton and Clark 1990; Laymon, 1992), at Victoria Island and the Amundsen Gulf (Hodgson, 1994; Stokes *et al.*, 2006) and Cumberland Sound (Kaplan *et al.*, 1999). In these regions, ice streaming has been inferred from several lines of geomorphological evidence, including attenuated streamlined bedforms and megascale-glacial lineations (MSGL) and the presence of ice-stream lateral shear margin moraines (e.g. Dyke and Morris, 1988; Stokes and Clark, 2002). Key to the reconstruction of former ice streams has been the detection of 'flow-sets' within the geomorphological record (c.f. Clark, 1999). Ice flow indicators are grouped into flow-sets on the basis of their orientation and parallel concordance, proximity to surrounding features and morphology (Stokes *et al*, 2006). The identification of cross-cutting landforms and superimposed flow-sets has been used to reconstruct the temporal changes in the internal configuration of ice sheets (e.g. Stokes *et al.*, 2006; Evans *et al.*, 2009). Ice streams have also been hypothesised in several locations relating to the British Ice Sheet (BIS) and have been used to suggest that the BIS was more dynamic than had previously been thought (e.g. Golledge and Stoker, 2006; Everest *et al.*, 2005) (see below in section 1.2). However, despite these recent advances, our understanding of the controls on ice streaming, ice stream configuration and the processes occurring at the ice stream bed and lateral margins remains limited. Consequently, there are still large gaps in our knowledge of the configuration and behaviour of palaeo-ice sheets.

1.2 The British Ice Sheet (BIS)

Various lines of evidence have been used to reconstruct the BIS. Models of the BIS vary greatly in terms of their inputs and outputs and consequently, there is little agreement on the maximum extent, configuration and thickness of ice cover and the occurrence and timing of advance and retreat phases (Fretwell *et al.*, 2008). In terms of ice thickness, Boulton *et al.* (1977) proposed that the BIS was at its greatest height over central Scotland, where it reached 1800 m. In this model, the BIS is inferred to have been composed of several coalescing ice domes over the Grampians, central Ireland and the Southern Uplands, the Lake District, the Pennines and Wales (Boulton *et al.*, 1977). Alternatively, Denton and Hughes (1981) suggested that a single dome of ice existed over central Scotland, with ice no thicker than 1750 m thick. More recently, geomorphological evidence in the form of trimlines and nunataks in Wales, southern Ireland and the Lake District (e.g. McCarroll and Ballantyne, 2000; Ballantyne *et al.*, 2006) have been used to infer ice thickness. The results suggest that the ice was thinner than is inferred in models that use solely theoretical models of ice dynamics, which do not often take into account the underlying topography (Fretwell *et al.* 2008). Since many of these models have been used to estimate ice volumes, glacio-isostatic adjustment rates and relative sea-level change, the inclusion of a topographic parameter within ice sheet models is essential. Indeed, Fretwell *et al.* (2008: 241) state that where topography has been ignored, such models "seriously overestimate ice volume". As with ice thicknesses, there has been little agreement on the lateral extent and timing of the BIS, in large part due to a lack of dateable deposits in lowland areas (Evans *et al.* 2005). Geomorphological evidence used to infer the ice limits include offshore glaciomarine deposits and tills (Sejrup *et al.* 1994; Hall *et al.* 2003) and onshore glaciolacustrine deposits and proglacial lake levels (Gaunt 1981). The BRITICE mapping project collated these numerous lines of evidence to produce the Glacial Map of Britain (Clark *et al.* 2004). The inferred limits of the Devensian ice sheet are shown in Figure 1.

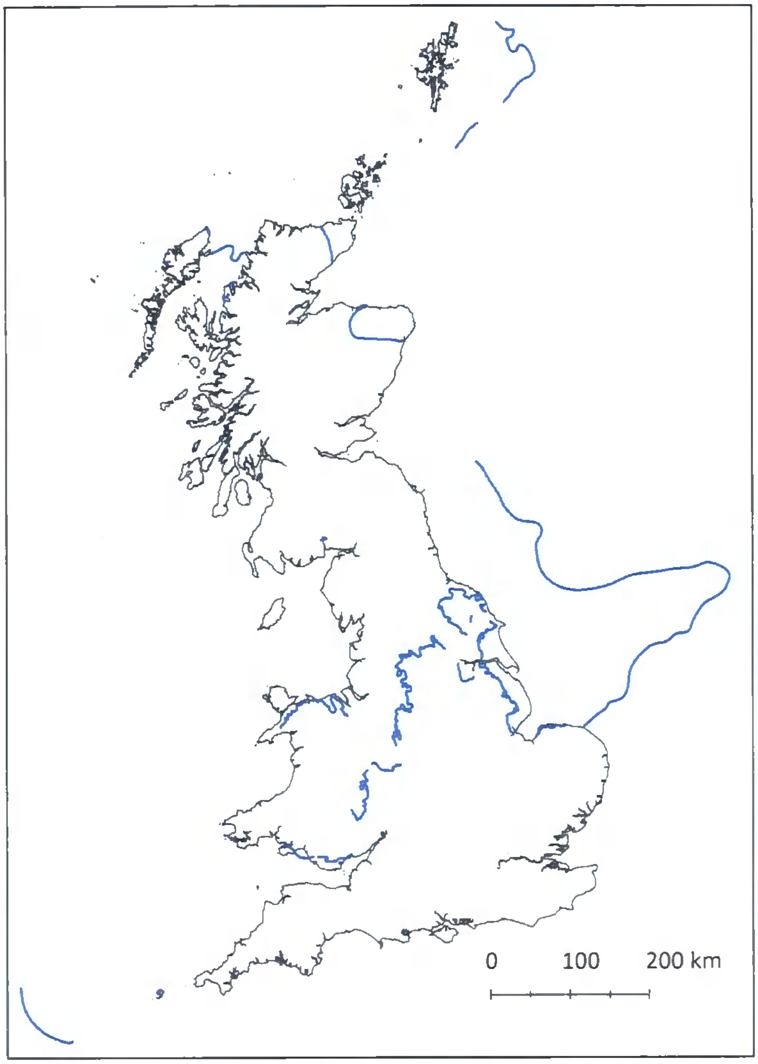


Figure 1: *The maximum inferred limit of the Devensian Ice Sheet in Britain (shown in blue). From the Glacial Map of Britain (Clark et al., 2004).*

More recent glaciological models are more complex and attempt to verify model findings against the glacial geomorphological and geological record (e.g. Boulton and Hagdorn, 2006; Brooks *et al*., 2007; Hubbard *et al*, 2009; Evans *et al*., 2009). These studies have revealed an extremely dynamic ice sheet with continuously migrating ice dispersal centres (Evans *et al*., 2009) and ice streams that "switch and fluctuate in extent and intensity" (Hubbard *et al*, 2009:758). Despite these recent advances, relatively little is still understood about the internal configuration of the BIS; there have been few investigations that center on identifying palaeo-ice stream signatures (Everest et al. 2005). Given the importance of ice streams in determining the stability of ice sheets (see section 1.1), there is a great need to identify zones of ice streaming within the BIS in order to understand in more detail the Devensian ice dynamics (Evans *et al*. 2005). Ice streams of the BIS that have been identified (see Fig. 2, page 5) include the Strathmore Ice Stream, that flowed offshore from the Firth of Tay (Golledge and Stoker

2006) and the Moray Firth Ice Stream (Merritt *et al.* 1995). Ice streaming is also inferred to have occurred in the Irish Sea Basin (Evans and Ó Cofaigh 2003; Thomas and Chiverrell 2007; Roberts *et al.*, 2007), in the Vale of York and Vale of Eden (Evans *et al.*, 2005) and in the Tweed Valley (Everest *et al.*, 2005).

Figure 2: *Location and names of ice streams of the BIS (from Everest et al., 2005).*

It has been highlighted in the BRITICE Glacial Map of Britain project (Clark *et al.*, 2004) that there are great spatial variations in the both the volume and quality of geomorphological data relating to the BIS. Furthermore, in areas where studies have been carried out, the resulting ice sheet reconstructions are highly varied due to the variety in methods utilised (Evans *et al.*, 2005). Evans *et al.* (2005) state that in order to produce more confident and accurate reconstructions of the BIS, detailed geomorphological surveys must be undertaken in regions where there has been little research into the glacial history. One such region is the northeast of England, in particular the Tweed Valley and northeast Northumberland. A few studies have aimed to reconstruct the activity of the BIS along the northeast coast (e.g. Carr *et al.*, 2006; Davies *et al.*, 2009), however this region remains little understood. Few dates exist for the last glacial maximum (LGM), the onset of deglaciation or the Younger Dryas (YD) (Harrison et al., 2006). Much of what is understood of the glacial history in northeast England is based on morpho-stratigraphic correlations with upland sites elsewhere in Britain (Lunn, 1980). The glacial sedimentary and geomorphic evidence throughout this region is highly fragmented (Lunn, 1980, 1995) and consequently, the Quaternary record in the north-east remains relatively incomplete. Compared to other upland regions of Britain, the Cheviots to the south of the Tweed Valley have remained relatively neglected in terms of glaciological research. Whilst detailed geomorphological and sedimentological studies have been carried at sites in this region (e.g. Gunn and Clough, 1895; Carruthers, *et al.*, 1932; Price, 1960; Clapperton, 1968, 1970, 1971a, 1971b), there has been little emphasis on reconstructing the regional ice dynamics from these records. It has been proposed that on at least two occasions during the Quaternary ice completely covered north-east England (Teasdale and Hughes, 1999); during the Anglian (MIS 12) and the Devensian (MIS2-4d) (Lunn, 1995). Little is known about the pre-late Devensian; much of the sedimentary evidence was eroded and modified by late-Devensian glaciations (Lunn, 1980; Huddart, 2002a). Whilst more is known about the late-Devensian than earlier glacials and interglacials, there are still large gaps in both the record and our knowledge of the glacial history of this region. What is known about the Devensian glacial history of northeast Northumberland and the Tweed Valley is outlined in section 2.1 (page 11).

1.3 Aims and Objectives

The main aim of this project is to investigate the glacial geomorphology of the Tweed Valley and surrounding area with a view to reconstructing the glacial history of the region. As mentioned above, detailed investigations of individual landform groups have previously been carried out (c.f. Gunn and Clough, 1895; Carruthers, *et al.*, 1932; Clapperton, 1968), however, there has been little emphasis on reconstructing the regional ice dynamics. In order to address the aim, four broad objectives have been set:

(1) Review the current literature on the field site (see chapter 2, page 11);
(2) Map and describe in detail the geomorphology of the Tweed Valley and surrounding areas;
(3) Conduct sedimentological surveys at all available exposures to supplement the geomorphological data;
(4) Reconstruct the ice dynamics of the region through the interpretation of the geomorphological and sedimentological evidence.

It is proposed that by combining the geomorphological record of this sector of the BIS with information on glacial processes observed in contemporary glacial settings, accurate interpretations of this little-studied region can be made (c.f. Evans and Twigg, 2002). This will lead to a wider understanding of the ice dynamics and configuration of the BIS, which, as mentioned above, are highly contested.

1.4 Physiology and Pre-Quaternary Geology of Study Site

The chosen field site covers approximately 3800 km^2 of northeast Northumberland and the southeast Scottish Borders (Fig. 3). The study site incorporates the Tweed Valley and Lammermuir Hills to the north, the North Northumberland Coastal Plain (NNCP) to the east and, in the centre, the most north-easterly section of the Cheviot Hills, the Cheviot massif. The Cheviot massif covers an area of around 625 km^2 and is characterised by its distinctive topography and geology (Clapperton, 1970b). The highest point is The Cheviot at 815 m OD. Steeply incised stream and river valleys (Fig. 4, page 8) extend radially from the plateau-like summits to meet the lowest slopes of the Cheviots, which fall sharply towards the Tweed and Till river plains to the north and the River Coquet to the south (Robson, 1966). Rivers draining

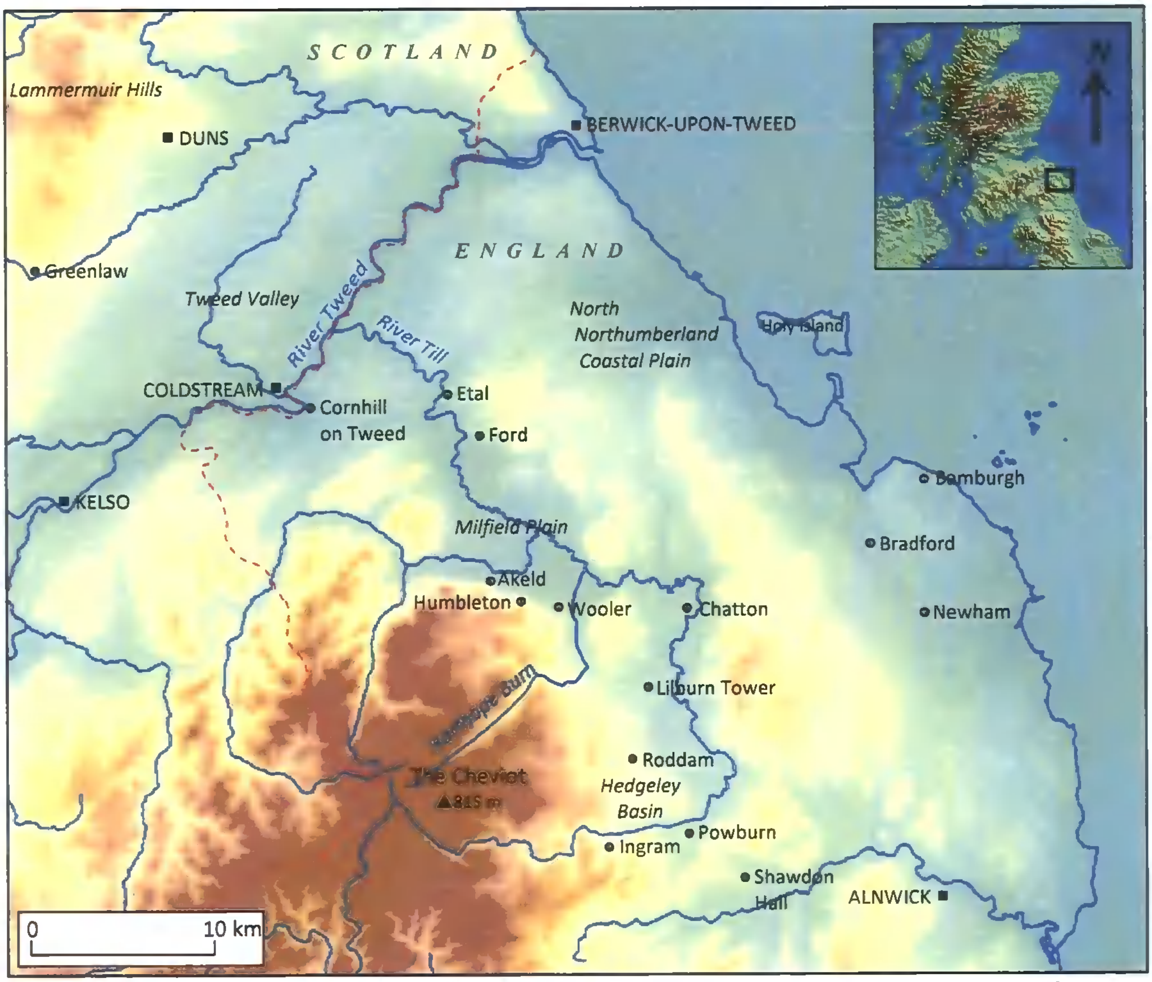


Figure 3: Main locations mentioned in text and topography of study site. Basemap relief-shaded NEXTmap® DEM (NEXTmap® Britain, Intermap Technologies 2008). Inset map shows field site location within the north of the British Isles.

the north and northeast Cheviots flow predominantly towards the Milfield Plain at the northeast edge of the Cheviot Hills. The River Breamish flows from the east of the Cheviots, converging with Wooler Water and the River Till on the Milfield Plain, just north of the town of Wooler. From here the River Till flows northwest towards the River Tweed. Towards the southeast of the field site the River Aln flows easterly towards Alnmouth on the coast. The River Tweed has its source in the Southern Uplands to the west of the field site. From here it flows east between the higher land of the Lammermuir Hills to the North and the Cheviot Hills to the south to discharge into the North Sea at Berwick-upon-Tweed. South of Berwick-upon-Tweed stretches the NNCP, the coastline of which is characterised by sand dunes and beaches with mudflats, sandflats and saltmarshes (Horton *et al.*, 1999).

Figure 4: *View SE towards Harthope Burn, southwest of Wooler. For location see Figure 3. Photo: KEHS*

In terms of the geology of the region (Fig. 5, page 9), the high ground of the Cheviot Massif is composed of granite (shown in bright blue in Fig. 5, page 9) of Old Red Sandstone (Devonian) age that was intruded into an andesitic and basaltic lava dome (shown in green in Fig.5, page 9) (Tomkeieff, 1965; Mitchell, 2008) around 380 Ma (Harrison, 1996). Intrusive porphyrite has formed dykes within the lava dome (shown in red in Fig. 5, page 9). The granite forms the highest ground due to its high resistance to weathering. The surrounding andesitic and basaltic lavas are generally purple or grey and sometimes flecked with white spots (Robson, 1966). The Cheviot Massif has been deeply incised by streams, some of which have formed along fault lines cut through the granite (such as Harthope Burn, see Fig. 4). To the north and west of the Cheviot Massif, sandstone and argillaceous rocks are found with mafic lavas, microgabbroic and microgranitic rock. Further to the west, Upper Old Red Sandstone is found (Tomkeieff, 1965). To the east of the Cheviot Massif, along the NNCP, the limestone, sandstone and argillaceous rocks are heavily intersected by faults running predominantly E-W.

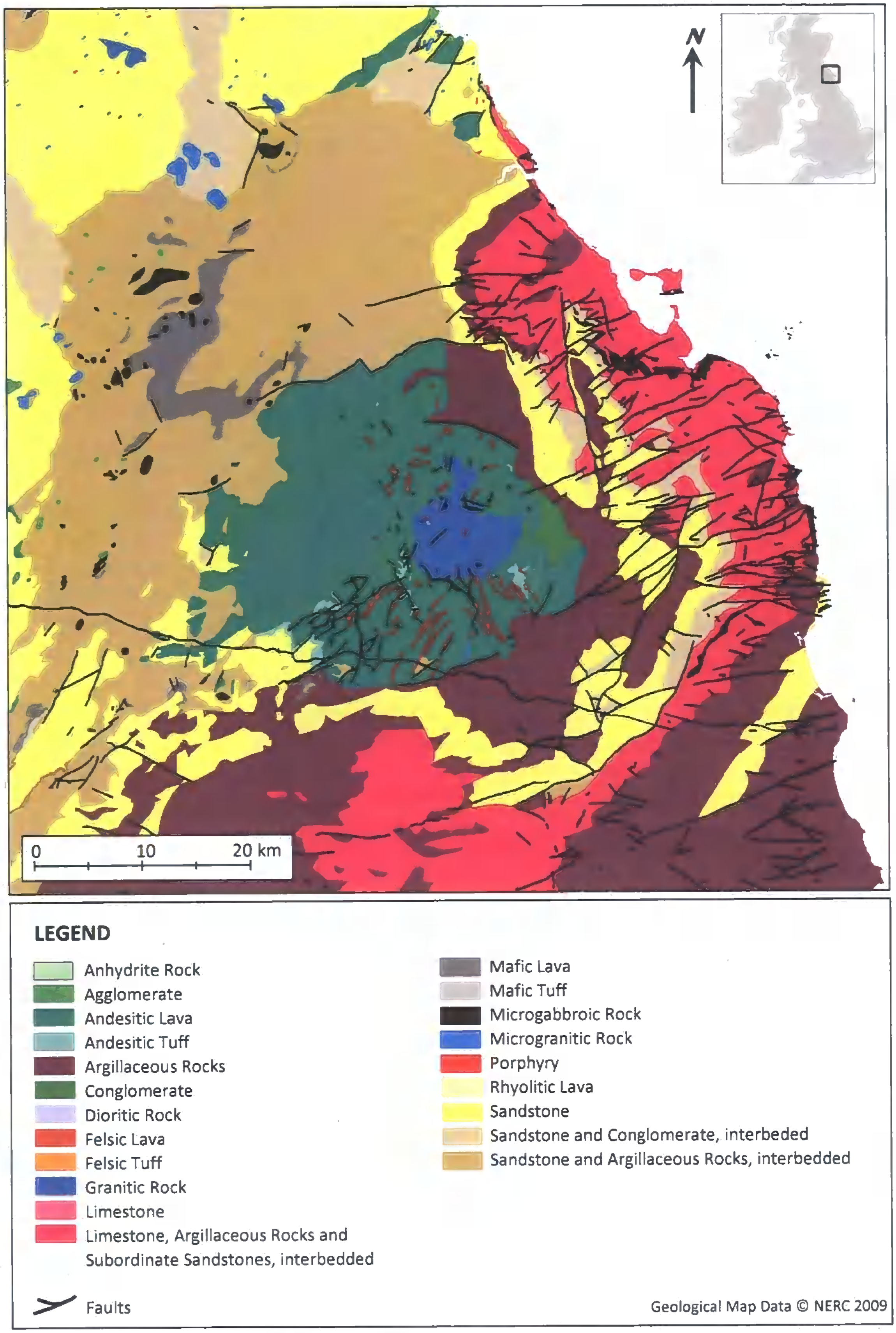


Figure 5: *Bedrock Geology of Northeast Northumberland and the Tweed Valley. Inset map shows location within the British Isles.*

1.5 Thesis Structure

The second chapter provides an overview of what is understood about the ice dynamics of the region together with a review of the existing literature on the landforms and sediments of the field site. The third chapter describes the key methods utilised in the project. This is followed by the results and interpretation chapter, in which the data is presented and analysed in the context of existing knowledge on glacial processes. The fifth chapter discusses the implications of the results for the reconstructions of the regional ice sheet dynamics. The final chapter summarises the key findings and conclusions, along with suggestions for further studies.

2

LITERATURE REVIEW: GLACIAL HISTORY OF FIELD SITE

As mentioned in section 1.2, the Quaternary record of northeast England is relatively incomplete and is based on morphostratigraphic correlations with upland sites elsewhere in Britain. What little is understood of the glacial dynamics is based on the presence of erratics, striae, till fabrics, glaciofluvial deposits, drumlins and glacial drainage channels throughout the region (Lunn, 1980). In this chapter, the inferred glacial dynamics of north east England are presented. This is followed by a review of the current literature on the glacial geomorphology and sedimentology of the Tweed Valley and the Cheviots.

2.1 Quaternary Glacial History

The Devensian began around 115 ka BP. During the Middle Devensian cooling intensified and by 30 ka BP tundra conditions existed across the British Isles (Huddart, 2002a). Temperatures continued to drop and by 30–26 ka BP, polar conditions marked the start of the Late Devensian. From 26–13 ka BP, ice cover in north-east England was at its greatest extent, although there is much debate over the maximum areal extent of the ice sheet and the timing of the LGM across the region (Huddart, 2002a). In Northumberland, ice was thought to have been at its maximum extent around 18 ka BP (Lunn, 1995). Ice flowed into the region from the Southern Uplands, the Solway basin and the Scottish Highlands. Flow directions have been inferred from onshore and offshore evidence, including erratics, streamlined bedforms, striae and till fabric analysis. It is also thought prior to 24 ka BP, the FIS and BIIS were confluent in the central northern sector of the North Sea basin (Nygard *et al*, 2007; Bradwell *et al*., 2008b). The generally accepted relative chronology of ice flow events in Northumberland and the Scottish Borders is as follows. Around the LGM (~18 ka BP), flow was predominantly from the west, with ice flowing from accumulation areas around the Solway Firth and in the Southern Uplands (Lunn, 1980; 1995) (see Fig. 6, page 12). Flow of this westerly ice was diverted around the Cheviots in Northumberland, with the Tweed Ice Stream (TIS) flowing around the northern margins and Solway Ice (flowing through the Tyne Gap) around the south (Fig. 6).

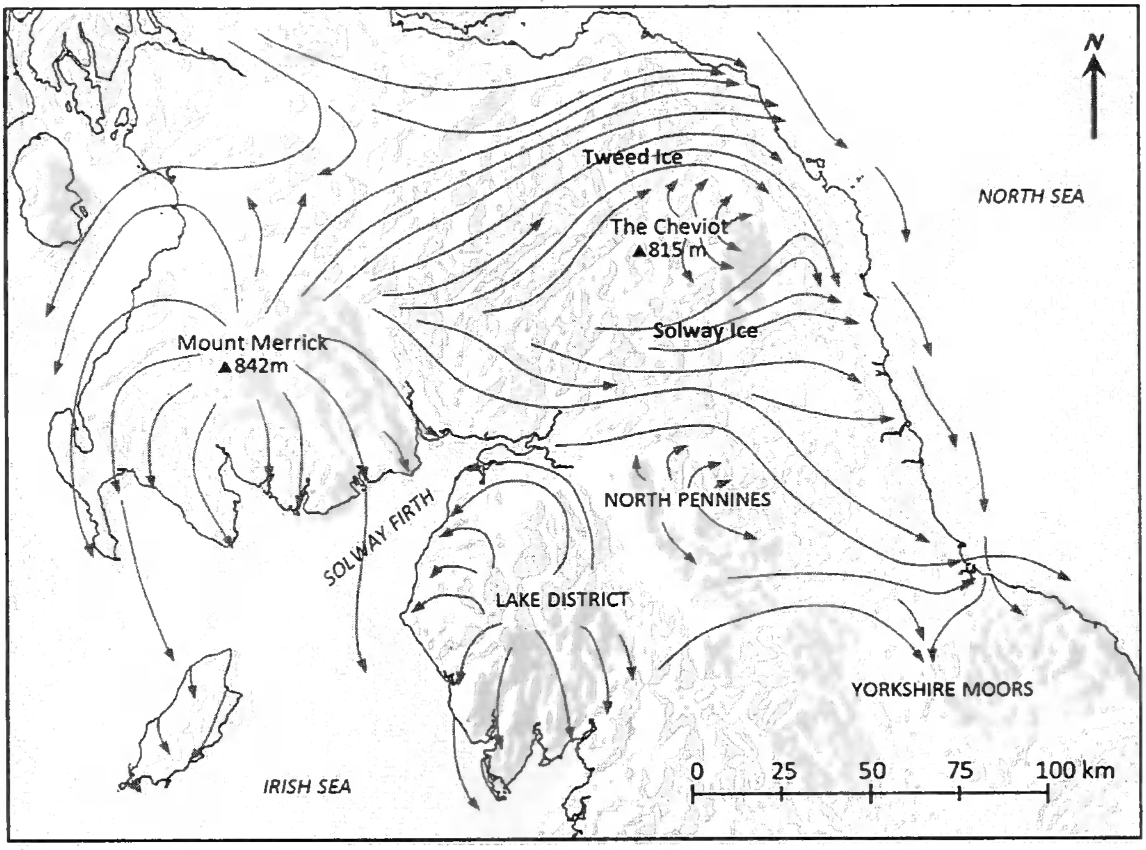


Figure 6: *Ice flow directions (grey arrows) in northern England and southern Scotland during the Late Devensian (modified after Clapperton. 1968; Clapperton, 1971a; Huddart and Glasser, 2002a). Basemap NEXTmap® DEM shows topography of region (NEXTmap® Britain, Intermap Technologies 2008).*

Geikie (1876, cited in Clapperton, 1970b) proposed that glacial erratics, tills and striations throughout Northumberland were evidence that foreign ice from the west had overridden the summits of the Cheviots. Conversely, Clough (1888), proposed that the highest summits were independent centres of glaciation, from which ice radiated to coalesce with the foreign ice on its lowest flanks. Charlesworth (1957) supported this theory, suggesting that the summits had been occupied by a small ice cap or semi-consolidated nevé field. Carruthers *et al.* (1932) disagreed with the theories of both Geikie and Clough, instead proposing that there was little evidence to support the notion that ice occupied or crossed the centre of the massif. If an independent ice mass did exist, they proposed that it was at a very early stage and retreated before foreign ice withdrew from the lower flanks. Despite several subsequent researchers supporting this theory (e.g. Common, 1957; Sissons, 1964), it is now widely agreed that the pattern of ice configuration in the region was that proposed by Clough: the Cheviots were entirely buried by a cold-based ice cap that was present during a relatively late stage of glaciation (Clapperton, 1970b; 1971b; Huddart, 2002c; Mitchell, 2008), although there remains some debate on the timing (Harrison et al., 2006). This ice flowed from the dispersal centre on the plateau summits to coalesce with the TIS and Solway Ice on the lower flanks of the hills.

At the LGM, ice flow in northern England was, as mentioned above, predominantly from west to east (Lunn, 1995). Following the LGM, the dominance of ice flow from the Lake District and the Solway Firth area was reduced. From the distribution of erratics along the NNCP it has been proposed that Tweed, Cheviot and Solway ice were diverted southwards to flow along the east coast (Sissons, 1964; Huddart and Glasser, 2002). This diversion is understood to be the result of the growth of the North Sea Lobe (NSL) that flowed southwards from the Scottish Highlands into the North Sea Bain (NSB) (Huddart and Glasser, 2002). More recently, a complex multi-lobate late-Devensian ice sheet has been invoked along the east coast of Britain, with ice lobes flowing from dispersal centres in southern and northeast Scotland, followed by ice flowing from the Scottish Highlands along the east coast (Davies *et al.*, 2009). Several reasons have been suggested for the southerly flow direction of the NSL, including the geology of the glacier bed influencing meltwater and ice flow; the growth of the Fennoscandian Ice Sheet in the North Sea Basin (e.g. Boulton *et al.*, 2001); and glacio-isostatic depression of the east coast of England creating a topographic low that channelled flow south (Teasdale and Hughes, 1999).

Following the retreat of the NSL around 18 ka BP (c.f. Lambeck, 1993), glacial lakes formed between the NSL and the western ice masses (Teasdale and Hughes, 1999). Lakes that have been hypothesised here include Glacial Lake Humber (e.g. Bateman *et al.*, 2008), Glacial Lake Pickering (e.g. Day, 1995) and Glacial Lake Wear (e.g. Raistrick, 1931). By 15 ka BP, the onshore margin of the BIS is inferred to have been north of the Scottish border, with only small areas of ice existing in the Lake District (Huddart, 2002a). It is estimated from correlation with Scottish glaciations that the region was fully deglaciated around 13 ka BP and that deglaciation was an uninterrupted process (Lunn, 1980, 1995). More recently, it has been suggested that the BIS re-advanced into the NSB after 18 ka BP (e.g. McCabe *et al.*, 2005; Nygard *et al., 2007*), however, the timing and extent of these readvance episodes remains highly contested (Fretwell *et al.*, 2008). McCabe *et al.* (1998; 2005) proposed that a significant readvance occurred around 14,7 – 14 ^{14}Cka BP (~15 ka BP). This readvance of the NSL has been linked to Heinrich Event 1 and the associated cooling in the North Atlantic brought about by the collapse of the LIS (McCabe *et al.*, 1998; 2005). Nygard *et al.* (2007) proposed that the margin of the NSL oscillated at least twice between 17 – 15.5 ka BP.

During the Younger Dryas (Loch Lomond Stadial), around 10,800 ^{14}C BP (~13 to 11 ka BP), ice readvanced in the Scottish Highlands, the Lake District and the Southern Uplands (Lowe and Walker, 1997). It has been proposed that periglacial conditions existed at this time, apart from in the Bizzle just north of The Cheviot, where a small cirque glacier developed (Harrison et al., 2006). Following the Younger Dryas, temperatures increased into the Holocene. Around 10 ka BP, sea levels began to rise and by 6 ka. BP much of the characteristic coastline of Northumberland was formed (Horton *et al.*, 1999).

2.2 Glacial Geomorphological and Sedimentological Evidence

Reconstructions of the extent, dynamics and configuration of the Devensian ice sheet in the Cheviots and Tweed Valley are based on, as mentioned above, a limited body of sedimentological and geomorphological evidence. Dating back to the late 19th century, glacial studies have focussed heavily on detailed landform description and classification, with little consideration given to reconstructing regional ice dynamics and the relationships between landform groups. Landforms identified and attributed to glacial activity include moraines, kames, eskers, lake plains, orientated and streamlined bedforms, streamlined tors, glacial striations and meltwater channels. In this section, the existing literature on the geomorphology and sedimentology of the Cheviots and Tweed Valley is summarised.

2.2.1 Meltwater channels.

Arguably the most frequently described of the landforms in and around the Cheviots are meltwater channels (Fig. 7, page 15). These features are distributed unevenly throughout the area, with the vast majority carved into the north-eastern and south-eastern slopes of the Cheviots. Channels have also been identified in the upper reaches of the Tweed Valley (c.f. Clough, 1888; Carruthers *et al.*, 1932; Common, 1957; Price, 1960; Derbyshire, 1961; Clapperton, 1968; 1970a; 1970b; 1971a; 1971b; Huddart, 2002c). Channels vary in form from small, metre-wide depressions to large, complex anastomosing systems with deeply incised v-shaped profiles, multiple inlets and multiple outlets (Lunn, 1980; 1995). The largest channels are found in cols and valley heads, are frequently steep-sided with occasionally undulating long-profiles (Lunn, 1995). In many instances, channels are cut across spurs (Common 1957). Most channels are currently dry or are occupied by small streams and marshy ground.

It is widely agreed that these meltwater channels were cut by meltwater during deglaciation as ice began to downwaste and extensive meltwater systems developed within the ice (Clough, 1888; Carruthers *et al.*, 1932; Common, 1957; Price, 1960; Derbyshire, 1961; Clapperton, 1968; 1970a; 1970b; 1971a; 1971b; Huddart, 2002c). The orientation of these channels, which is in large parts controlled by ice flow (Clapperton, 1971a; 1971b; Lunn, 1995), has permitted reconstructions of local ice flow trajectories. Around the northeast margin of the Cheviots the Tweed ice flowed in a south-easterly direction, approximately at right angles to the valleys and spurs (Clapperton 1968; 1971a). Despite the agreement that these channels were eroded by glacial meltwater, the exact mechanism by which these channels were cut varies greatly between authors. Theories include erosion by over-spilling lake water, ice-marginal and sub-glacial erosion and superimposition of englacial conduits.

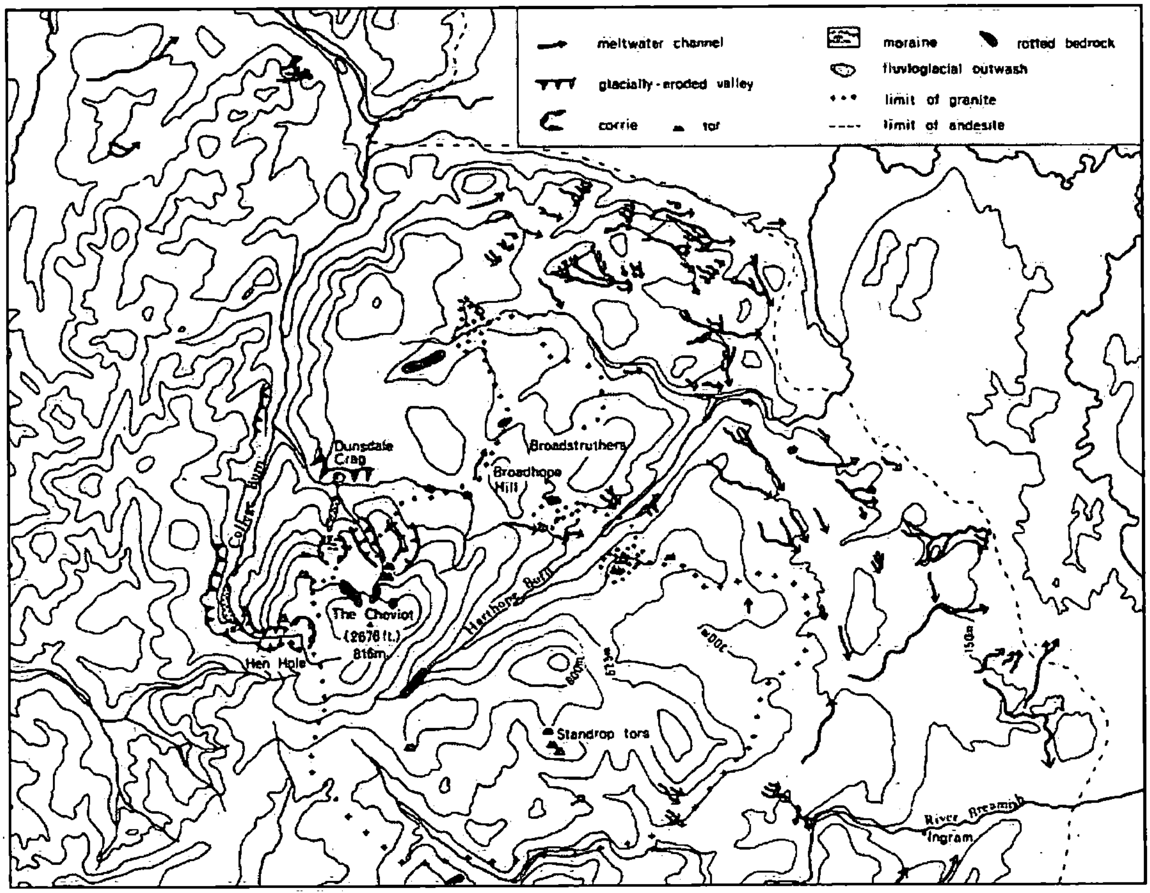


Figure 7: *Meltwater Channels of the northeast Cheviots, as mapped by Clapperton (1970). Map taken from Clapperton (1970, page 120).*

Initially, Clough (1888) attributed such channels in the Cheviots to erosion by glacial meltwater streams, although the position of these streams in relation to the ice was not expanded upon. A few decades later, theories emerged that erosional channels (e.g. in Cleveland Hills, Yorkshire) were cut as water overspilling from glacial 'lakelets' (Kendall, 1902). This theory proposed that where small, glacially-impounded lakes drained over a col or spur, channels were cut (Kendall, 1902). A lack of evidence supporting the existence of such lakes in the Cheviots led to the widespread rejection of this theory (c.f. Carruthers *et* al., 1932; Common, 1957; Derbyshire, 1961). Furthermore, several channels identified in col gullies exhibit up-and-down long profiles, double intakes, double outlets and are frequently occupied by glacial tills, which together have been interpreted as evidence of subglacial erosion (Derbyshire, 1961). The up-and-down long profiles of many channels is also used to reject the lake-overflow theory, as there is "no reason to suppose that outflowing glacial waters should erode their channels of the lake-ward side of the point of overspill to such a marked degree" (Derbyshire, 1961:38).

Carruthers *et al.* (1930, 1932) suggested that channels cut across spurs and along valley sides in the Cheviots were the product of ice-marginal meltwater erosion. This theory was supported by Common (1957), who proposed that channels at progressively lower heights, such as those in the Hedgeley Basin, were cut by retreating ice, with each channel recording a progressive lowering in the ice surface. Like Common (1957), Price (1960) attributed ice-marginal meltwaters with channel formation in the upper reaches of the Tweed Valley (Price 1960). Price (1960:487) interpreted that channels running down the hillsides at "considerable angles to the contour lines" were subglacial chutes, formed by water plunging down slopes and underneath the ice (Clapperton, 1968). Common (1957) suggested that channel form was heavily influenced by the rate of ice retreat, which varied in different localities. Their resulting form was understood to be a reflection of the length of time the channels were occupied by meltwater and the source and routing of the meltwater (Common, 1957). The bedrock geology is also proposed to have had an influence on the form of these channels; Derbyshire (1961) proposed that interrupted sections along the longitudinal profiles of subglacial channels were a result of changing rock types along the channel length, and also structural weaknesses such as rock joints, dykes and sills.

Price (1960) interpreted a small number of severed spur channels in the Tweed Valley as evidence of erosion by supraglacial or englacial channels. When these channels came into contact with the hillside, erosion occurred and channels were cut. Price (1960) concludes that only a relatively small number of channels in the upper Tweed Valley were formed by this 'superimposition' mechanism. This theory was expanded by Clapperton to explain the occurrence of meltwater channels in the Cheviots. Clapperton (1968) proposed superimposed channels are most likely to be englacial in origin, rather than supraglacial (Fig.8, page 17). Supraglacial meltwater systems tend to be ephemeral and in, most instances, find the easiest course down through crevasses. He points out that rarely is an entire conduit superimposed; sections may remain englacial whilst others are subglacial. This may lend an explanation to the interruptions along the long profiles of certain channels (c.f. Common, 1957; Derbyshire, 1961).

Clapperton (1968) also states that the properties of the ice mass govern the depth to which meltwaters can penetrate, suggesting that water within the BIS could penetrate only around 300-400 ft (~122 m) and that channels only formed when the ice was this thickness.

Clapperton (1970a) observes that of all the channels mapped, few exhibit an outlet that its more than 400 ft below the inlet, which he argues is a clear indication that the hydrologic properties of the ice influence channel formation. The theory of channel superimposition as a causative mechanism of channel formation in the Cheviots led Lunn (1995) to propose that a 'thermally layered' ice sheet must have existed, with a cold impenetrable base and a temperate region above.

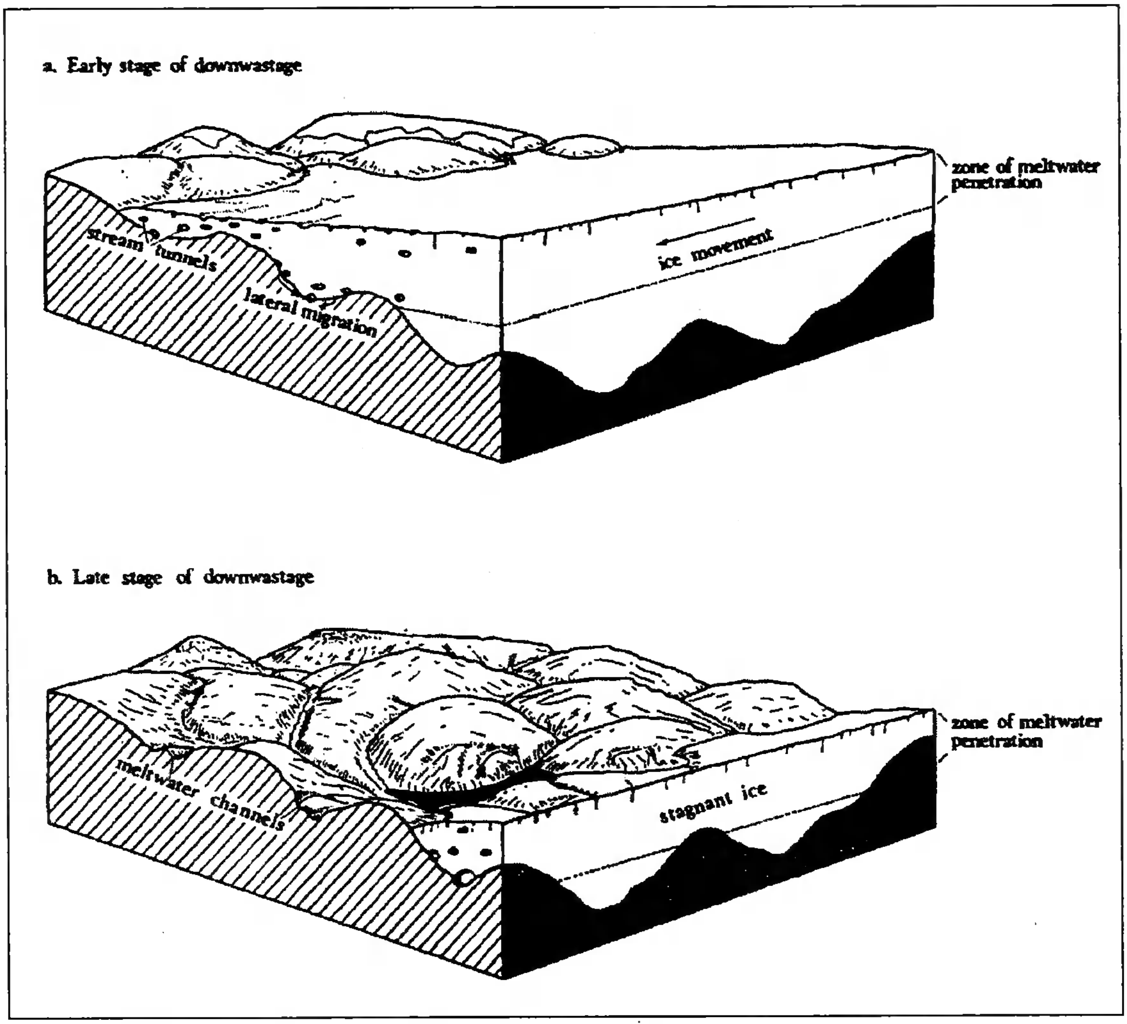


Figure 8: *Clapperton's model of englacial channel superimposition (image taken from Clapperton, 1968, page 209).*

2.2.2 Meltwater deposits

Where many of these meltwater channels reach the lower slopes of the Cheviots, extensive complex sand and gravel deposits are found (Clapperton, 1968; Lunn, 1980). These deposits have an uneven distribution across the region, although they occur mainly in low-lying areas (Clapperton, 1971a), with the most extensive systems found around Bradford, Cornhill on Tweed and Wooler (for location see Fig. 3, page 7). The Bradford Kame is "more characteristic and better developed than any other in the county" (Carruthers *et al.* 1927) and is well documented in the literature. The complex can be traced for approximately 20 km from Belford in the north, to the Aln Valley near Alnwick in the south (Carruthers *et al.*, 1930, see Figure 9, page 1919). Elongate mounds and sinuous ridges along this distance are composed primarily of sand, silt and gravel (Huddart, 2002b). Carruthers *et al.*, (1930) also identified mounds of heterogeneous assemblages of matrix-supported boulders. An early theory suggested a pro-glacial 'glacieluvial' (glaciofluvial) origin associated with ice flowing from the west (Gregory, 1922). Carruthers *et al.* (1930) rejected a western ice origin, and instead proposed the deposits represent a moraine formed at the southern margin of the lobe of the Tweed glacier. More recently, it has been suggested that the complex represents a subglacial esker that supplied meltwater and sediment into a series of open crevasses (Parsons, 1966, cited in Huddart, 2002). Huddart (2002b) interprets the system as a combination of subglacial eskers, pro-glacial and glaciolacustrine deltas, glaciofluvial fans and supraglacial kames, formed during active ice retreat. However, due to the lack of sedimentary exposures along the length of the Bradford Kame complex, many of the inferences made on the processes of formation and the pattern of ice retreat remain purely speculative.

Gunn and Clough (1895) described in detail the deposits south of Cornhill, to the north of the Cheviot Massif. The mounds, banks and ridges (frequently referred to as 'kaims') are in places composed of coarse, 'sharp' gravel and in others composed of well-rounded coarse gravel with pockets of dirty sand and clay. Porphyrite, which is found in the nearby Cheviots (shown in red in Fig. 5, page 9), makes up a large proportion of the deposits (Gunn and Clough, 1895). Frequently, hollows occupied by small ponds or peat bogs (Gunn and Clough, 1895) are found between these mounds. The presence of these pits within the sands and gravels led Carruthers *et al.* (1932) to refer to this system as a moraine of pitted and kettled sands, that formed on the ice surface at the southern lateral margin of Tweed Glacier. This system is believed to extend to Wooler and further south to the Breamish and Hedgeley Basin (Fig. 10, page 20), where accumulations of poorly-sorted sands and gravels have also been associated with hill-washed sands and gravels and stagnating ice (Carruthers *et al.*, 1930). They argue that the large proportion of 'hill-washed' gravels within this system is evidence that the Cheviots did not support an independent ice cap, as these gravels must have come from "ice-free ground fully exposed to denudation" (Carruthers *et al.* 1932). The characteristic hummocks and pits are interpreted to have formed as ice buried by supraglacial deposits slowly stagnated during the retreat phase of the Tweed glacier (Carruthers *et* al., 1932). The theories proposed by Carruthers *et al.* (1930) and Carruthers *et al.* (1932) were later rejected by subsequent researchers as was the use of the word 'moraine' to describe these features.

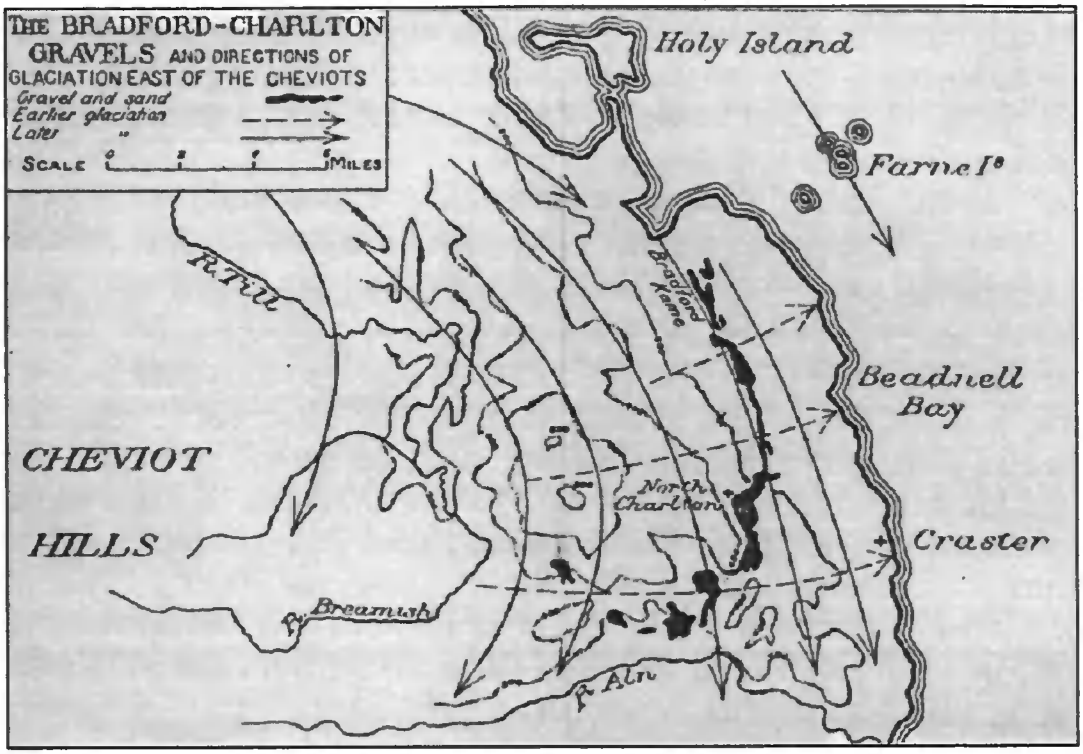


Figure 9: The Bradford-Charlton Gravels (shown in black) as mapped by Carruthers et al (1930). Arrows show direction of ice flow. Map taken from Carruthers et al (1930).

Sissons (1964; 1967) proposed this 'kame moraine' was formed at the southern edge of a large lobe of ice that extended down the Tweed basin during the Aberdeen-Lammermuir readvance. Alternatively, Derbyshire (1961) attributes the formation of the deposits south of Wooler with fluvioglacial deposition between two phases of meltwater channel erosion from a rapidly stagnating ice mass over the Cheviots. The deposits in this region are predominantly well sorted, show clear bedding structures and are interpreted as eskers, hummocks and kame terraces (Clapperton, 1971b). Much of the material comprising these features is thought to have been derived from superimposed meltwater channels (see section 2.2.1), re-worked till deposits and englacial debris (Clapperton, 1971b), and not from ice-free hillslopes as suggested by Carruthers *et al.* (1930, 1932). These glaciofluvial landforms are interpreted to have formed during deglaciation, when supra-, en- and sub-glacial tunnels formed in stagnating ice and transported and deposited vast volumes of debris (Clapperton, 1968; 1971a). Clapperton (1971b) assigned these landforms to a period of deglaciation during which meltwater drainage was 'ice directed', i.e. controlled by the surface slope of the ice. During this time, meltwater channels were cut and esker and kame systems deposited.

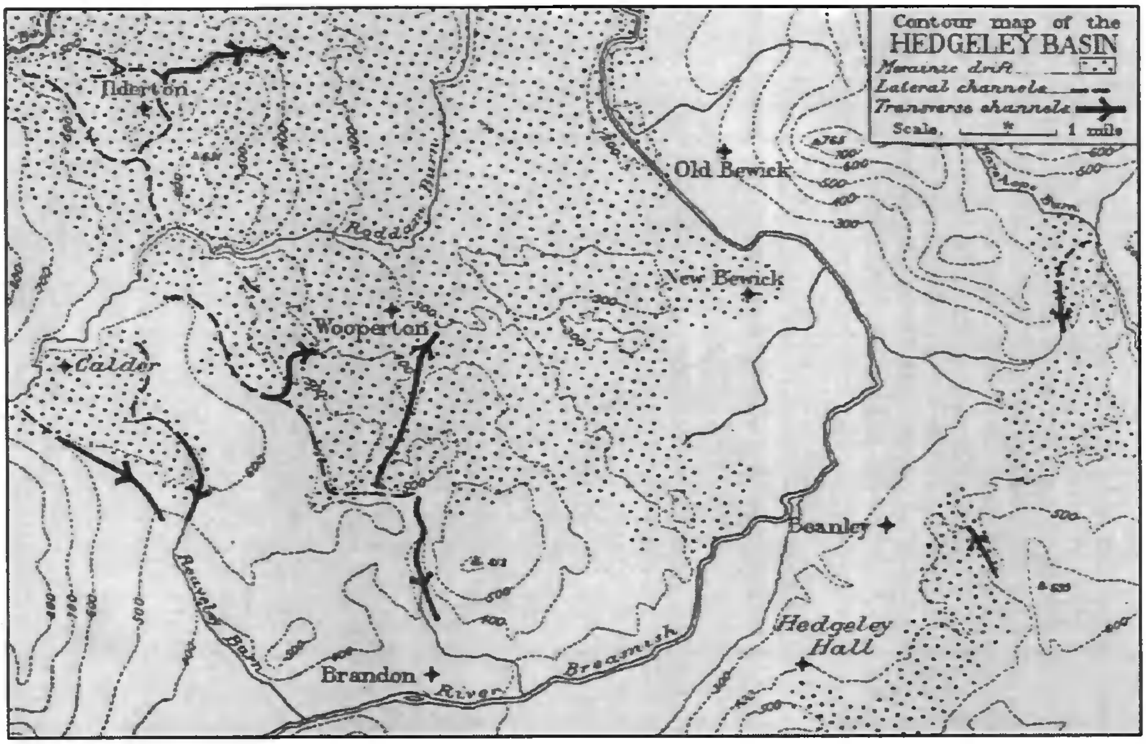


Figure 10: *'Morainic Gravels' of the Hedgeley Basin, as mapped by Carruthers et al (1930, p.90). Black arrows show location of 'dry-valleys' (most likely meltwater channels). For the location of the Hedgeley Basin see Fig. 3, page 7.*

2.2.3 Lake deposits

Following the period of 'ice-directed' meltwater drainage (Clapperton, 1971b), glacial lakes formed where the ice masses began to separate along their zones of confluence. In the Hedgeley basin (see Fig. 10, above) and the mid-Aln valley, which are separated by Shawdon Dene, laminated silts and clays and sand and gravel deltas were interpreted by Clapperton (1971b) as glaciolacustrine deposits. In the mid-Aln valley, the separation of the southern and northern ice masses allowed a lake to form to an altitude of 60 m (Clapperton, 1971b). In the Hedgeley basin, meltwater ponded to an altitude of 85-90 m as the northern (Tweed) ice mass separated from the Cheviot ice. Evidence in the form of an incised delta in Shawdon Dene suggests the mid-Aln lake drained before the Hedgeley basin lake, which drained when the Tweed ice had retreated from the valley of the present River Till (Clapperton, 1971b).

Before the Tweed ice had fully retreated, lakes also formed in the Wooler-Chatton area and across the Milfield Plain (Fig. 11, page 21). Gunn and Clough (1888) described sand and gravel deposits at the edges of the Milfield Plain (NT 990 330), arranged in a flat-topped terrace south of Milfield, small mounds north-east of Milfield and a 'great spread' around Fenton Mill (NT 970 342). They suggest that they are in some way related to an 'old lake' that occupied the plain. Carruthers *et al.* (1932) further described the deposits of the Milfield Plain and associated these with the end of the glacial period. Lunn (1980) interpreted thick deposits (>22 m) of laminated lacustrine clays and silts to indicate that a glacial lake formed here. This lake, known as Glacial Lake Ewart (Butcher, 1967) is believed to have formed during a late stage of glaciation, reaching levels of 45 m (Evans *et al.*, 2005). A sand and gravel delta also formed as sand and gravel were washed from an ice lobe blocking the nearby Glenn Valley (Lunn, 1995). Lake Ewart drained northwards beneath the receding Tweed Ice, cutting deeply into the bedrock beneath, forming the incised meanders occupied by the present day River Till (Lunn, 1980).

Figure 11: *View west across Milfield Plain, which was occupied by a lake during deglaciation. Photo: KEHS. For location of the Milfield Plain see Fig. 3, page 7.*

2.2.4 Glacial Tills and Drumlins

The tills of this region (referred to as 'boulder clay' or 'drift') were described in detail by the British Geological survey in early 20th century. Along the NNCP and further to the south along the coast and up to approximately 15 km inland, an upper and a lower till are identified (Lunn, 1995). Carruthers *et al.* (1927) identified that these tills were separated by fluvially-deposited sands and gravels or laminated clays. The lower till is a coarse blue-grey clay, with a high proportion of clasts of both local and foreign provenance and occasional rip-up clasts (Carruthers *et al.*, 1927; Lunn, 1995). The upper till is purplish-red/brown with a high proportion of Cheviot and northern clasts and is less coarse than the lower till (Carruthers *et al.*, 1927; Lunn, 1995). Carruthers *et al.* (1927: 131-132) suggested that the two tills formed in the presence of one ice sheet; the lower till is "a true ground moraine" and the upper "englacial detritus", i.e. the tills were emplaced by lodgement and then melt-out respectively (c.f. Boulton, 1980). The mechanism of till emplacement has been proposed to have been a combination of 'unconformable facies superimposition', a result of the ice sheet switching between erosion and deposition (Eyles *et al.*, 1982) and post-glacial weathering (Eyles and Sladen, 1981; Eyles *et al.*, 1982). More recent work on the Devensian tills along the North East coast further to the south in County Durham has suggested that the upper and lower tills are subglacial traction tills (Davies *et al.*, 2009). The lower till has been associated with ice from the Midland Valley and western Southern Uplands and the upper with ice an ice lobe flowing from the eastern Grampian Highlands (Davies *et al.*, 2009).

Within the valleys of the Cheviot Hills, glacial till is distributed unevenly. Till is found up to an altitude of 455 m (Clapperton, 1970b) predominantly on lee slopes facing down-glacier (Lunn, 1995). Directly south of the River Tweed, just to the north of the Cheviots, only one till has been identified (Gunn and Clough, 1895). The till is red and sandy, with few boulders and is thicker towards the west (Gunn and Clough, 1895). Around Wark Common (NT 820 367), bore holes indicate that the drift thickness is over 100 ft (30 m). Gunn and Clough (1895) also observed in many locations around Coldstream that the till formed long parallel ridges, interpreted to be drumlins. From the long axes of these, ice flow direction is inferred to range from north-east to east (Gunn and Clough, 1895; Carruthers *et al.*, 1932). Drumlins and glacial striations throughout the Tweed Valley and along the NNCP have been mapped by Clapperton (1970) (Fig. 12, page 23). The till composing the drumlins were found to have a higher proportion of stones than the surrounding lower ground (Gunn and Clough, 1895). Around Ford (NT 946 375) and Etal (NT 925 395), a high proportion of porphyritic boulders (i.e. erratics) are found. Clapperton (1971a) suggested that a few drumlins around Kelso (for location see Fig. 3, page 7) composed of re-worked glaciofluvial deposits, although acknowledged that the majority are composed of till. It was also considered possible that some drumlins are rock cored (Carruthers *et al.*, 1932). Sissons (1964, 1967) suggested the drumlins were formed during the Aberdeen-Lammermuir readvance at the same time as the Cornhill esker and kame complex. This was largely rejected by Clapperton (1971a), who stated that the drumlins must pre-date the Cornhill complex as the glaciofluvial deposits overlie the drumlins in many situations (as observed by Gunn and Clough, 1895; Carruthers *et al.* 1932). Furthermore, the deposits around Cornhill are as 'fresh' as those around Wooler, further evidence that the former were not deposited by a late-stage readvance (Clapperton, 1971a).

Clapperton (1971a) asserted that the highly elongate form of these drumlins indicate that the flow of the Tweed glacier was fairly rapid. This led Everest *et al.* (2005) to propose that the Tweed Ice Stream (TIS) flowed between the Cheviot and Lammermuir Hills towards the North Sea from its onset zone in the upper Tweed Valley. Fast ice flow here has been associated with high pore-water pressures facilitating basal sliding (Everest *et al.*, 2005), which streamlined the underlying highly deformable red sandy till. Everest *et al.* (2005) inferred ice flow directions in the Tweed Valley from the megaflutes, streamlined terrain and drumlins mapped by Clark *et al.*, (2004) in the Glacial Map of Britain. The closely-spaced, highly parallel nature of the drumlins, megadrumlins and megaflutes were interpreted as evidence that only one flow event occurred (Everest *et al.*, 2005). The TIS is thought to have flowed during a late stage of deglaciation, and may have even initiated deglaciation across the wider region (Everest *et al.*, 2005), although the exact timing is unknown. Whilst it has previously been acknowledged that Tweed and Solway ice were diverted southwards by growth of the NSL (e.g. Sissons, 1964; Huddart and Glasser, 2002), this is not considered by Everest *et al* (2005). Indeed, there has been little detailed work undertaken on the TIS; its flow dynamics, timing and interactions with the surrounding ice and topography are not fully understood.

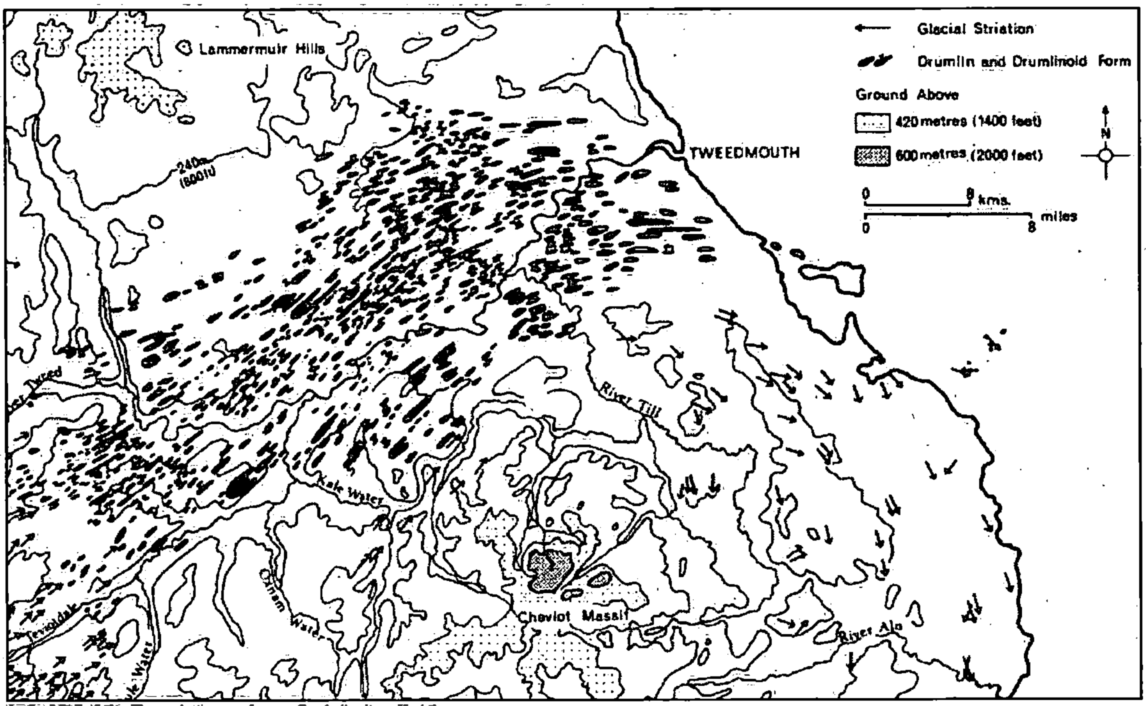


Figure 12: *Glacial Striae and Drumlins of the Tweed Valley and NNCP mapped by Clapperton (1970). Map taken from Clapperton (1970, p.116).*

3

METHODOLOGY

3.1. Introduction

To describe, understand and explain the impact of the last ice sheet in the Tweed Valley and the surrounding area, it was necessary to conduct a detailed investigation of the glacial geomorphology and sedimentology. It is possible to establish the processes responsible for the formation of landforms (and therefore ice dynamics) through combining a detailed analysis of their morphological and sedimentary characteristics (Vitek 1996) with "empirical and theoretical studies" of the processes associated with landform development (Napieralski et al., 2007; 1). The latter has been achieved by undertaking a thorough study of the literature surrounding the field site and Pleistocene ice histories at local, regional and continent-wide scales (see previous two chapters). The main techniques employed in this study to classify and understand the landforms of the field site were geomorphological mapping and sedimentary analysis.

3.2 Geomorphological Mapping

The technique of mapping is integral to glacial geomorphology and Quaternary studies, and has long been utilised by researchers to display spatial data. Geomorphological maps are generated by identifying and grouping landforms on the basis of their age, origin and common morphologic characteristics and are used to display the "spatial distribution of phenomena" (Vitek et al., 1996; 234). Until recently, the majority of geomorphological maps were produced in the field by mapping features onto topographic base maps (Smith et al., 2006). With regards to this study, it was deemed impracticable to carry this out over the entire study area. The field site covers an area of around 3800 km^2; such a large (and relatively inaccessible) area could not be mapped effectively at the scale required due to time constraints and lack of manpower. In order to map the area effectively, it therefore was necessary to identify an alternative technique.

3.2.1 Remote Sensing and GIS technologies

Remote sensing of the Earth's surface produces data that that has been "generated without direct human measurement" (Richards, 1981; 41) and includes aerial photography, satellite imagery (e.g. Landsat Thematic Mapper, SPOT) and digital elevation models (DEMs) (Smith et al., 2006). The spatial and elevation data acquired by remote sensing can be explored using a geographical information system (GIS). The development of GIS technologies since the 1960's has led to new and efficient ways of dealing with geographical data (Vitek et al., 1996). Geographically referenced objects can be stored, referenced, manipulated, analysed and visualised easily (Smith et al., 2006). Napieralski et al. (2007) emphasise the importance of GIS to palaeo-ice sheet reconstructions, as vast volumes of data requiring a variety of analytical approaches can be stored and worked on with great ease. Although GIS software is often relatively expensive (as opposed to the materials required for 'pen and ink' mapping), it is becoming cheaper, is easily accessible and 'user friendly', and allows large areas to be studied and mapped by fewer people in shorter time periods (Smith and Clark, 2005). In many geomorphological studies, mapping remotely has replaced field mapping as the method of choice (Smith and Clark, 2005). For this study therefore, mapping from remotely acquired data was considered to be the most appropriate method.

Aerial photographs were not used due mainly to the inherent complexities surrounding image variability and processing, which results in highly variable mapping results (Smith *et al.,* 2006), high costs, and the difficulty of obtaining them for the chosen study area. In many studies it has proved near impossible to map from aerial photographs (c.f. Smith *et al.,* 2006). Satellite imagery (e.g. Landsat Thematic Mapper) often has a much greater spatial coverage than aerial photography and is relatively cheaper. The resolution and accuracy of the data however allows only moderate detail to be identified, and is not appropriate for detailed mapping in palaeo-glaciological terrain (Smith *et al.,* 2006). This study required a high resolution dataset, in order to identify glacial landforms and perform the appropriate morphometric analysis. Mapping from a DEM in a GIS was therefore considered to be the most appropriate method as pixel resolutions can be around 10 m and height accuracies around 1.0 m (Smith and Clark, 2005). A DEM is define as a "regular gridded matrix representation of the continuous variation of relief over space" (Gülgen and Gökgöz, 2004:2). DEMs hold data on the absolute elevation of a landscape and can therefore be used to "visualise landscapes" (Smith *et al.,* 2006; 149).

3.2.2 The NEXTmap® Britain Database

The DEM chosen for this study was generated by the NEXTmap® Britain mapping programme (Intermap Technologies™, 2008), which uses airborne single-pass Interferometric Synthetic Aperture Radar (IfSAR) technology to remotely survey the landscape. Two radar antennae fixed to an aircraft emit and receive radar signals reflected by the earth's surface. DEMs are created by laying the recorded elevation measurements over a grid (Intermap Technologies™ www.intermap.com, 2008). The main reason for using NEXTmap® (as opposed to DEM's derived from other sources) was the access to the data. It is also widely acknowledged that airborne surveys generate higher detail than spaceborne sensors (such as Satelite Radar Topography Mission (SRTM) and Landmap) due to a higher level of sensitivity (Smith et al., 2006). Resolution and accuracy values of SRTM DEM's are around 90 m and 6 m respectively, whereas the NEXTmap® DEM has a vertical resolution of 1 m and a horizontal resolution of 2 m (Bradwell et al., 2008). Only LiDAR (light detection and ranging) derived data is of a higher resolution and accuracy. This however was not available for the chosen field site. Furthermore, when mapping glacial terrain, NEXTmap® DEMs have been shown to provide a better estimate of ground truth than DEMs derived from other sources (Smith *et al.*, 2006).

3.2.3 Map Generation

Two geospatial databases were used in which to generate the map – Erdas Imagine 9.1 and ESRI ArcMap. Erdas was used to mosaic DEM tiles and map landforms and ArcMap was used to generate the end-result map. Two DEMs were used in this study (Fig. 13, page 27); a digital surface model (DSM), which is a geometrically-correct base map and a digital terrain model (DTM), which is digitally 'smoothed' to remove artefacts such as buildings, vegetation and roads. Both these visualisations were used to reduce error in the landform identification process. The mosaiced images were visualised using mulitple-azimuth relief-shading to highlight variations in the topography of the field site (Smith and Clark, 2005). The limitations of these visualisation methods are examined in the next section. To generate the map, vector layers were generated, each including one landform assemblage, for example, 'hummocks', 'elongate features'. Care was taken to identify and group landforms solely on the basis of their morphology. Landforms that were identified by previous authors (see chapter 2) were also included as were those in existing GIS databases (e.g. the Glacial Map of Britain, Clark *et al.*, 2004). The boundaries of the features were delimited through creating polygon, point and line shapefiles. From the map, certain areas of interest were chosen to map in detail at scale of 1:25,000. This increased the number of features that could be mapped in detail (Hubbard and Glasser, 2005). These large-scale maps were generated in much the same way as the compilation map using ERDAS and ARC, although the DEMs were overlain by a semi-transparent 1:25,000-scale digital OS maps (from Edina digimap www.edina.ac.uk/digimap). The compilation map and the OS maps also highlighted suitable areas for sedimentary analysis (section 3.3).

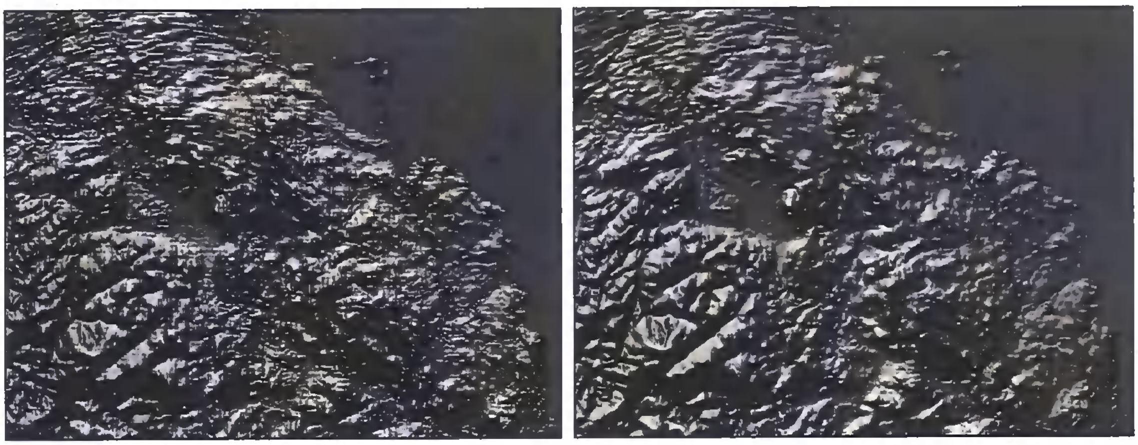


Figure 13: *DEMs of field site* ***LEFT:*** *DSM of the area around Wooler* ***RIGHT:*** *DTM of the same area. Note the flat area in the centre of the image (The Milfield Plain) is 'smoother' in the DTM – artefacts such as trees and buildings have been removed (NEXTmap® Britain, Intermap Technologies, 2008)*

3.2.3.1 Limitations and reducing errors

The quality of the interpretation depends heavily on the quality of the maps produced, so every effort was taken to reduce errors when mapping. To maintain consistency and reduce variability in landform identification, only one person (the author) carried out both computer and field mapping. Furthermore, certain areas were re-mapped a few weeks after the initial mapping process to ensure the consistency of landform identification. For all of the maps generated, features were extensively ground-truthed. This is important as in previous studies, mapping from NEXTMap DEMs has resulted in the mis-identification of certain landforms (c.f Smith *et al*. 2006). Clearly, generating a map of glacial landforms requires prior knowledge of glacial landforms and processes, although every attempt was made during mapping to group landforms solely on their morphology.

An important issue in the generation of landform maps from relief-shaded DEMs is that of azimuth biasing (Smith and Clark, 2005). Through altering the angle of the 'sun' over the image, landforms can be concealed, revealed or incorrectly identified. This is particularly relevant for linear landforms which are concealed when parallel to the chosen azimuth (Cooper, 2003). Therefore, to reduce the likelihood of mis-identifying features, the sun-angle was frequently altered during mapping. The DTM and DSM were used for several reasons, mainly due to the inherent limitations of the NEXTMap DEMs, which are as follows. Firstly, the DSM is an un-altered representation of the height of the land surface, so artefacts such as settlements and roads appear as hummocks and ridges. The likelihood of identifying these as landforms was reduced by mapping onto the DTM and DSM at the same time using two monitors. Secondly, IfSAR technology often performs poorly in forested areas, due to its large radar footprint (Smith *et al.*, 2006). To overcome the possibility of identifying these as glacial landforms, the vector layers were lain over 1:25,000-scale digital OS map tiles (www.edina.ac.uk/digimap).

3.3 Sediment Analysis

It has been suggested that sedimentary evidence from formerly glaciated regions is often more important than the geomorphic record in palaeo-environmental reconstructions (Lowe and Walker, 1998). Through the description and classification of glacigenic deposits, inferences can be made on the glacial processes and climatic conditions at the time of deposition (Krumbein and Sloss, 1951; Evans and Benn, 2004). Information can be gained on the types of ice sheets, their limits, dynamics and flow directions, the source of sediments and the modes of sediment transport and deposition. In view of this it was decided that sedimentary surveys would be carried out to supplement the mapping results, as this reinforces the validity of any palaeo-environmental inferences (Lowe and Walker, 1997). All available sites within the study area were selected for sedimentary investigations from the landform compilation map and from 1:25,000-scale OS maps. Effort was made to find sites within each landform group that would cause minimal disruption to the site and wildlife, for example river cuttings, quarries and pits. At each site the same method was employed; sediments were logged in the field and samples were taken for laboratory analysis.

3.3.1 Field Logging

Once the appropriate sections were identified, field logging was carried out. Firstly, notes were made on the general characteristics of the site, such as the aspect, dimensions and surroundings of the section. This permits an assessment of the contemporary processes affecting the section (e.g. fluvial and biological activity) and highlights any health and safety issues (Hubbard and Glasser, 2005). GPS readings were taken in order to accurately locate the site on the geomorphic map and make the site easy for subsequent researchers to locate. Secondly, slumped or disturbed material was removed from the section and loose debris was gently cleaned away using a trowel. A scaled detailed field sketch of the exposure was then drawn and sites were chosen at which to log. Every attempt was made to space these sites at equal distances along the exposure, although this depended greatly on the stability of the face and access to it. A tape measure was attached to the exposed face and detailed notes were made on the characteristics shown in table 1. Lithofacies codes from Evans and Benn (2004) were used to record lithofacies properties when logging (see Appendix A). Facies analysis was undertaken through using informal lithofacies codes (see Appendix 1a), which permitted the description of the identified facies in a descriptive and objective manner (Evans and Benn, 1998). Through utilising lithofacies analysis, bias at the interpretation stage was reduced.

Table 1: *Recorded characteristics of logged sediments (modified from Boggs, 1995; Lowe and Walker, 1998; Evans and Benn, 2004; Hubbard and Glasser, 2005).*

Characteristic	***Features recorded***
BEDDING	Beds (>1 cm think) were distinguished from each another by distinct vertical changes in clast characteristics. Their dimensions were measured and recorded, along with their shape (lenticular, tabular, wedge or trough). The nature of the contacts between beds (e.g. erosional, deformed or gradational) were recorded (Evans and Benn, 2004).
Internal structures	The presence of structures such as laminae (<1 cm thick), ripples, cross-strata, cross-laminae and foresets (Boggs, 1995; Evans and Benn, 2004) was recorded, as these are important indicators of varying flow velocities.
Sorting and Grading	The degree of sorting is established using a standardised scale that ranges from very well sorted to very poorly sorted (Appendix 1d). Sorting is a measure of "the range of grain sizes present and the magnitude of the spread or scatter of these sizes around the mean size" (Boggs 1995; 87). Grading (inverse, normal or massive) refers to the vertical change in grain size and reflects temporal changes in flow velocities and sediment supply.
Palaeocurrent data	Palaeocurrent data was generated using a compass-clinometer to measure the orientation of imbricated elongate clasts and the orientation of ripple crests. This allows former currents and flow directions to be inferred.
DEFORMATION STRUCTURES	The presence of folding, thrusting, faulting and de-watering structures reveal post- or syndepositional reworking of deposits, and can reflect fluctuations in ice margins, melting of stagnating ice or changes in flow regimes (Evans and Benn, 2004).
CLAST TEXTURE AND LITHOLOGY	The size, shape and lithology of clasts within each bed were recorded to provide information on clast provenance and mode of transport. The a- b- and c- axes of at least 50 randomly selected clasts from each bed were measured using callipers. In order to remain objective, Wentworth's (1922) particle size chart (Appendix 1b) was used to aid the description of clast size and Powers' (1953) roundness index (Appendix 1c) was used to record the angularity of clasts. It was ensured that only one person carried this out. It is particularly important to record the shape of clasts, as this data can be used to infer transport pathways and depositional settings (Lowe and Walker, 1998).

3.3.2 Laboratory Analysis

Samples were taken from beds for laboratory particle-size analysis. By combining these results with clast morphology data (obtained through sedimentary logging), information on the source material, transport processes and the environment of deposition can be inferred (McCave *et al.*, 1991; Evans and Benn, 2004). Samples were first air-dried, weighed and then dry-sieved. The fraction below 2mm (-1 on the phi (ϕ) scale) was subjected to analysis by a laser diffraction particle size analyser. The material was first riffled to obtain a >0.5 g representative sample and 20 ml of 20% hydrogen peroxide was added to destroy organic material. The samples were placed in a 40°C water bath for several hours, then were centrifuged and the acid decanted off. They were then rinsed twice before adding 20 ml of the deflocculent sodium hexametaphosphate. The samples were then analysed by a multi-wavelength Beckman-Coulter granulometer. As the laser beam passes through the sediment suspension it diffracts. The degree of scatter is recorded by a multielement ring detector (McCave *et al.*, 1991) and the flux pattern produced allows the distribution of particle sizes to be inferred.

4

DATA DESCRIPTION AND INTERPRETATION

Three distinct sediment-landform assemblages have been identified in the Tweed Valley and surrounding area. These are, (A) a track of numerous, elongate asymmetric ridges; (B) hummocks and sinuous ridges of varying shape and size found at the margins of this track; and (C) channels cut into the hillsides of the Cheviot and Lammermuir Hills. The spatial distribution of these landforms is shown in figure 14 (page 32). Features have been identified and differentiated from one another on the basis of their morphology and also their sedimentology.

4.1 Assemblage A: Elongate Ridges

4.1.1 Description

Elongate, closely spaced, sub-parallel ridges are identified in the Tweed Valley and across the North Northumberland Coastal Plain (NNCP) south of Berwick-upon-Tweed (Fig. 15, page 33). These ridges form a sharply delineated arcuate track that is bordered in the Tweed Valley by the higher ground of the Cheviots to the south and the Lammermuir Hills to the north. The track is convergent from west to east, narrowing from 35 km wide in the upper Tweed Valley, to around 15 km between the Cheviots and Lammermuir Hills. South of Berwick-upon-Tweed, the track extends to the south along the NNCP. In plan, the elongate ridges are ovoid to oval in shape and display an almost symmetric axis along their length (Fig. 16, page 33). Along the track, ridges are distributed sub-parallel to one another and are orientated along their long axis. Occasionally, ridges are connected, forming multiple-crested features (Fig. 17, page 33). In cross section, the majority of elongate ridges are asymmetric with a relatively steep-sided high point at one end and a gently sloping tapered tail at the other (see Fig. 16, page 33). The vast majority of features are orientated with their tapered ends pointing in the same direction: N around Kalemouth (NT 709 274), NE around Kelso (NT 725 345), ESE to the south of Berwick- upon-Tweed (NT 995 524) and SSE south of Bamburgh. Southwest of Berwick-upon-Tweed several features are orientated NE, which differs from the majority of the surrounding features (Fig. 18, page 34). Between the elongate ridges, subtle, parallel-sided linear and tapered features have been identified. These features are far longer and narrower than the elongate ridges but are orientated in the same direction. Elongate ridges and linear features range from 120 m to over 2.5 km in length, 10 m to nearly 80 m in height and 100 m to 500 m in width.

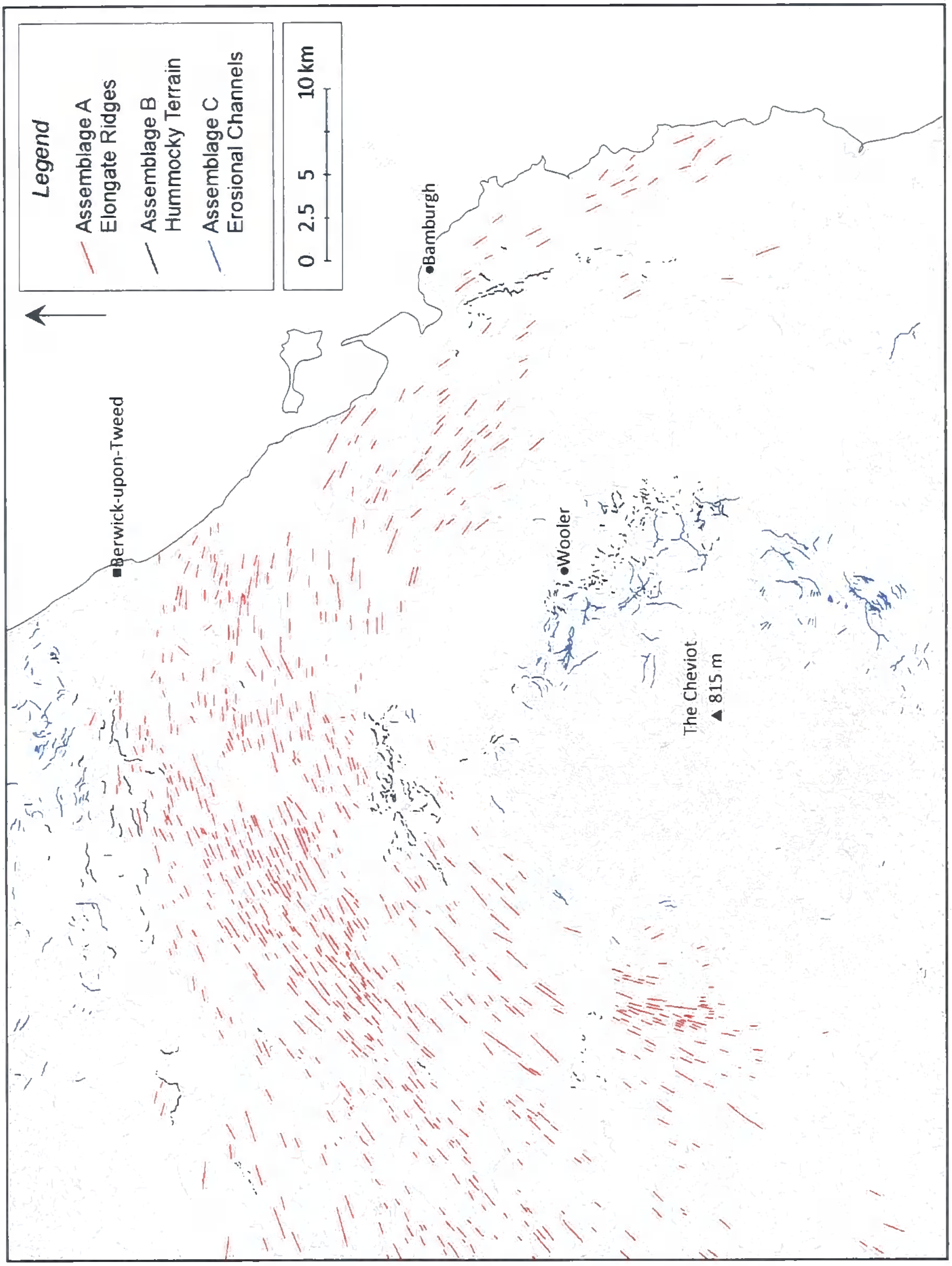


Figure 14: *Compilation map of glacial landform assemblages in the Tweed Valley and surrounding area.*

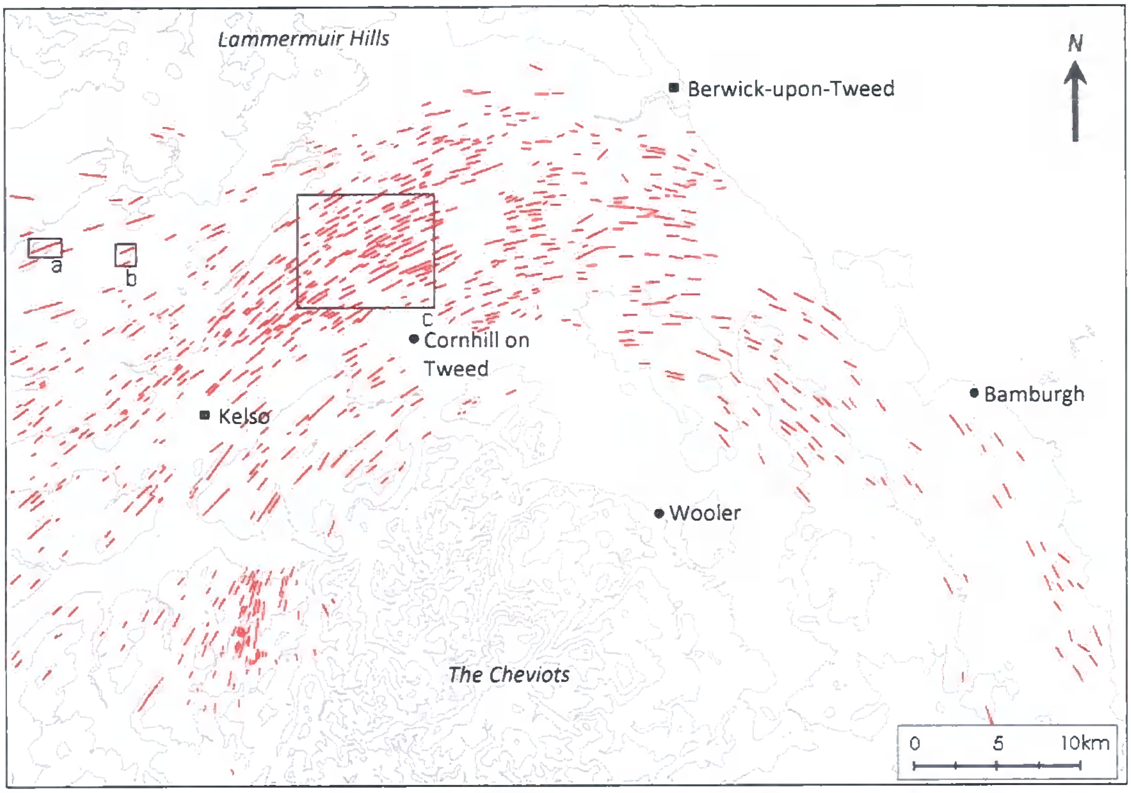
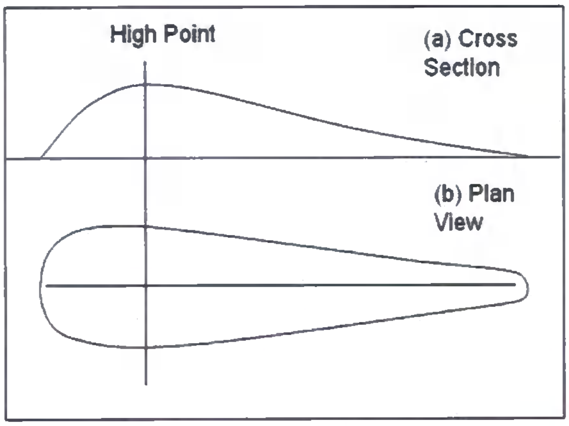
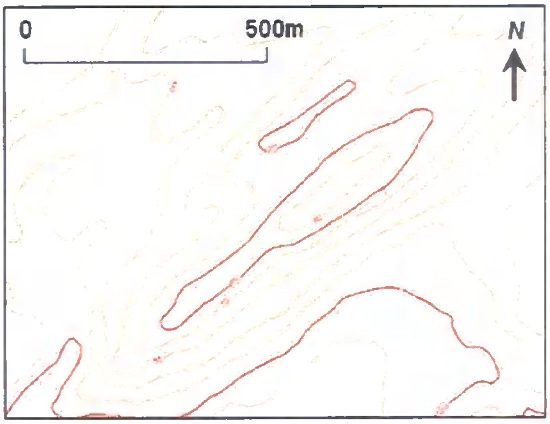


Figure 15: *Distribution of elongate ridges (shown in red). Contours shown in grey at 50 m intervals. Inset boxes a, b and c indicate areas shown in figures 21, 17 and 20 respectively.*





LEFT Figure 16: *Shape characteristics of elongate ridges of the Tweed Valley and North Northumberland Coastal Plain.*
RIGHT Figure 17: *Double-crested elongate features approximately 5 km NE of Kelso. Crests shown in red. Location shown in inset box b, Fig.15 (Basemap OS Landform Profile © Crown Copyright/ database right 2008. An Ordnance Survey/EDINA supplied service).*

Elongation (length to width) ratios of the ridges vary greatly across and along the track. The most elongate ridges found along the centre of the track, where they are most closely spaced and most numerous. The majority of bedforms longer than 1000 m are identified here (Fig. 19, page 34). In the vicinity of Coldstream, ratios frequently exceed 1:7 (Fig. 20, page 35). Nowhere along the track do elongation ratios exceed 1:12. Towards the lateral margins of the track, elongate ridges are less closely spaced and elongation ratios are lower. For example, in the vicinity of Gordon (NT 648 431), whilst elongate ridges are well-defined, steep-sided and exceed 70 m in height, elongation ratios are rarely higher than 1:3 (Fig. 21, page 35). Elongate ridges to the south of Berwick-upon-Tweed on the North Northumberland Coastal Plain have uneven edges, irregular contours and are more subtle than those further northwest.

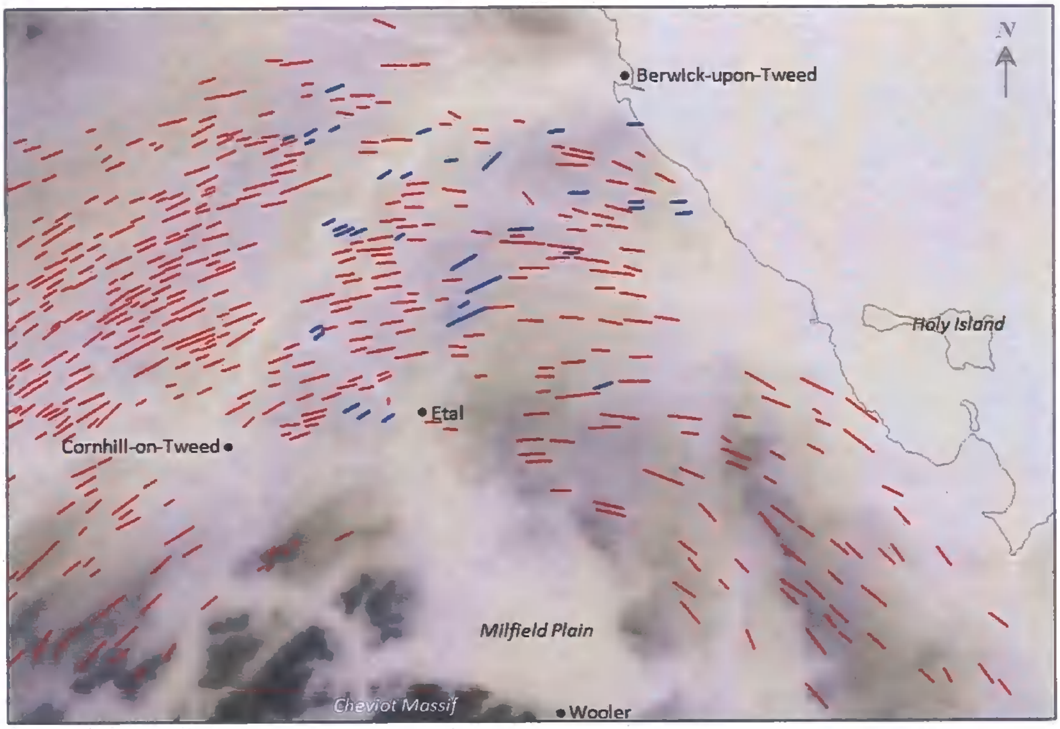


Figure 18: Position of elongate features southwest of Berwick, showing two distinct orientations. Those shown in blue are orientated towards the NE, pointing offshore. Those in red loop round the Cheviot Massif in an arcuate track. (Relief-shaded DTM from NEXTmap® Britain, Intermap Technologies)

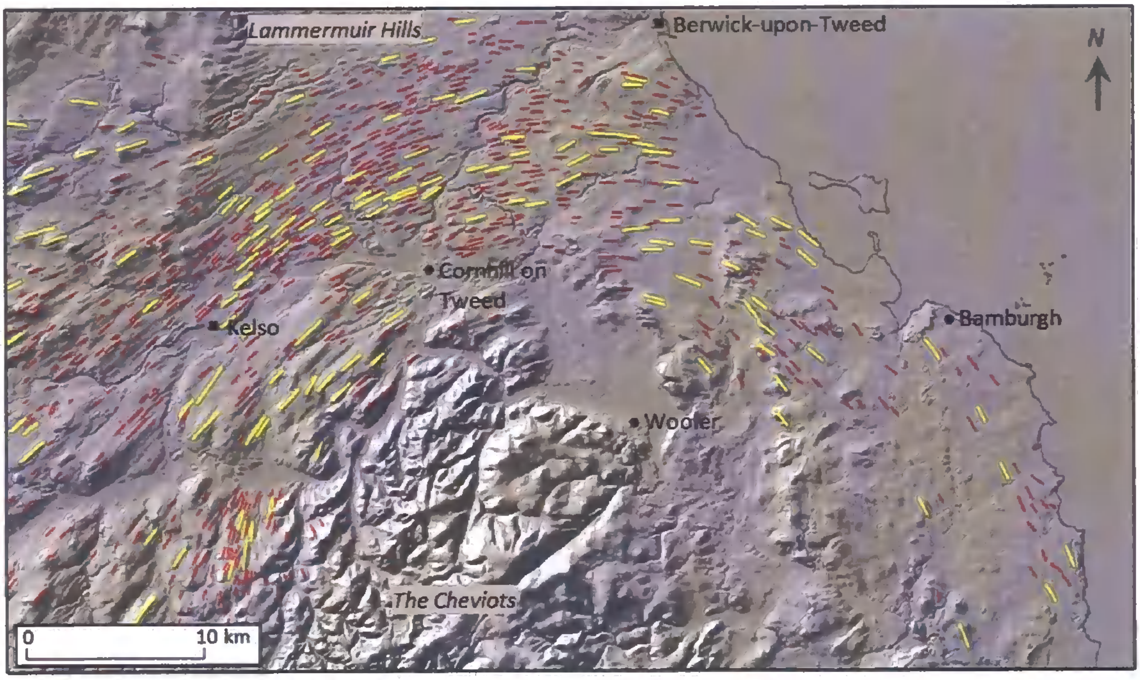


Figure 19: Elongate features over 1000 m in length (shown in yellow). Features in red <1000 m in length. (DTM from NEXTMap® Britain, Intermap Technologies)

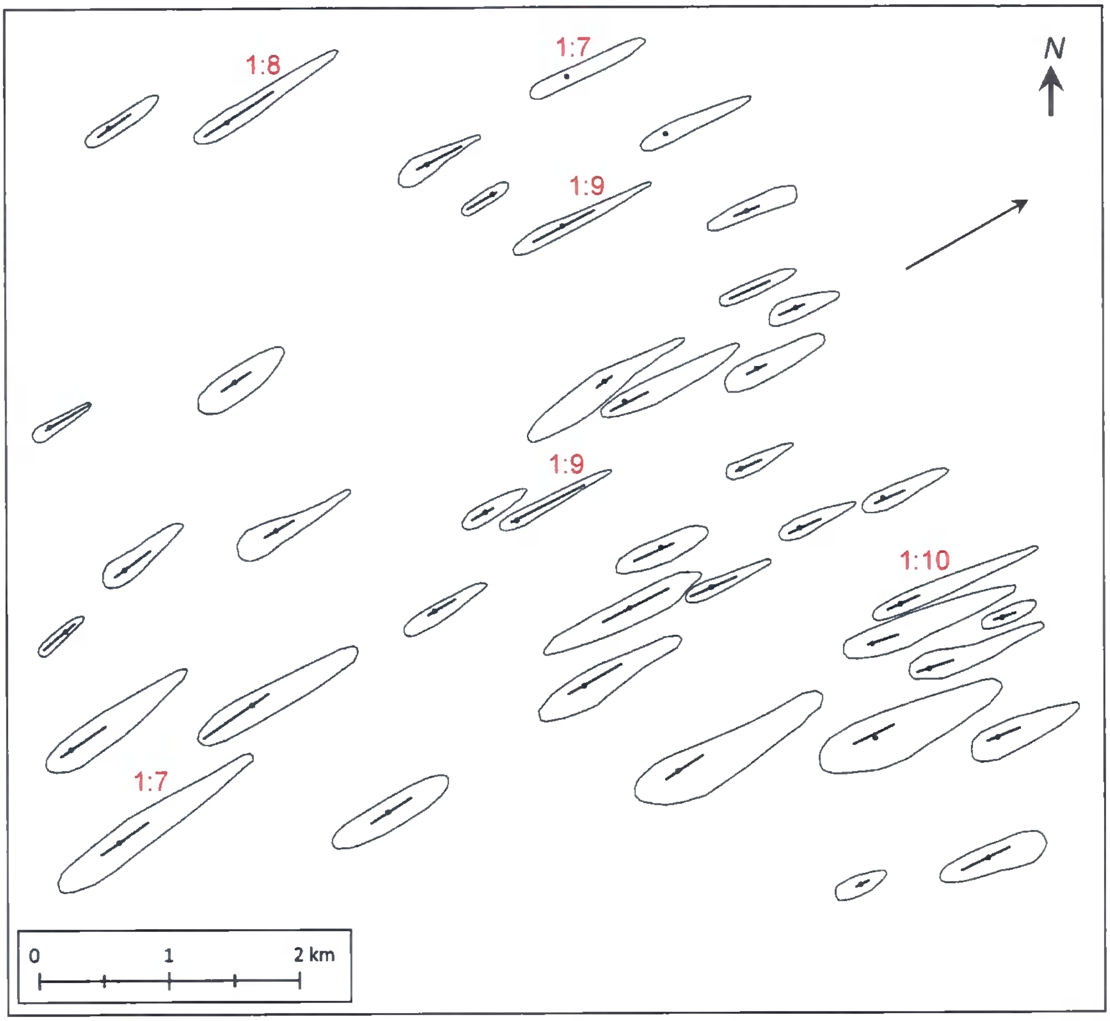


Figure 20: Highly attenuated ridges northeast of Coldstream (NT 841398). Ridge crests are marked with black lines, highpoints with a black dot. Elongation ratios are marked in red. Ratios range from 1:3 to 1:10. Black arrow denotes average orientation (61.2°). (Location shown in Figure 15, box c).

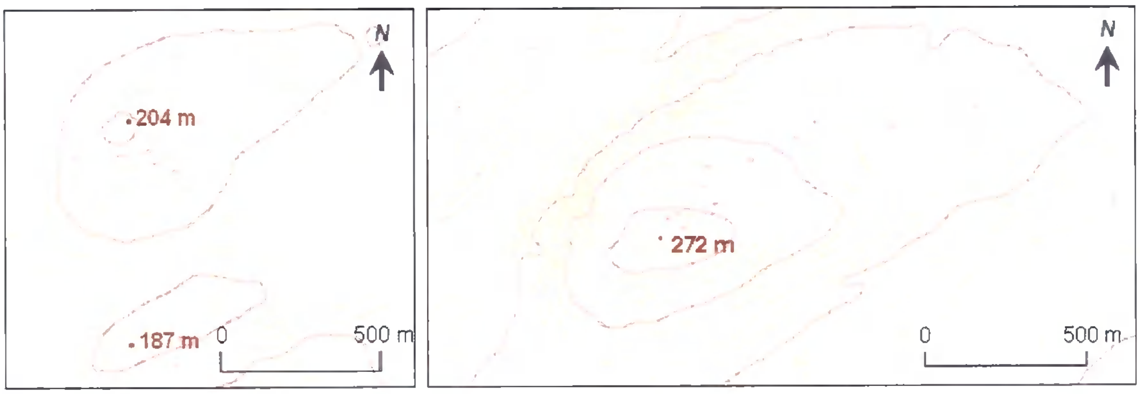


Figure 21: Well-defined elongate mounds in the vicinity of Gordon (location shown in inset box a, figure 15). Note elongate mounds with high point towards the west and tapered end towards the east. Orientation of long axes is approximately towards the northeast. (Basemap OS Landform Profile © Crown Copyright/database right 2008. An Ordnance Survey/EDINA supplied service).

With regards to the internal structure of the elongate ridges, exposures were not identified for sedimentological investigation. Therefore the geology of these features is inferred from descriptions in the literature and British Geological Survey geology maps (Geological Map Data © NERC 2008). A surficial geology map of the area (Fig. 22, page 36) shows that a large proportion of the region is covered by Devensian glacial till and pockets of glaciofluvial sand and gravel. From this map it is tentatively inferred that these elongate features are composed of glacial till as their distribution broadly coincides with the till cover (Fig. 23, page 37). This is in broad agreement with descriptions of previous authors who identified long parallel ridges formed of an upper and a lower glacial till (e.g. Gunn and Clough, 1895, see section 2.2.4, page 21). It has also suggested in the literature that some elongate ridges are composed of glaciofluvial sands and gravels (e.g. Clapperton, 1971a) and that some exhibit bedrock cores mantled by till (e.g. Carruthers *et al.,* 1932). Several elongate ridges found on higher ground to the west of the Cheviot Massif and to the west of the NNCP, do not appear to coincide with the Devensian Till cover and are suggested therefore to be formed of bedrock (circled in red in Fig. 23, page 37). This inference, is of course, somewhat speculative and would require further field surveys to verify. These landforms are some of the longest mapped, with lengths exceeding 1 km. In long profile, these features appear asymmetric. In terms of the bedrock structure, to the east of the Cheviot Massif faults trend approximately E-W within the limestone, sandstone and argillaceous bedrock (see Fig. 5, page 9). These faults lie roughly perpendicular to the orientation of the elongate ridges here.

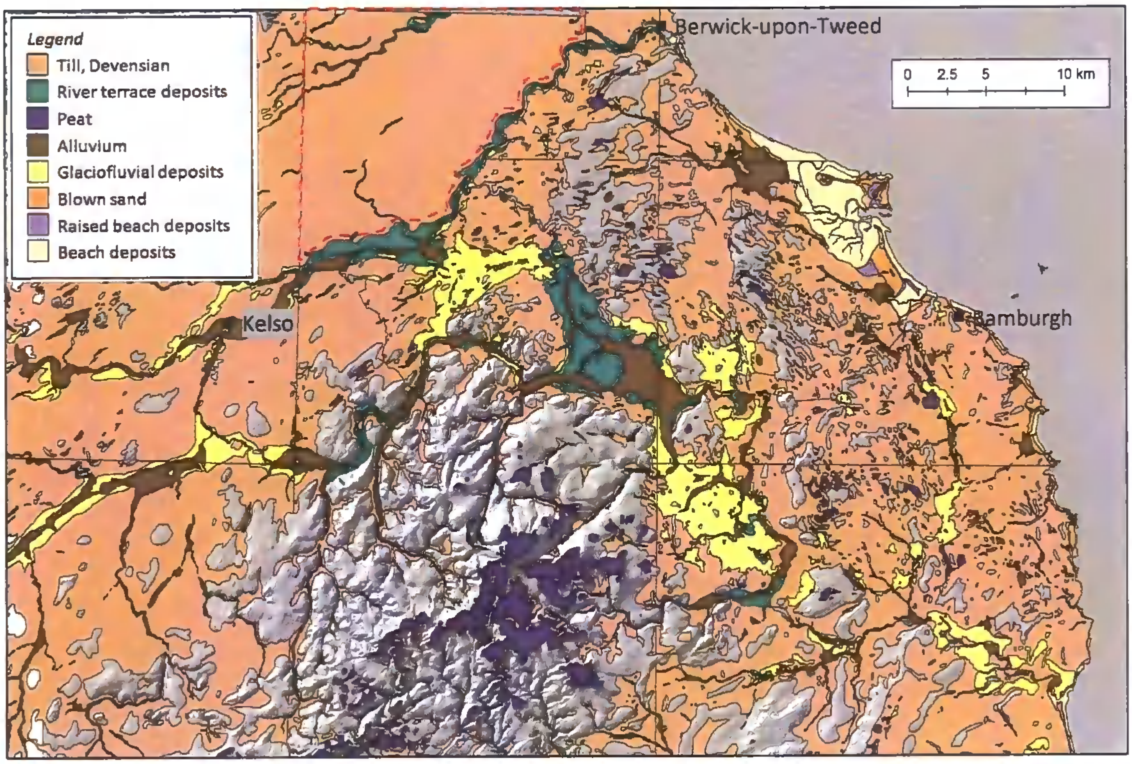


Figure 22: *Distribution of surficial deposits across northeast Northumberland and the Tweed Valley. Image underlain by hillshade DEM, highlighting areas where bedrock is not covered by deposits. Area outlined in red has not been mapped by the BGS, however the till cover is inferred from Geological Survey Memoirs. (Geological Map Data © NERC 2008, DTM from NEXTMap® Britain, Intermap Technologies).*

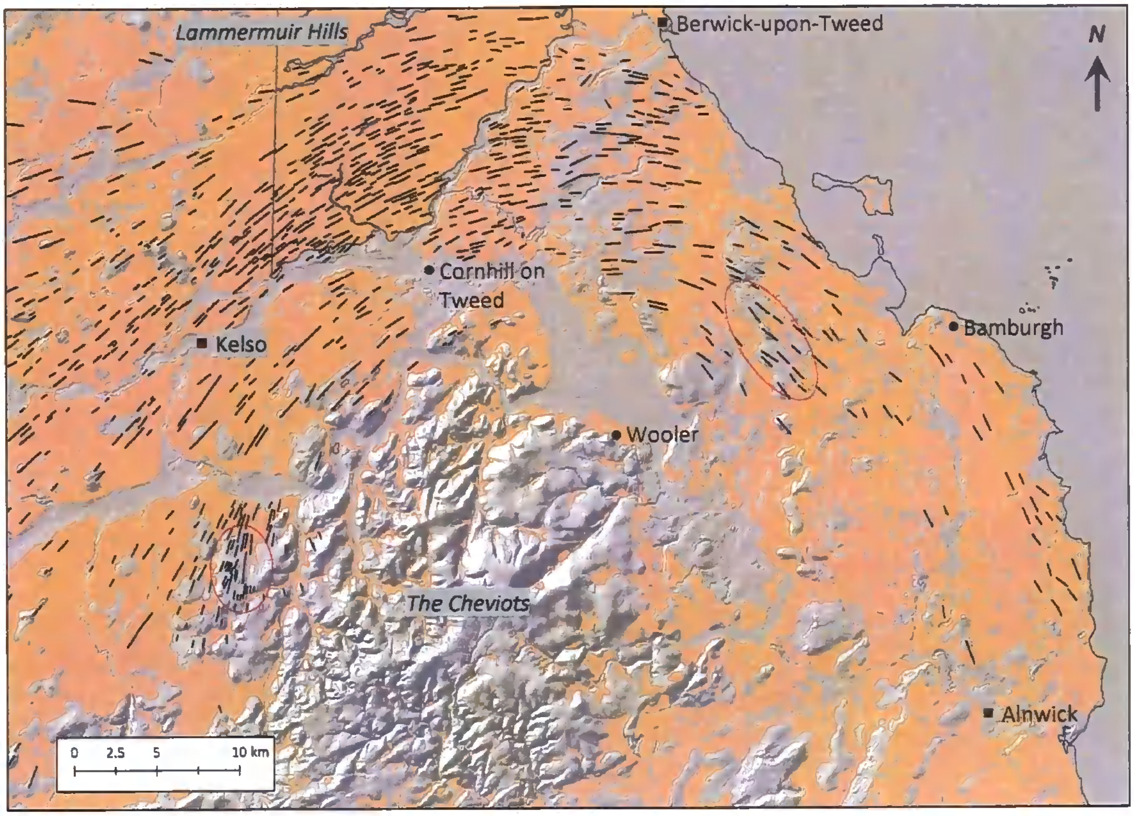


Figure 23: *Location of streamlined bedrock ridges (circled in red) and distribution of Devensian Till (shown in orange) (Geological Map Data © NERC 2008, DTM from NEXTMap® Britain, Intermap Technologies)*

4.1.2 Interpretation

The numerous, highly attenuated sub-parallel ridges identified in a broad track along the Tweed Valley and the North Northumberland Coastal Plain are interpreted as longitudinal subglacial bedforms. On the basis of their morphology, bedforms are interpreted as flutings, drumlins and streamlined hills. These landforms are differentiated from each another on account of their dimensions and elongation ratios, and also their distribution and relationship to neighbouring landforms. These distinctions are, however, somewhat arbitrary, as these features are part of a landform continuum (Rose, 1987). The smooth, parabolic to spindle-shape of these landforms, their number and high degree of parallel conformity are diagnostic characteristics of streamlined subglacial bedforms (Benn and Evans, 1998; Hillier and Smith, 2007). Elongate, asymmetric ridges with long axes >100 m and elongation ratios <1:7 are interpreted as drumlins (Rose, 1987). Asymmetric, streamlined ridges inferred to be formed of bedrock are referred to as rock drumlins. More subtle, elongate features between the drumlins orientated in the same direction are interpreted as flutings (<100 m) and megaflutings. The larger, drumlinoid features with long axes >1000 m, as identified along the centre of the Tweed Valley are interpreted as streamlined hills (Benn and Evans, 1998). The above interpretation is in agreement with previous researchers, who have all proposed a subglacial origin for these landforms (e.g. Gunn and Clough, 1895; Sissons, 1964; Clapperton, 1971a; Eyles *et al.*, 1982; Everest *et al.*, 2005).

Due to a lack of sedimentary exposures throughout the region, the exact mode of formation of these subglacial bedforms is speculative and is therefore based on their mapped morphology and morphometry (and to a lesser extent, their inferred sedimentology, see Figures 22 and 23, page 37). The theories on drumlin and flute formation are varied, although it is agreed that longitudinal, streamlined bedforms are a reflection of the dynamic conditions at the ice-bed interface (Rose, 1987). The most widely accepted theory is that of subglacial deformation, where sediment deforms around obstructions and protrusions at the ice-bed interface (Boulton, 1987; Boulton *et al.*, 2001). Rock drumlins are proposed to be the result of focussed glacial abrasion on the stoss side of bedrock protrusions. Reduced abrasion on the lee-side of these bedforms, resulting in their tapered form, is likely to have occurred where cavity formation was inhibited or where lee-side quarrying was inhibited by stable water pressures (Dionne, 1987; Benn and Evans, 1998). Rock-cored drumlins mantled by till (as identified by Carruthers *et al.*, 1932) are inferred to have been formed in a similar manner as rock drumlins, with till being deposited over the top of the streamlined bedrock protuberance (e.g. Hill, 1971). When considering the likely genesis of these rock drumlins and bedrock-cored drumlins, the bedrock structure must be considered. Rock drumlin orientation in the New York Drumlin Field has been associated with the strike of the underlying bedrock, which has resulted in drumlins whose orientation is independent of glacial erosion and ice flow (e.g. Kerr and Eyles, 2007). Faults running through the sandstone, limestone and argillaceous rocks of northeast Northumberland trend E-W (see Figure 5, page 9), trending approximately perpendicular to the orientation of the bedrock drumlins. It is therefore considered unlikely that bedrock structure has played a significant role in bedform genesis, instead selective glacial erosion of bedrock protuberances is attributed with the genesis of these features.

Clapperton (1971a) identified drumlins formed primarily of reworked glaciofluvial material in the Tweed Valley. Whilst this has not been verified in this study, their inferred structure can be explained in terms of subglacial deformation. As ice flowed over the Tweed till, which has a high proportion of clay and sand (Carruthers *et al.*, 1932), pockets of coarser glaciofluvial sands and gravels would have acted as an obstruction to ice flow, around which the weaker, finer grained till deformed (Boulton, 1987). Twin crested drumlins mapped around Kelso (see Fig. 17, page 33) are suggested to have formed where a crescentic pocket of coarser sediment was streamlined, forming two tapered downstream ends (Fig. 24, page 39). Local changes in basal conditions or sediment supply may have prevented lee-side infilling of the crescentic ridge (c.f. Boulton, 1987), leading to the formation of the observed twin-crested drumlins.

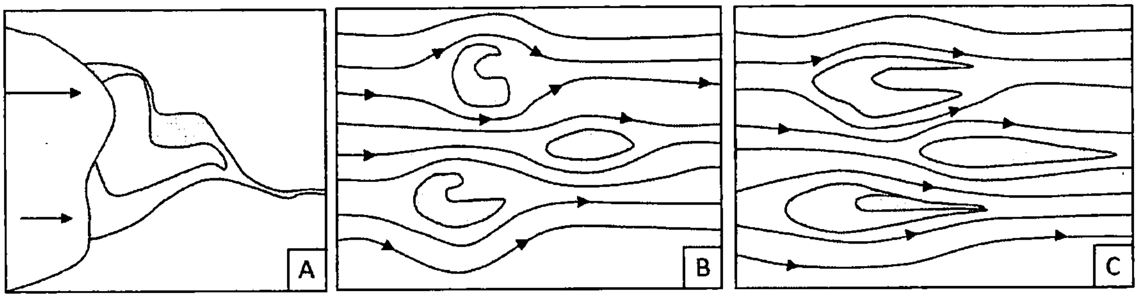


Figure 24: *Twin-crested drumlin formation.* ***Picture A:*** *coarse-grained sediment (shown in dark grey) is deposited, over which the glacier advances* ***(picture B)****, streamlining the bed.* ***Picture C:*** *the coarse grained material an obstacle to ice flow, resulting in deposition in the lee-side. Twin-crested drumlins are proposed to have formed where lee-side filling did not occur (Adapted from Boulton, 1987).*

The shape, elongation, dimensions and organisation of the drumlins, flutings and streamlined hills provide information on flow trajectories and ice-flow dynamics. As drumlins and flutings form with their long axes orientated parallel to ice flow, with their tapered end on the lee (downstream) side (Jauhiainen, 1975; Martini *et al.*, 2001), flow trajectories can be reconstructed. In the central section of the Tweed Valley, between the Cheviots and Lammermuir Hills, ice flow was broadly from southwest to northeast, as has been recognised by previous researchers (e.g. Lunn, 1995: Everest *et al.*, 2005). In the lower reaches of the Tweed Valley, southwest of Berwick-upon-Tweed, bedforms have been identified orientated towards the NE, out of trend with the overall orientation of the bedforms in arcuate loop around the Cheviots (see Fig. 18, page 34). The orientation of the bedforms in this area and their proximity to neighbouring longitudinal landforms has led to the identification of two flow-sets (c.f. Clark, 1999) (Fig. 25, page 40). These flow-sets were differentiated from each other by the parallel concordance and close spatial association of the bedforms with adjacent bedforms (Clark, 1994, 1999). Those drumlins orientated towards the northeast (shown in blue in Fig. 18, page 34) indicate ice flowed directly into the North Sea Basin (indicated by blue arrows in Fig. 25, page 40). Streamlined bedforms orientated towards the southeast along the NNCP (shown in red in Fig. 18, page 34) indicate that ice flowed around the Cheviots and southwards along the NNCP, parallel to the present-day coastline (indicated by red arrows in Fig. 25, page 40).

The elongation ratios of the bedforms provide an insight into the spatial variations in flow velocities across and along the bedform track, as with increased ice velocities, it is proposed that the length and elongation of bedforms increases (Rose, 1987; Stokes and Clark, 2002a; Briner, 2007). It is therefore suggested that ice velocities were highest along the centre of the subglacial bedform track; the longest bedforms are identified here (> 1000 m in length), with elongation ratios frequently exceeding 1:10, such as around Kelso and Coldstream (see Fig. 16, page 32). At the lateral margins of this track, shorter, less elongate bedforms (see Fig. 18) suggest velocities were reduced here. Along the NNCP, whilst features are less well defined and harder to identify than those in the Tweed Valley, elongation ratios exceeding 1:10 imply ice velocities were high. The highly convergent nature of the streamlined bedforms mapped in the upper reaches of the Tweed Valley (see Figure 15, page 33) implies convergent, and therefore accelerating ice flow (Benn and Evans, 1998), which is in agreement with previous researchers (e.g. Clapperton, 1971; Everest *et al.*, 2005).

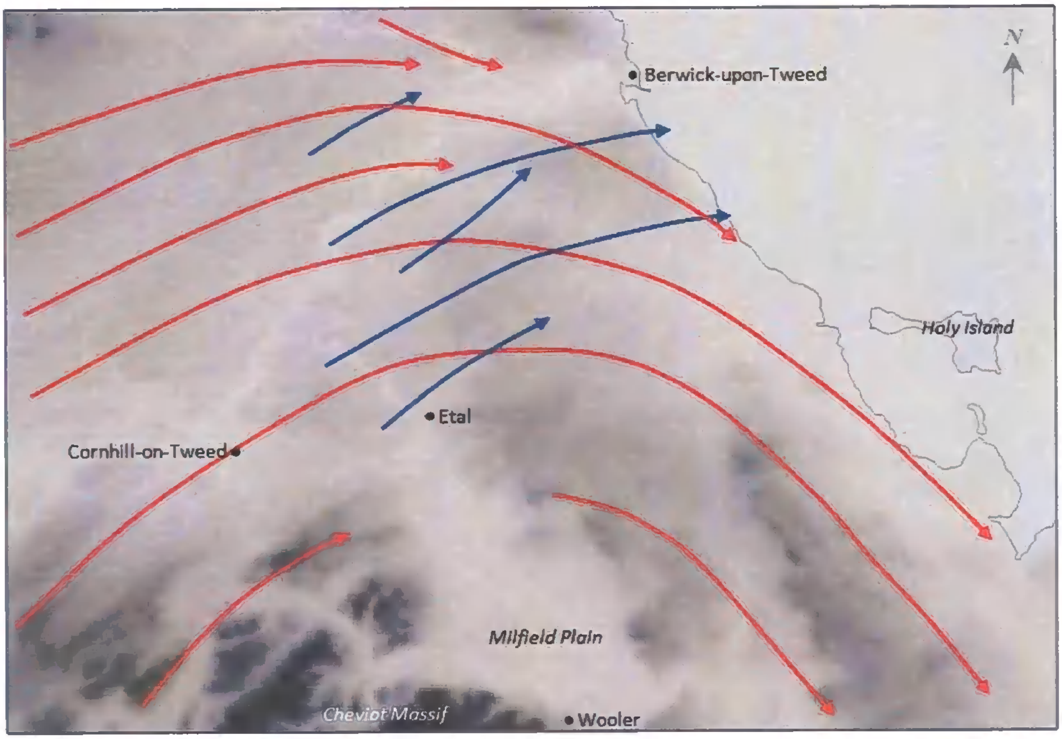


Figure 25: *Simplified trend of the two flow-sets identified in the Tweed Valley and northeast Northumberland. Blue arrows show flow direction inferred from drumlins orientated towards the NE, with the red arrows showing those that are orientated in an arcuate loop around the Cheviot Massif. (Relief-shaded DTM from NEXTmap® Britain, Intermap Technologies).*

It has recently been proposed that extensive fields of streamlined subglacial bedforms are evidence of ice streaming (Stokes and Clark, 2002a). Based on a comparison with the expected geomorphological criteria for the identification of palaeo-ice streams (Table 2), it is proposed that an ice stream flowed along the Tweed Valley and NNCP. The existence of a Tweed Ice Stream (TIS) is based on the identification of five of the eight geomorphological criteria presented by Stokes and Clark (1999). These include its characteristic shape and dimensions, highly convergent flow patterns, highly attenuated bedforms varying in size across and along the track and abrupt lateral margins. This, and the glaciodynamic significance of the TIS is discussed in detail in section 5.1.

Table 2: Geomorphologic criteria for the identification of palaeo-ice streams with evidence for a Tweed Ice Stream (from Stokes and Clark, 1999; Stokes and Clark, 2001; Clark and Stokes, 2001).

Contemporary ice stream characteristics	Proposed Geomorphological signature	Observed features in the Tweed Valley and surrounding area
Characteristic shape and dimension	Characteristic shape and dimensions (>20 km wide, 150 km long) of distinct flow pattern Highly convergent flow patterns leading into a trunk	Drumlin track 35 km wide in places Convergent flow pattern in upper Tweed Valley, trunk visible between Cheviot Massif and Lammermuir Hills
Rapid velocity	Highly attenuated features with elongation ratios >1:10 Boothia-type erratic dispersal train	Elongation ratios <1:12 Not seen
Distinct velocity field	Variation in streamlined bedform sixe	Features shorter and smaller towards the lateral margins, larger, longer and more elongate towards the centre of the track
Deformable bed conditions	Glaciotectonic and geotechnical evidence of deformed till	Not seen
Sharply delineated shear margin	Abrupt lateral margins (<2 km) Ice stream shear margin moraines	Abrupt margins visible, particularly well-define between the Cheviot Massif and Lammermuir Hills
Spatially focussed sediment delivery	Submarine till delta or sediment fan	Ice stream terminus not identified – inferred to be offshore (e.g. Everest *et al.*, 2005)

4.2 Assemblage B: Hummocky Terrain

At the lateral margins of the TIS track, three distinct extensive areas of hummocky terrain have been identified. Hummocks are mapped as black lines in figure 10 (page 29). Landforms within these three complexes include small regular-shaped mounds, through larger irregular mounds to elongate sinuous ridges. All are found in close association spatially.

4.2.1 The Wooler Complex

4.2.1.1 Description

The first extensive area of hummocky terrain is located south of Wooler in the lower ground between the Cheviot massif to the east and the Fell Sandstone scarp to the west. The system can be traced for 9 km as far south as the River Breamish, just north of Ingram and Powburn (Fig. 42). The Wooler Complex (Fig. 27, page 43) consists of near-circular mounds 5-15 m high;

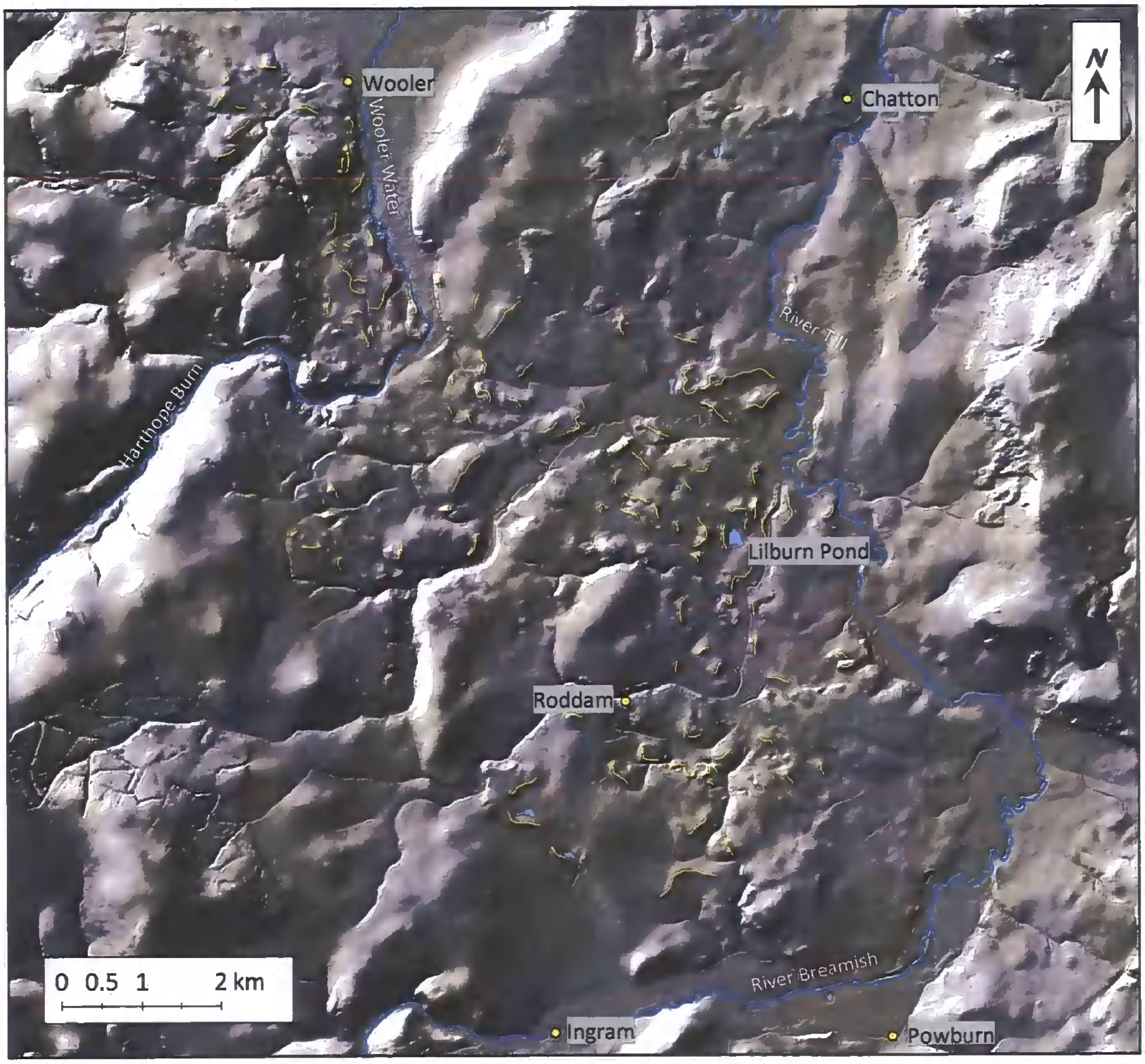


Figure 26: *NEXTmap DEM of Hummocky terrain south of Wooler and locations mentioned in description. (DTM NEXTmap® Britain, Intermap Technologies, 2008).*

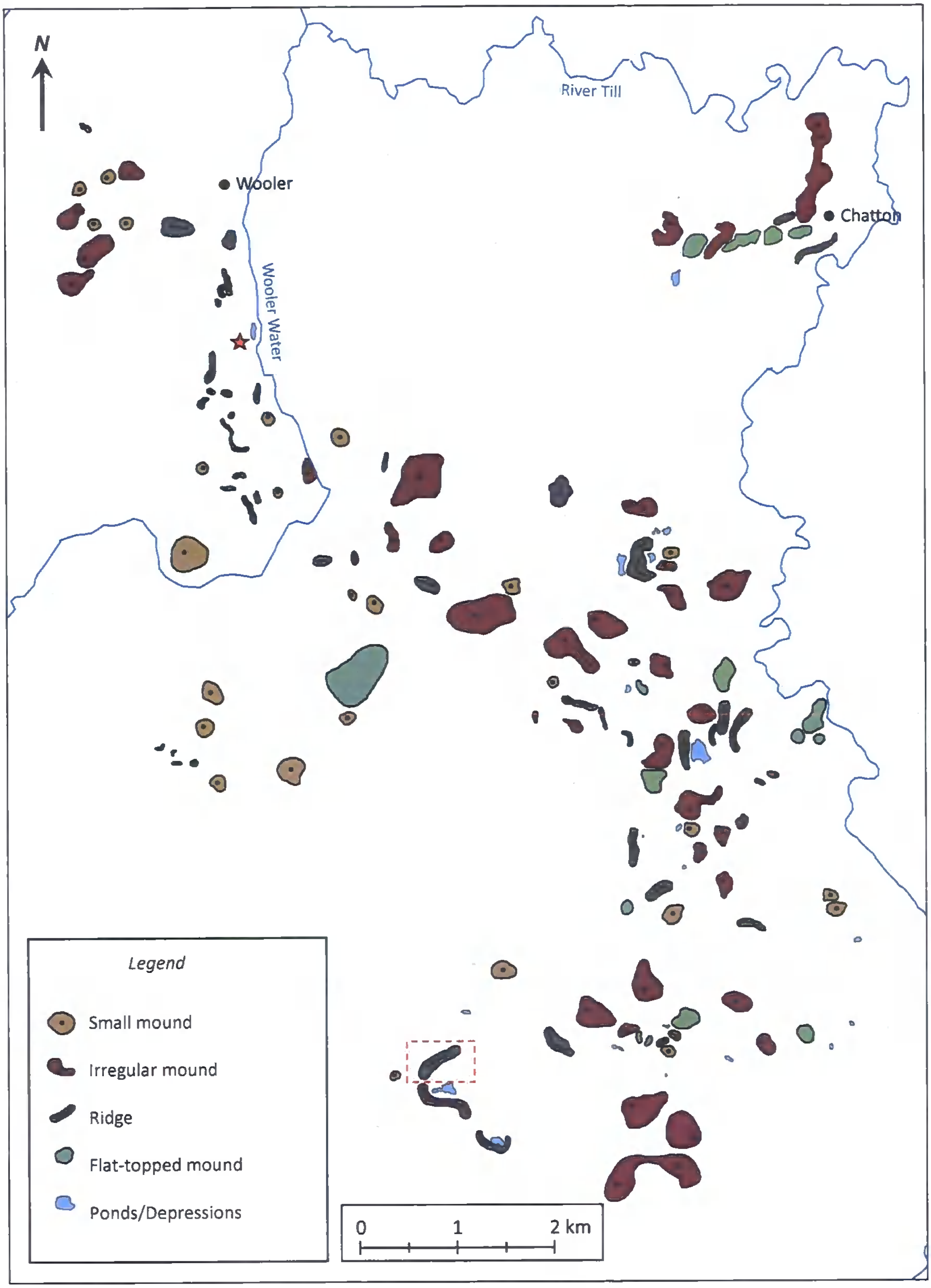


Figure 27: *The Wooler Hummocky Terrain Complex. Star shows location of Figure 33, page 50. Inset box shows location of Roddam Bog – site of sedimentary survey.*

sinuous ridges 100 to 700 m long and irregular larger mounds, often with a ridge extending from a high point. These larger mounds are up to 30 m in height. Gently-sloping flat-topped mounds of varying size are also found, 10-12 m in height. In many instances, small depressions occur between these hummocks and ridges. Several of these depressions are occupied by ponds or marshy ground, such as at Roddam and Lilburn Pond. There appears to be little arrangement or pattern to the distribution of the mounds, although the sinuous ridges do exhibit some degree of alignment locally; 1.5 km south of Wooler a series of sinuous ridges trend approximately N-S, as do those around Lilburn Pond. These ridges, the crests of which frequently undulate, show little alignment with the topography. A site 1 km southwest of Roddam at Roddam Bog provides an insight into the internal stratigraphy and sedimentology of these mounds. At this site, three gently-sloping broad ridges enclose two small ponds (inset box in Fig. 27, page 43). The northernmost ridge has been excavated and forms a 55 m long, 3.5 m high southeast facing exposure (Fig. 28, page 45 and Fig. 29, page 46). From west to east across section, the first 31 m are characterised by interbedded coarse sands and gravels, with sharp erosional contacts. At the base of the exposure, beige/brown laminated, horizontally-bedded and massive sands are identified (Sl, Sh, Sm). These beds are trough-shaped and are overlain by a laterally extensive bed of poorly-sorted, matrix-supported beige/grey gravel (Gms), identifiable along this entire stretch of the exposure. In places, this bed is overlain by pockets of coarse brown sand with outsize well-rounded pebbles (Sm). At 31 m along the exposure, Log 1 was drawn (Fig. 30, page 47).

The lowest bed (Bed 1) is 40 cm thick and consists of poorly-sorted, steeply-dipping planar cross-bedded coarse sands (Sc) interbedded by laminated, poorly-sorted coarse sands (Sl). Bed 2 is 10 cm thick and is formed of up-fining gravels (Gfu). Bed 3 (10 cm thick) is formed of massive granules with isolated outsize clasts (GRmc). This is overlain by Bed 4, which is comprised of 12 cm of up-fining gravels (Gfu). Bed 5 is composed of up-fining gravels (Gfu) in a coarse yellow/grey silty-sand matrix. Measurements of 50 clasts from here (Fig. 30b, page 47) indicate clasts are predominantly blade shaped and around 35 % of the total sample are sub-angular (SA). Clasts were not separated on the basis of their lithologies. The C_{40} value, which measures the percentage of clasts in a sample that exhibit c:a axis measurements
and Benn, 1998), is 44.4. The uppermost bed (Bed 6) is the matrix-supported beige/grey gravel bed (Gms), visible for around 30 m along the exposure. Laboratory particle-size analysis (Appendix E) reveals the matrix is composed of poorly sorted medium silty coarse sand with a high proportion of fines (fine skewed). The contacts between all 6 beds are sharp (erosional).

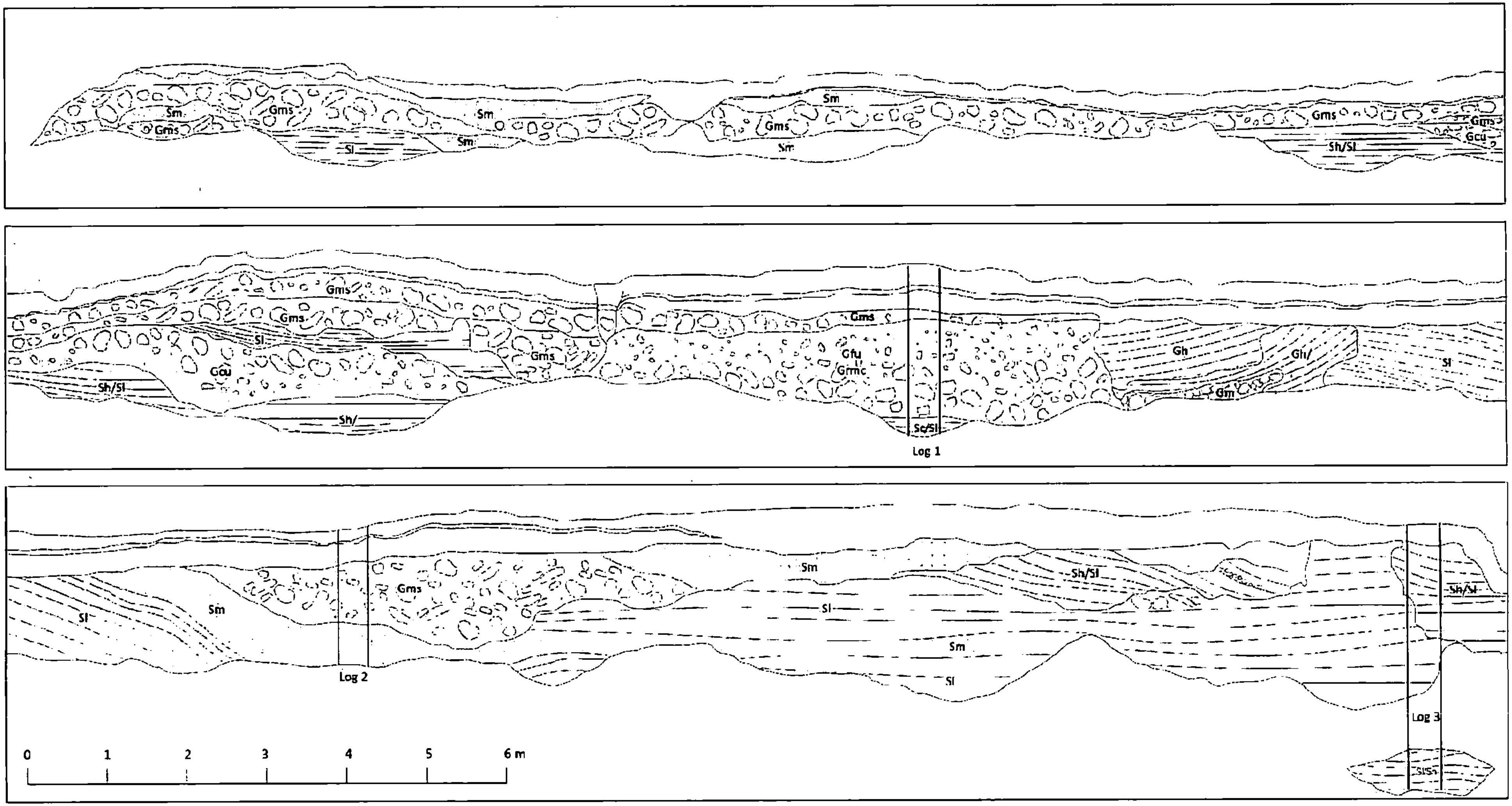


Figure 28: *Scaled sketch of the Roddam Bog Exposure. For key to lithofacies codes see Appendix A.*

Figure 29: *The Roddam Bog Exposure.* ***TOP:*** *Central section of Roddam Exposure, site of log 2 (Fig.31, page 48).* ***BOTTOM:*** *East end of section, site of log 3 (Fig. 32, page 48. Note fining towards the east.*

Log 2 (Fig. 31, page 48) is formed of 45 cm of steeply dipping, highly compacted massive and ripple cross-laminated beds (Beds 1-6) of medium to fine sand (Sm, Sr(B)), overlain by a 75 cm-thick trough-shaped bed of matrix-supported gravels. Morphological analysis of 50 clasts from this bed (Bed 7) indicates that clasts are predominantly elongate, compact blade- shaped or blade-shaped and are mainly sub-rounded and sub-angular (Fig. 31b, page 48). Clasts were not separated on the basis of their lithologies. The C40 value is 42. Log 3 (Fig. 32, page 48), was recorded from the eastern end of the exposure where the sediment in finest (see Fig. 29, bottom photo). The lower bed, Bed 1, is comprised of 15 cm of ripple cross-laminated sands (Sr(B)). Palaeocurrent readings from these ripples indicate flow was predominantly from the northwest. Above this bed is a 45 cm-thick, highly-consolidated bed of up-coarsening sand and massive fines (Fm/Suc). Statistically, the proportion of mud and sand in the lower section of this bed is 85.1 % and 14.9 % respectively. In the upper coarser section, proportions of mud drop to 21.1% and sand rises to 78.9%. Bed 3 (160 cm thick) is comprised of alternating layers of massive, moderately-sorted coarse silty-sand, interdigitated with ripple cross-laminated fine sands. Over 99% of the total weight of the samples taken from this bed were sieved below 2 mm in diameter. Above this bed is a 20 cm-thick bed of moderately-sorted medium sands, exhibiting trough cross-bedding and horizontal bedding. The uppermost bed, Bed 5, is comprised of cross-bedded granules in a moderately-sorted, coarse sandy matrix (Grt/St). Clast lithological analysis on a sample of 20 clasts from this bed reveal porphyritic andesites, granite, carboniferous chert and quartzite sandstone.

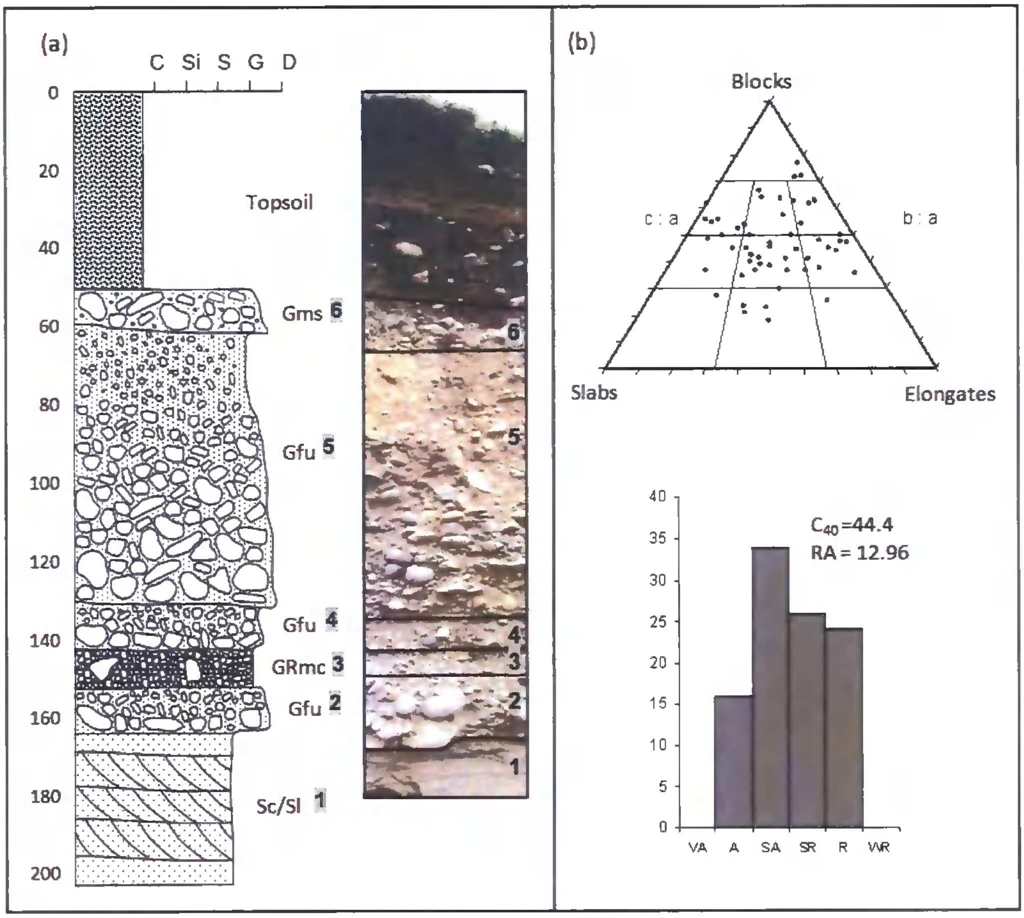


Figure 30: *(**a**) Roddam Bog Log 1. Location of log shown in figure 28, page 45. For lithofacies code see Appendix A. Numbers highlighted in grey denote bed number. For descriptions of beds see text (**b**) Clast roundness and shape data from Log 1, Bed 6.*

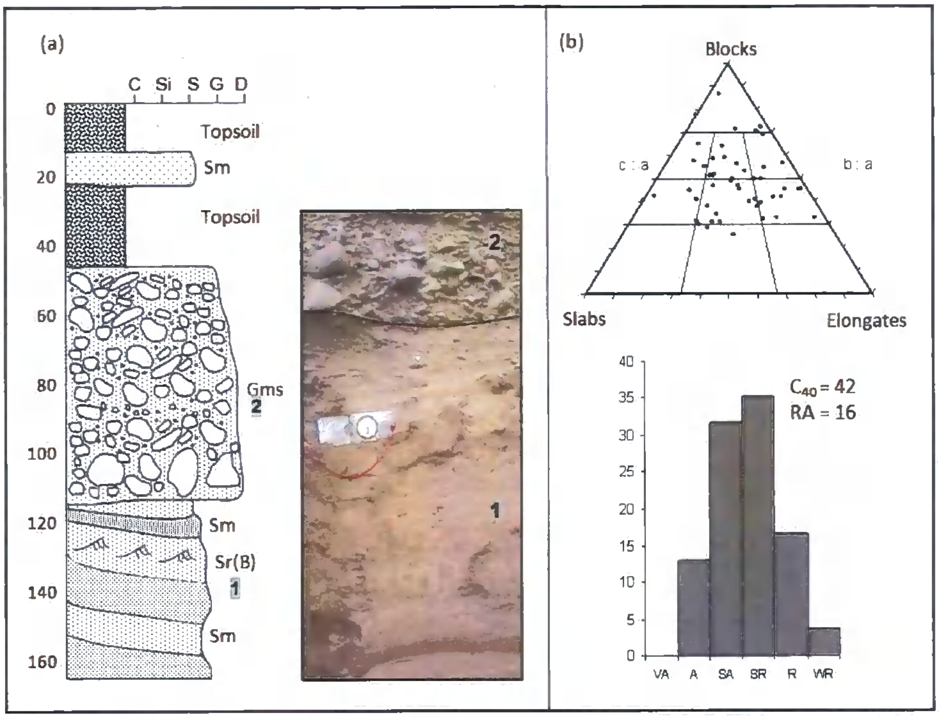


Figure 31:* *(a) *Roddam Bog Log 2. Location of log shown in figure 28, page 45. For lithofacies code see Appendix A. Compass clinometer for scale (approx. 15 cm long). Numbers highlighted in grey denote bed number.* ***(b)*** *Clast roundness and shape data from Log 2, Bed 2.*

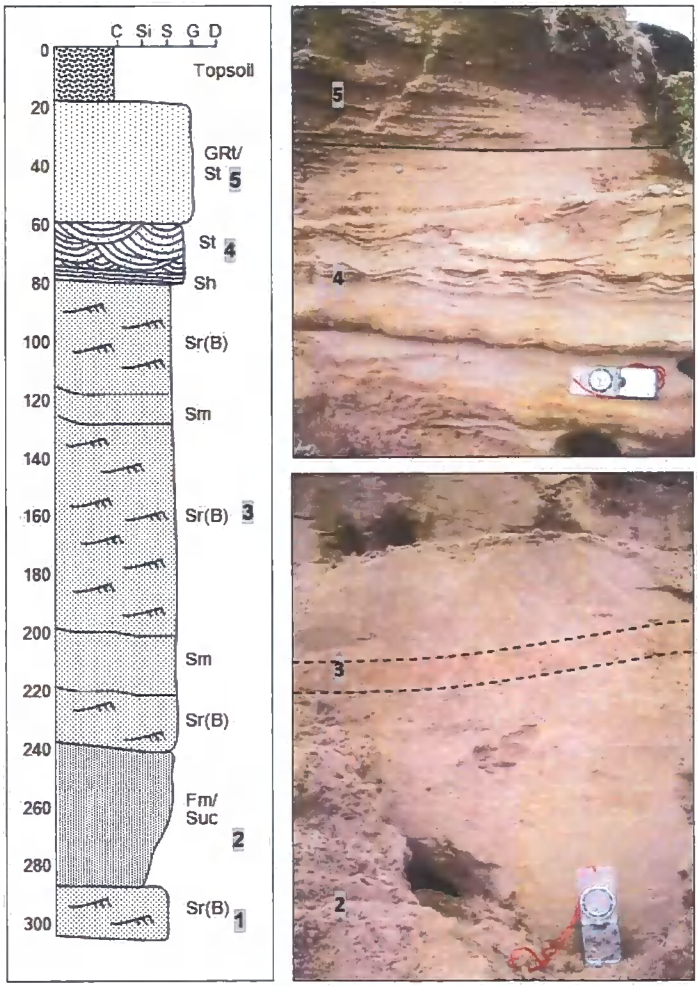


Figure 32: *Roddam Bog Log 3. Location of log shown in figure 28, page 45. For lithofacies code see Appendix A. Numbers highlighted in grey denote bed number.*
Top photo: *Upper two beds of Log 3. Note well-defined ripple features. Compass clinometer for scale (approx. 15 cm long).*
Bottom photo: *up-coarsening fine sands and silty-sand bed (highlighted with black dashed lines). Compass-clinometer for scale (approx. 15 cm long).*

4.2.1.2 Interpretation

The close spatial association and morphology of the sinuous ridges, hummocks, flat-topped mounds and small depressions south of Wooler has led to the interpretation that these sand and gravel features are part of the same landform assemblage. Sinuous ridges are interpreted as eskers. This interpretation is based on their sinuous long-profiles and apparent lack of alignment with the underlying topography, which is indicative of deposition within pressurised conduits (Bennett and Glasser, 1996). Those eskers showing little alignment to the underlying topography (e.g. south of Wooler) may be the result of the superimposition of englacial conduits, or may reflect deposition in subglacial tunnels (Warren and Ashley, 1994; Martini *et al.*, 2001). It is proposed that these eskers formed within active ice, as esker formation is inferred to be limited in areas of stagnating, slow-flowing ice (Punkari, 1997). Furthermore, in order to maintain the cryostatic pressure required to keep the conduits open, actively flowing ice (i.e. not stagnating) would have been required (Warren and Ashley, 1997). Shorter, closely-spaced ridges that are orientated at oblique angles to the main N-S trend of the eskers, such as those south of Wooler and east of Roddam, may represent bifurcated eskers (c.f. Evans and Twigg, 2002) that have been incised by post-glacial streams. The inferred sedimentology of these sinuous ridges supports the morphological interpretation that they are eskers; Clapperton (1971b) identified sinuous ridges composed of well-sorted, stratified, sands and gravels within the Wooler Complex and associated them with subglacial deposition, or englacial channel superimposition (see section 2.2.1). The undisturbed nature of the bedding implies *in situ* deposition and as such, earlier interpretations of these features as the product of melt-out and ice stagnation (c. f. Carruthers *et al.*, 1930, 1932) are rejected. Irregular mounds and small, near-circular mounds throughout the Wooler Complex are tentatively interpreted as supraglacial kames and hummocky moraine. The overall appearance of these landforms is relatively chaotic; there appears to be no accordance with the underlying topography. Kames and hummocky terrain have been differentiated on the basis of their morphology; kames exhibit smoother sides and are more gently sloping than hummocky moraine (Benn and Evans, 1998). This interpretation is somewhat tentative, as the mounds are likely to have been affected by slumping and post-glacial reworking. The lack of sedimentary exposures has also hindered their interpretation. However, through a comparison with morphologically similar features elsewhere, their mode of formation can be speculated upon.

Flat-topped mounds mapped in close association with the kames and eskers are interpreted as ice-contact fans and ice-proximal outwash deposits. Regarding the internal structure of these, the sedimentological survey conducted at Roddam (see Figs. 30-32, pages 47-48) has revealed much on the environment of deposition. The vertical and horizontal variations in the sedimentary characteristics of this exposure suggest a spatially and temporally changing flow regime. The abundance of subangular clasts suggests that the streams responsible for deposition were not particularly far travelled, whilst the elongate, blade-shape of the clasts suggests they may have been sourced subglacially (Evans and Benn, 2004). The variations in grain size across the exposure reflect changes in sediment supply and/or changing flow conditions. From west to east, the overall trend is that of sediment fining, which indicates that either volume of coarse sediment available for transportation and deposition was reduced, or that flow velocities were decreasing (Reading, 1996; Russell and Marren, 1999). Climbing ripples (i.e. ripple cross-lamination) with drapes within beds 1 and 3 of Log 3 (Fig. 32, page 48) are indicative of bottom currents switching on and off with fallout of suspended sediment following ripple formation (Ashley *et al*, 1981). The horizontally-bedded and laminated sands towards the east of the exposure are indicative of low energy deposition in relatively calm water, possibly ponded in cut-off channels or flooded depressions. It is therefore suggested that these changes in particle size reflect a shift from a proximal to intermediate environment of deposition (Evans and Benn, 2004). The interbedded sand and gravel troughs and prograded channel fills are therefore interpreted to represent a proglacial braided stream environment. Palaeocurrent data suggest these streams were flowing from the north-west, i.e. from the Cheviot massif.

Figure 33: *Hummocky terrain south of Wooler. Note gently-undulating hummocks. Location shown by red star in figure 27, page 43. View is towards the southeast.*

Depressions between the eskers, kames and fans are interpreted to be kettle holes, which are formed by the gradual melt-out of ice buried by glaciofluvial sediment. Kettle holes are frequently observed on contemporary outwash plains, such as at Breiðamerkurjökull often in close association with other glaciofluvial landforms (c.f. Evans and Twigg, 2002; Everest and Bradwell, 2003). The form of kettled surfaces depends greatly on the thickness of sediment cover, the organisation of the proglacial drainage system and the rate at which buried ice melts (Bennett and Glasser, 1996; Evans and Twigg, 2002). It is therefore suggested that the occurrence of kettle holes within the Wooler Complex implies that stagnating ice was present throughout the complex and that proglacial drainage was limited or that meltwater channels shifted their courses prior to ice stagnation. Where this happened, the melt-out of buried ice has resulted in inversion of the relief (c.f. Bitinas *et al*, 2004). Postglacially, these have been water-filled, such as at Roddam Bog and Lilburn Pond (see Figs. 21 & 22, pages 38 & 39). It is suggested here that the Wooler Glaciofluvial Complex formed during deglaciation when the ice was undergoing active retreat. The large spatial coverage of glaciofluvial sediments and landforms (eskers, braided channel systems, hummocky moraines, fans and kames) indicate that vast volumes of water were present, a condition known to exist during deglaciation. During the final stages of deglaciation, reduced meltwater flow and stagnating ice led to the formation of kettle holes. The association of the Wooler Complex with deglaciation is in agreement with Clapperton, 1971b, who proposed that the glaciofluvial phenomena in the area around Wooler were deposited during the first phase of deglaciation. This was followed by separation of the ice masses and the formation of glacial lakes and sand and gravel delta deposition (Clapperton, 1971b). The broader implications of this complex are considered in the next chapter (section 5.3).

4.2.2 The Cornhill-on-Tweed Complex

4.2.2.1 Description

The second extensive region of hummocky terrain is located to the south of Cornhill on Tweed at the southern margin of the Tweed drumlin field. The complex stretches from Pressen Hill to Etal, 10 km to the northeast (Fig. 34, page 52). The topography is characterised by smoothly undulating mounds (Fig. 35, page 52). Similar features to those identified in the Wooler Complex are found here, i.e. sinuous ridges, irregular mounds, near-circular mounds, plateau-topped mounds (Fig. 36, page 53). The features south of Cornhill are, however, less chaotically spread than around Wooler. Sinuous ridges are found mainly towards the centre of the complex, between Cornhill and East Learmouth, and range in length from 125 to 500 m. The long axes of these features are orientated predominantly E-W and NE-SW. To the east, between East Learmouth and Wark, and to the west between East Learmouth and Branxton are groups of irregular mounds (15 to 24 m high) and small near-circular mounds (5-15 m high). To the west of East Moneylaws are two elongate, gently-sloping plateau-topped mounds running along the lower flanks of Moneylaws Hill. Around Crookham, similar plateau-topped mounds are found. The largest of these is 25 m high, over 1 km long and 1 km wide. As in the Wooler complex, small depressions and ponds are found between the hummocks. It has been possible to infer the internal composition of the Cornhill features in the main complex by overlaying the landform map onto the BGS surficial geology maps (Geological Map Data © NERC 2008) (Fig. 37, page 54). The vast majority of features are composed of glaciofluvial sands and gravels, whilst some, particularly those to the southwest around Pressen Hill are composed of Till. Between the mapped features are spreads of glaciofluvial sand and gravel, which can be traced to the southeast into the Glenn Valley, south of Moneylaws Hill. This material runs in discontinuous ridges and mounds along this valley.

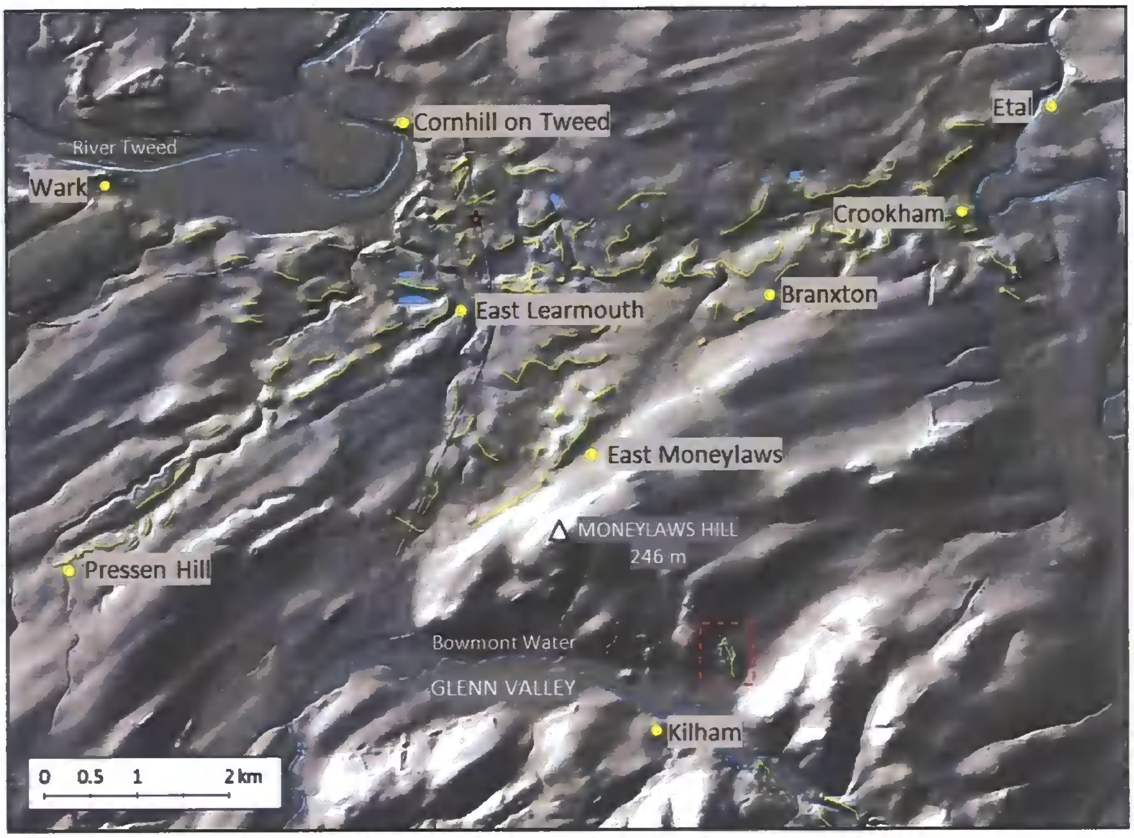


Figure 34: *NEXTmap DEM of the Cornhill Hummocky Terrain Complex. Red inset box shows site of sedimentological survey in the Glenn Valley.*

Figure 35: *Undulating topography of the Cornhill complex. View is south from NT 864 382 (marked by red star in Fig. 28 and 30).*

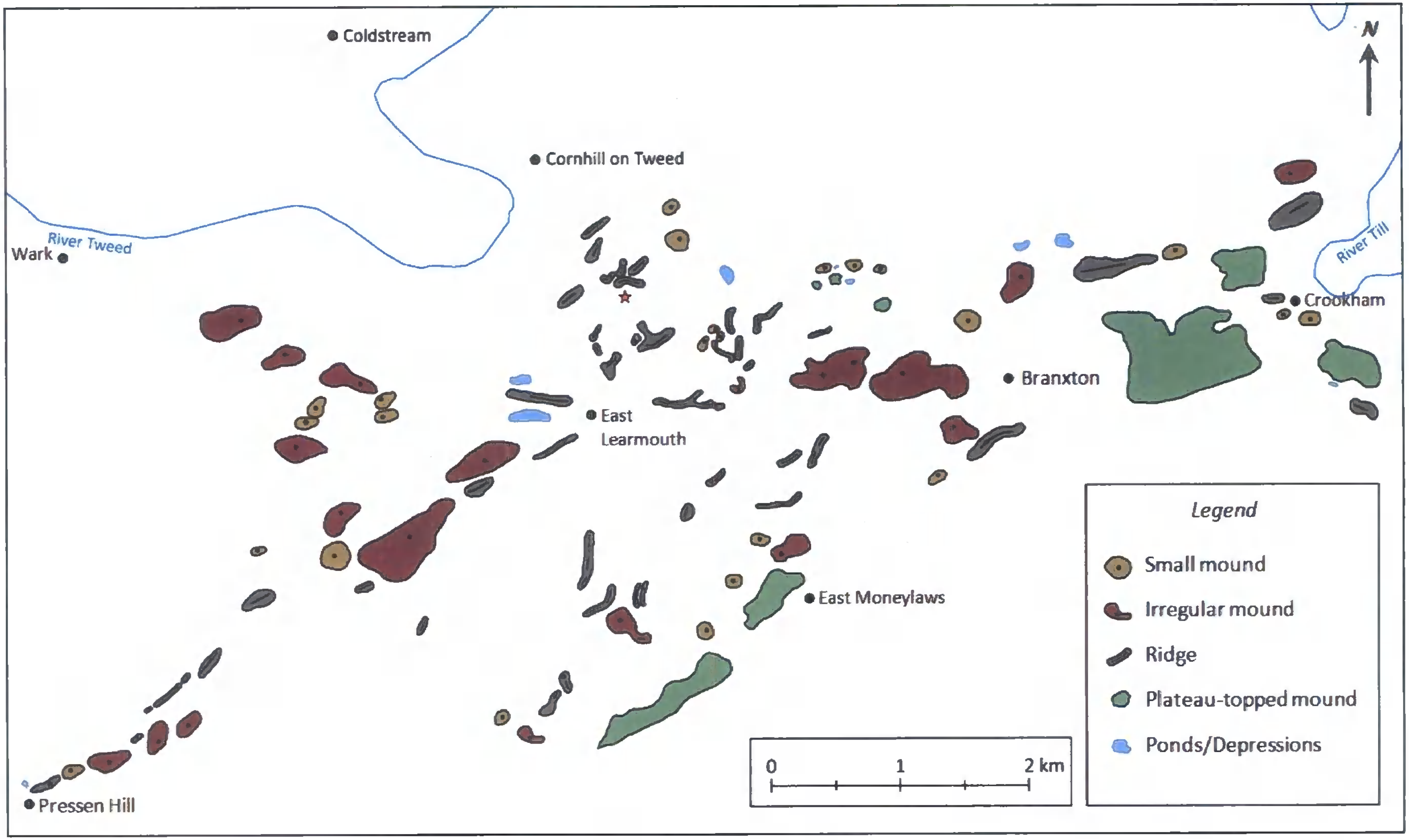


Figure 36: *The Cornhill Hummocky Terrain Complex.*

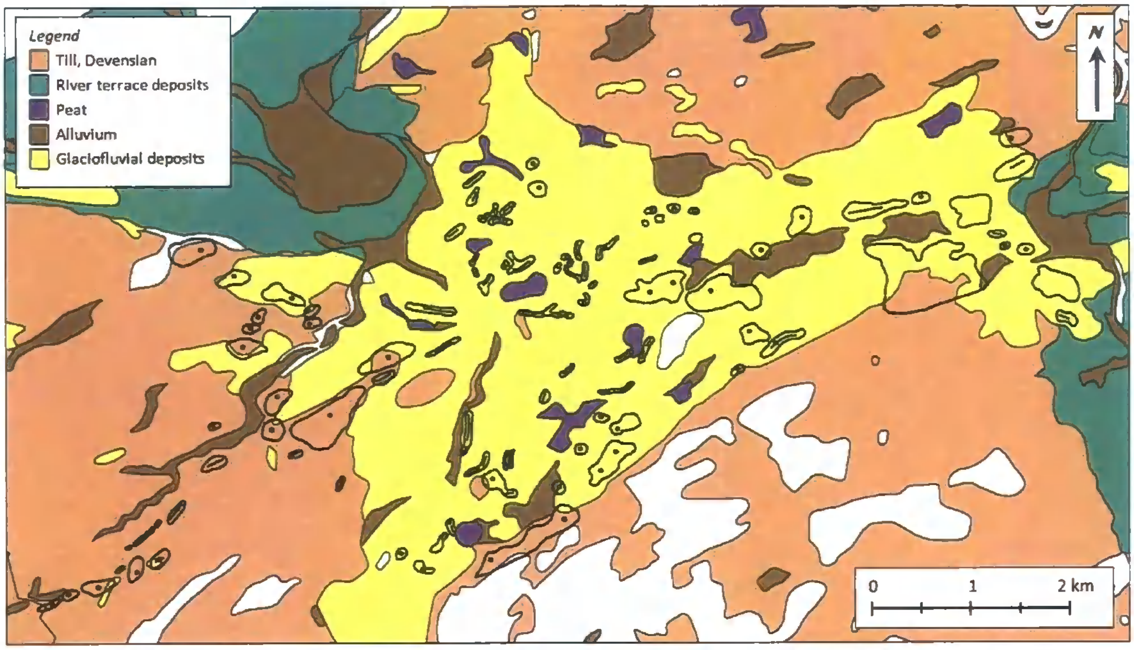


***Figure 37**: Surficial Geology of the Cornhill Hummocky Terrain Complex. (Geological Map Data © NERC 2008)*

In a NE-SW trending valley that joins the Glenn Valley at Kilham (NT 885 325), four logs were recorded at a 5m high southwest-facing exposure (Fig. 38, page 55). The results from this sedimentological investigation reveal sub-parallel beds of inversely and normally graded gravels, imbricated gravels and horizontally and trough cross-bedded granules and sands, dipping towards the southeast at an angle of around 15-20°. Log 1 (Fig. 39, page 55), at the northern end of the exposure exhibits overall up-fining. The lower bed of Log 1 is 160 cm thick and is comprised of matrix-supported massive and imbricated gravels and horizontally-bedded gravels. Clast shape analysis show sub-angular and sub-rounded clasts dominate this bed. Clasts are imbricated on their a-axes by an average of 5°, orientated N-S. Overlying this are 40 cm of interbedded massive sand and matrix supported gravel (Beds 2-6), 10 cm of highly compacted massive granules with large (>15 cm diameter) outsize clasts (Bed 7), 40 cm of alternating horizontally-bedded sand and granules (Bed 8), 10 cm of massive sand (Bed 9), 20 cm of up-fining gravels (Bed 10) and 10 cm of horizontally-bedded highly compacted sand (Bed 11).

The beds of log 2 (Fig. 39, page 55) are tilted by around 15° towards the southeast. The lowest bed is comprised of 20 cm of massive clast supported granules and is laterally continuous with Bed 7 of log 1. Overlying Bed 1 is 15 cm of horizontally-bedded sand (Bed 2), 15 cm of up-coarsening interbedded sand and gravels (Bed 3), 11 cm of horizontally-bedded sand with pockets of outsize granules (Bed 4), 45 cm of up-fining gravels and granules (Beds 5-7) and 15 cm of horizontally-bedded sand (Bed 8). This upper bed, unlike the beds beneath it, is not inclined and shows a sharp erosional contact between it and bed 7. Log 3 (Fig. 39) is approximately 2 m to the south of Log. The lower 8 beds are no thicker than 7 cm and are comprised of, from the base up, horizontally-bedded sand, massive granules, horizontally-bedded sand, massive sand, up-fining granules, massive sand, up-coarsening granules and horizontally-bedded sand.

Figure 38: *Sedimentary logging site in the Glenn Valley. Log sites are shown with white lines.*

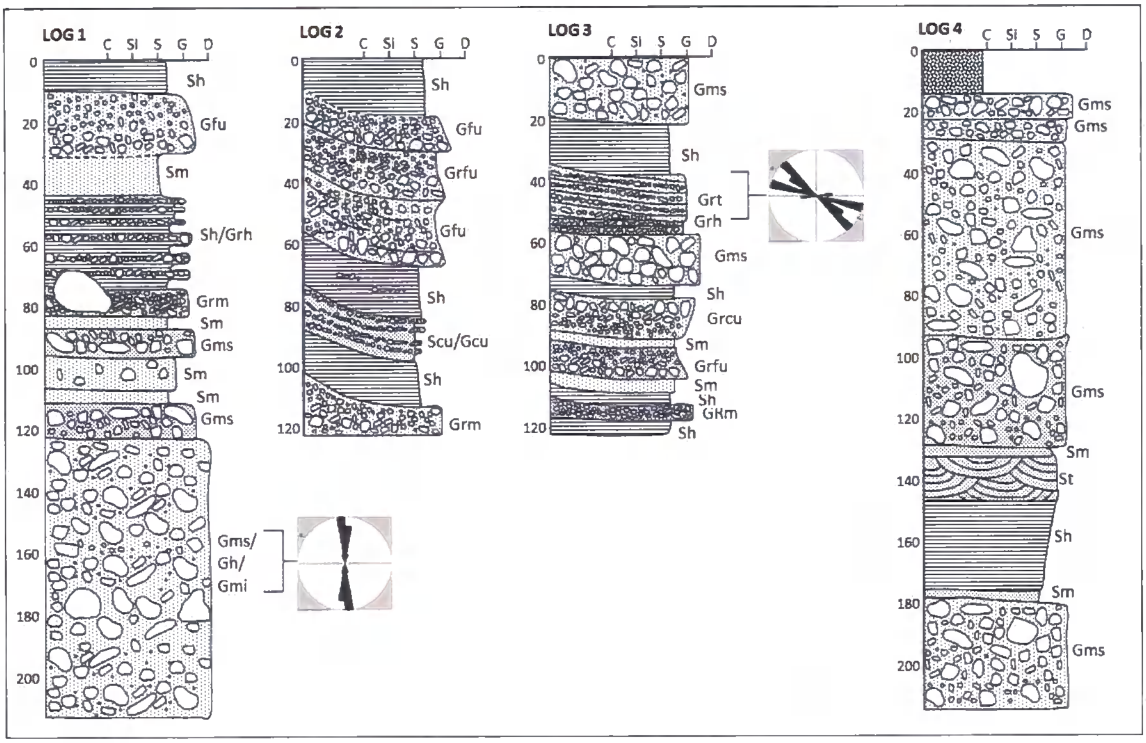


Figure 39: *Sedimentary logs from exposure in the Glenn Valley near Kilham. Location of logs in relation to each other shown in Fig. 38. Rose diagrams show clast orientation.*

Overlying these lower beds is 10 cm of matrix-supported gravels (Bed 9) and horizontally-bedded and trough cross-bedded granules (Beds 10 and 11). Clasts within these beds dip by an average of 14° NW-SE and are predominantly sub-angular. The upper two beds are comprised of 17 cm of horizontally-bedded sand (Bed 12) and gravels in a coarse sandy matrix (Bed 13). Log 4 (Fig. 39) at the southern end of the exposure is the highest log recorded and, unlike the other three logs, the beds of it do not dip and are instead plane bedded. The lower bed is comprised of 35 cm of matrix-supported gravels, overlain by 50 cm of massive and horizontally-bedded sand, trough cross-bedded sand and massive sand (Beds 2, 3, 4 and 5 respectively). These are overlain by 4 beds of matrix-supported gravels, displaying an overall up-coarsening (beds 6-9). The matrix of these upper four beds is medium to coarse sand (Appendix E).

4.2.2.2 Interpretation

In terms of morphology, the landforms of the Cornhill Complex are very similar to those identified at Wooler (see section 4.2.1). Based on their comparable morphologies, it is interpreted that this complex formed in much the same way. Sinuous ridges are, like at Wooler, interpreted as eskers. To the southwest of the Cornhill Complex, these are orientated parallel to flow of the TIS, as is inferred from their relationship with the adjacent NE-orientated streamlined bedforms. These eskers are suggested to be the remnants of superimposed englacial conduits, as they are found draped over the drumlins mapped in this area. Sinuous ridges identified just south of Cornhill-on-Tweed are also interpreted as eskers. However, their shorter lengths, bifurcating nature and near-parallel orientation to adjacent eskers suggest that the conduits in which they were deposited shifted over time (c.f. Shreve, 1985) or that deposition was not continuous (Brennand, 1994). Water-filled depressions mapped between the eskers are interpreted as kettle holes, which indicate stagnating ice was present within this complex (c.f. Evans and Twigg, 2002; Everest and Bradwell, 2003). Irregular and near-circular shaped mounds mapped west of East Learmouth and west of Branxton are interpreted as supraglacial kames. This interpretation is based on their morphology and inferred sedimentology; these mounds exhibit highly disturbed bedding (Carruthers *et al.*, 1932) which is a diagnostic criteria of supraglacial kames as bedding structures are disturbed by the melt-out of the surrounding supporting ice.

Flat-topped mounds on the lowest flanks of Moneylaws in the Cornhill-on-Tweed Complex are interpreted as ice-contact kame terraces. This interpretation is based primarily of the morphology and location of these mounds. Kame terraces form through glaciofluvial deposition in ice-marginal lakes (Bitinas *et al.*, 2004). The terraces at Cornhill are therefore inferred to have formed between the TIS and Moneylaws Hill as the TIS retreated (Fig. 40, page 57). Their poorly defined form suggests the ice margin was relatively shallow, resulting in a large area of ice being covered glaciofluvial and glaciolacustrine sediments (Bennett and

Glasser, 1996). Their undulating surfaces are attributed either with the stagnation of this ice from beneath the supraglacial deposits or from post-depositional disturbance.

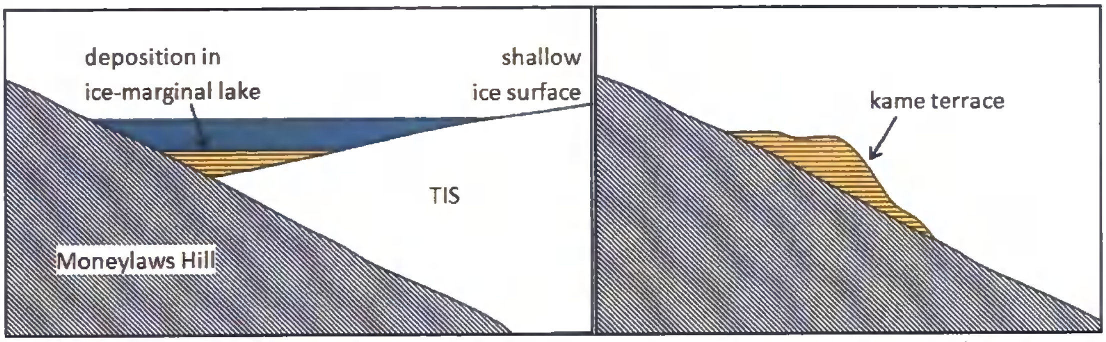


Figure 40: *Cartoon of inferred formation of kame terraces of the Cornhill Complex.*

The sedimentological survey conducted at Kilham (see Fig. 39, page 55) reveals the palaeo-environmental conditions of the Glenn Valley during deglaciation. It is proposed that this plug of glaciofluvial sediment is an ice-contact delta. This interpretation is based a comparison with the sedimentary characteristics of ice-contact deltas identified elsewhere (c.f. McCabe and Eyles, 1988; Lønne, 1993, 1995) and also on the interbedded, large-scale clinoforms of sub-parallel sands and gravel dipping towards the southeast, which represent foreset bedding and are indicative of a prograding depositional system (Lønne, 1995). The coarse lower bed of Log 1 is interpreted as re-sedimented glacial till, which is proposed on the basis of the high proportion of sub-angular and sub-rounded bullet-shaped clasts within this bed. Diamict facies in a Northern Irish ice-contact glaciomarine delta are attributed with subaqueous debris flows (McCabe and Eyles, 1988). It is suggested that this lower bed may have been deposited by a debris flow, as clasts are orientated with their a-axes parallel flow, with the a-axis imbricated upstream, a characteristic of mass flows (Evans and Benn, 2004). The presence of debris flow material reflects the proximity of Log 1 to the glacier margin (c.f. Nemec *et al.*, 1999). The lower 7 beds of Log 2 and all of the beds of log 3 are interpreted as foresets, constructed from reworked glaciofluvial outwash as glacial streams discharged into the proglacial lake. The upper bed of Log 2 and all the beds of Log 4, which are horizontally-bedded, represent the stream-deposited topset (Lønne, 1995). Palaeocurrent data from the bottomset and foresets imply that flow from the ice margin shifted from N-S to NE-SW, possibly reflecting a reorganisation of the ice margin during retreat. Given the palaeocurrent data, it is speculated that this delta formed in a lake at the TIS margin that flowed over Moneylaws Hill. The lake is hypothesised to have been dammed by an ice lobe that blocked the Glenn Valley (c.f. Lunn, 1995).

4.2.3 The Bradford Complex

4.2.3.1 Description

The third area of hummocky terrain is located in a narrow band southwest of Bamburgh on the North Northumberland Coastal Plain. This complex has been mapped for over 10 km from Spindlestone to Rock Moor House (Fig. 41). No sites were located for sedimentary analysis and therefore internal composition is inferred from BGS surficial geology maps (Geological Map Data © NERC 2008). Features identified here are primarily ridges (Fig. 42, page 59), varying in height, sinuosity, width and length (Fig. 43, page 60). The majority of these ridges are orientated N-S, apart from that directly south of Spindlestone, south of Newham and west of Doxford Farm. To the east of Spindlestone and Bradford, a near-continuous, slightly sinuous narrow ridge, approximately 10 m high, extends N-S for 1.4 km. Small gaps no longer than 100 m punctuate this ridge.

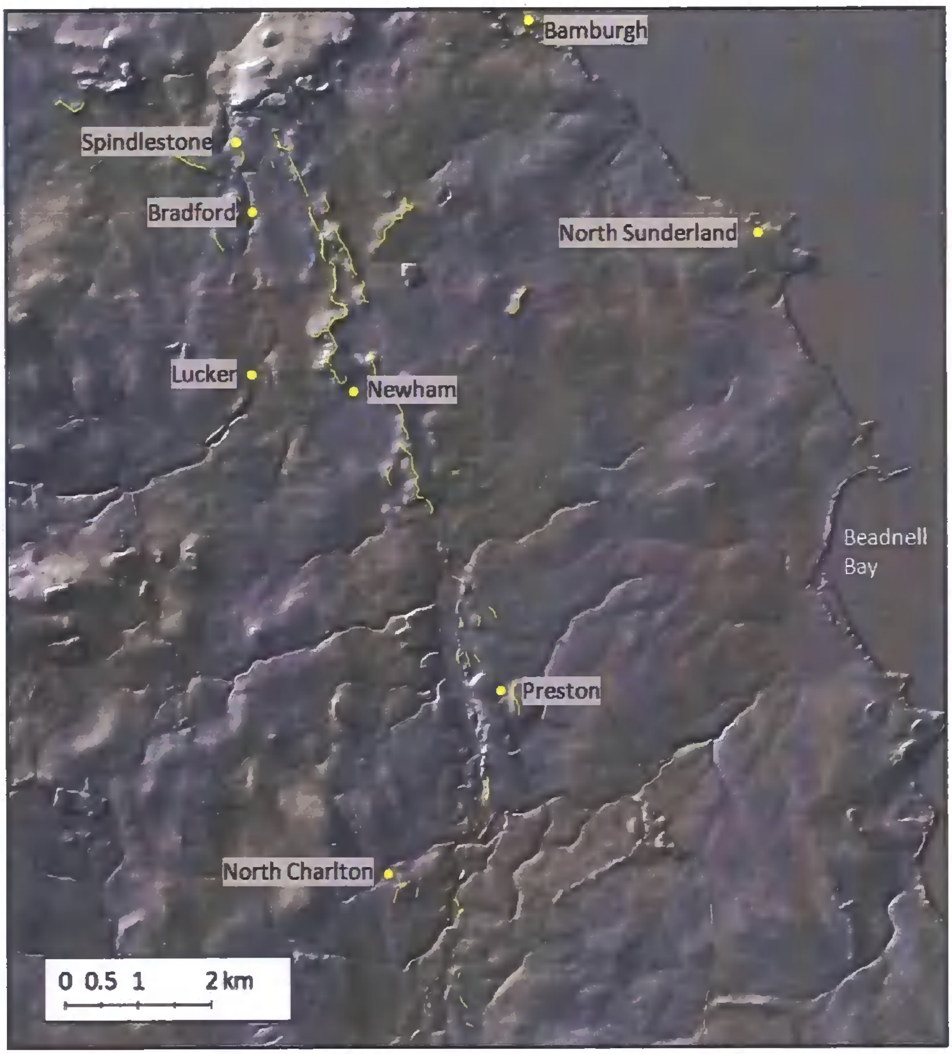


Figure 41: *NEXTmap DEM of the Bradford Hummocky Terrain Complex showing locations mentioned in text. Hummocks highlighted in yellow.*

Figure 42: *Sinuous ridge and pond in small depression. View east towards Spindlestone. Location of photograph shown by red star in figure 43, page 60.*

This ridge is found in close association with small water-filled depressions. At the southernmost point of this ridge at Hoppen Hill, an irregular, fan-shaped, sloping flat-topped mound, measuring over 1 km in length and 500 m wide spreads towards the south. Between Hoppen Hill and Newham, ridges are wider (250 m) and higher (up to 20 m) than the near-continuous ridge to the north. These slightly sinuous ridges have uneven lateral slopes and are inferred to be composed of both till and glaciofluvial deposits (Geological Map Data © NERC 2008). Interspersed among the elongate ridges are a small number of closely-spaced irregular mounds and plateau-topped mounds. Between the hummocks and ridges are spreads of sands and gravels, up to a kilometre in width between Bradford and Chathill. South of Chathill this widens to around 1.5 km. The sand and gravel spreads can be traced as far south as Alnmouth, where they cover an area of approximately $13.5km^2$. Features at the southern end of the complex are generally smaller, less well defined and more widely spaced. The largest feature here is a ridge 20 m high, 250 m wide and 550 m long found just east of Preston. Four small parallel ridges, <250 m long and 10 m high are located between Chathill and Preston, all aligned N-S.

4.2.3.2 Interpretation

Morphologically, the sinuous ridges, irregular hummocks and flat-topped mounds of sand and gravel are similar to those mapped in the Cornhill and Wooler complexes. However, in terms of the overall organisation of the landforms, the Bradford Complex differs, in that the landforms are distributed in a elongate corridor along the NNCP (see 43, page 60). From the orientation of the adjacent elongate bedforms (see Fig. 14, page 32), it is proposed that the Bradford Complex is orientated sub-parallel to flow of the TIS. Elongate ridges of glaciofluvial sands and gravels aligned sub-parallel to ice flow are classified as eskers (Warren and Ashley, 1994). The orientation of these parallel to ice flow indicates that the configuration of the internal drainage

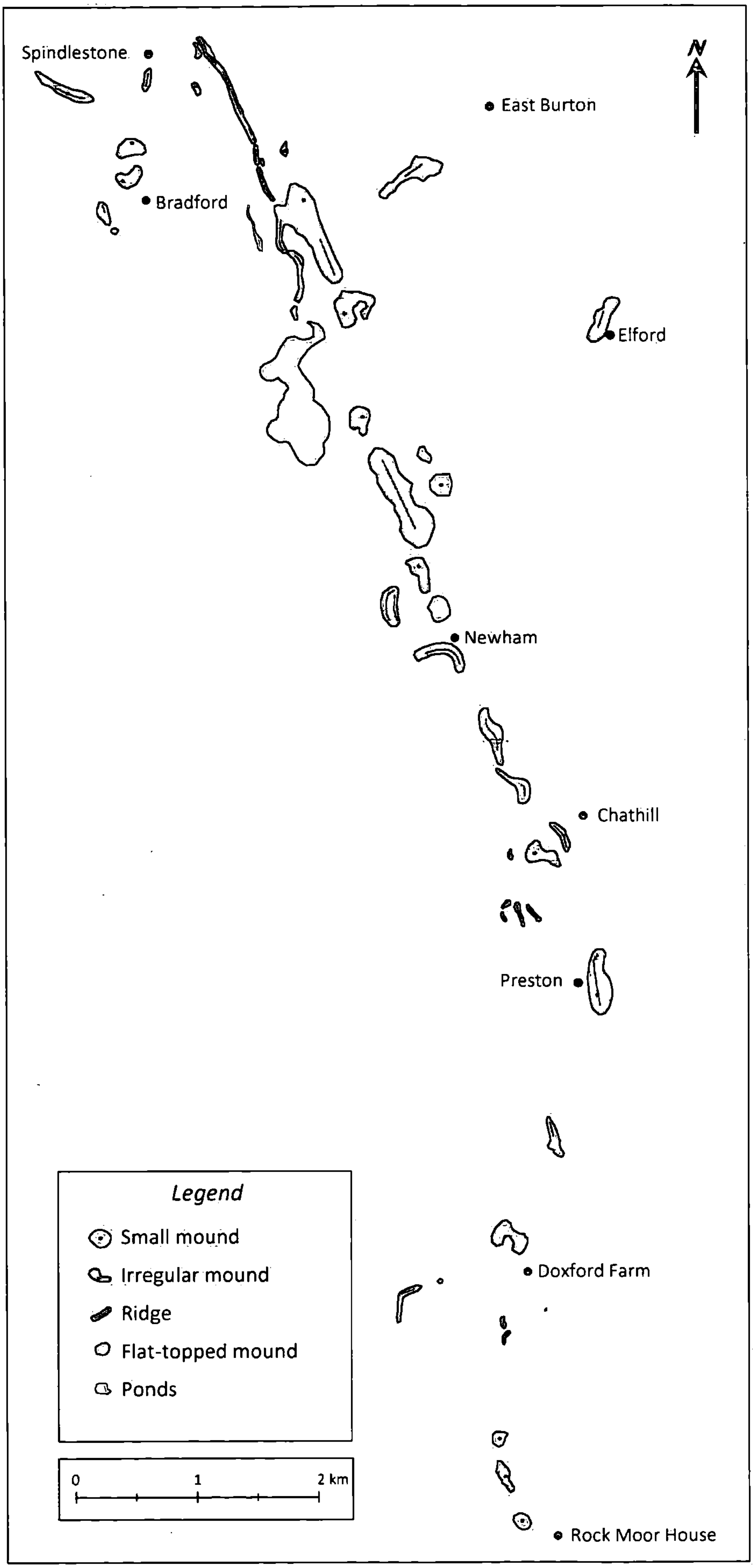


Figure 43: *The Bradford Hummocky Terrain Complex. Red star shows location of figure 42, page 59. Yellow triangle shows location of figure 44, page 61 and figure 45, page 61.*

system was driven by the ice surface slope. The interpretation of these features as eskers is in agreement with previous interpretations of this complex (c.f. Parsons, 1966 cited in Huddart, 2002b; Huddart, 2002b). It is proposed here that the northernmost esker was formed in a subglacial conduit, which is evident from its narrow, well-defined, smooth-sided appearance (Fig. 45).

The morphology of the subglacial conduit-fill eskers and abundance of glaciofluvial deposits reveal the nature of the drainage network that operated here. Gaps between the northernmost esker ridge are evidence of either discontinuous deposition within the subglacial tunnel, a constriction in the subglacial tunnel or post-depositional erosion by streams (Brennand, 1994). An alternative explanation is offered by comparison with 'beaded' eskers in central Ireland (Warren and Ashley, 1994) and Quebec (Banerjee and McDonald, 1975). These discontinuous, segmented eskers are interpreted to have been deposited in the distal regions of a conduit that was discharging into a standing body of water at the glacial terminus (Warren and Ashley, 1994).

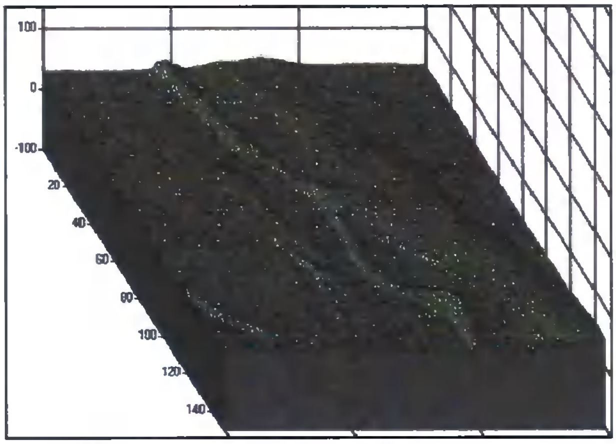


Figure 44: *Esker ridges of the Bradford Complex, just to the southwest of Bradford. Location shown in figure 43, page 60.*


Figure 45: *Smooth-sided esker at the northern end of the Bradford Complex. View is towards the north. Location shown by yellow triangle in Fig. 43, page 60. Photo: KEHS*

As the conduit drained into the lake, the reduction in velocity triggered sedimentation (Ashley and Warren, 1997, Delaney, 2001). It is therefore proposed that the northernmost esker in the Bradford Complex was formed in a subglacial conduit that fed into a proglacial lake. The gaps between the ridges record the steady retreat of the TIS and the North Sea Lobe (c.f. Warren and Ashley, 1994). Further evidence to support the presence of a proglacial lake at the terminus of these two ice masses is the fan-shaped mound at Hoppen Hill. From its close spatial association with the subglacial esker to the north, this feature is interpreted as a subaqueous fan delta, formed as the esker discharged into a lake. This is supported by morphological similarities with fan-shaped, flat-topped mounds of the Carstairs Esker system in Scotland (Thomas and Montague, 1997) and the eskers of central Ireland (Warren and Ashley, 1994; Delaney, 2001), which are interpreted as having formed in proglacial water bodies as water discharged from subglacial tunnels. This interpretation is supported by the sedimentology of Hoppen Hill; interbedded laminated clays and silts (as identified by Carruthers *et al.*, 1927) are evidence of deposition in a relatively quiescent water body.

Highly linear, elongate ridges at oblique angles to the main N-S trend of the Bradford Complex are identified to the east and west of the northernmost esker, south of Newham and west of Doxford Farm towards the south of the complex (see Fig. 43, page 60). The genesis of these features is difficult to infer without sedimentological surveys, however, on the basis of their morphology, it is considered likely that they are eskers. As mentioned above, ice flow is broadly parallel to the Bradford Complex, although their does appear to be a degree of flow convergence (explored further in section 5.2.1). As eskers form subparallel to ice-flow, it is not implausible that these are eskers. An alternative theory is that these oblique-trending ridges are supraglacial crevasse fills, deposited during ice stagnation. However, given the low preservation potential of supraglacial crevasse fills (c.f. Evans *et al.*, 1999), a subglacial esker origin is preferred.

As at Wooler and Cornhill, the presence of water-filled depressions and irregular sand and gravel hummocks in close association with the eskers are indicative of stagnating ice, which reinforces the notion that these glaciofluvial complexes formed during deglaciation. The above interpretation of the landforms of the Bradford Complex is in agreement with that of Huddart (2002b), who proposed the complex was formed as a series of subglacial eskers, glaciolacustrine deltas and supraglacial kames. The implication of the above interpretation for the local and regional glacial dynamics is discussed in the next chapter (see section 5.2)

4.3 Assemblage C: Meltwater Channels

Erosional channels have been mapped between 50 and 450 m OD in the Cheviots (Fig. 46, page 64) and 50 and 250 m OD in the Lammermuir Hills (Fig. 47, page 65). Channels range from short, shallow depressions to long, deeply incised, meandering channel systems with multiple inlets and outlets. Most channels are dry, although some are occupied by small misfit streams.

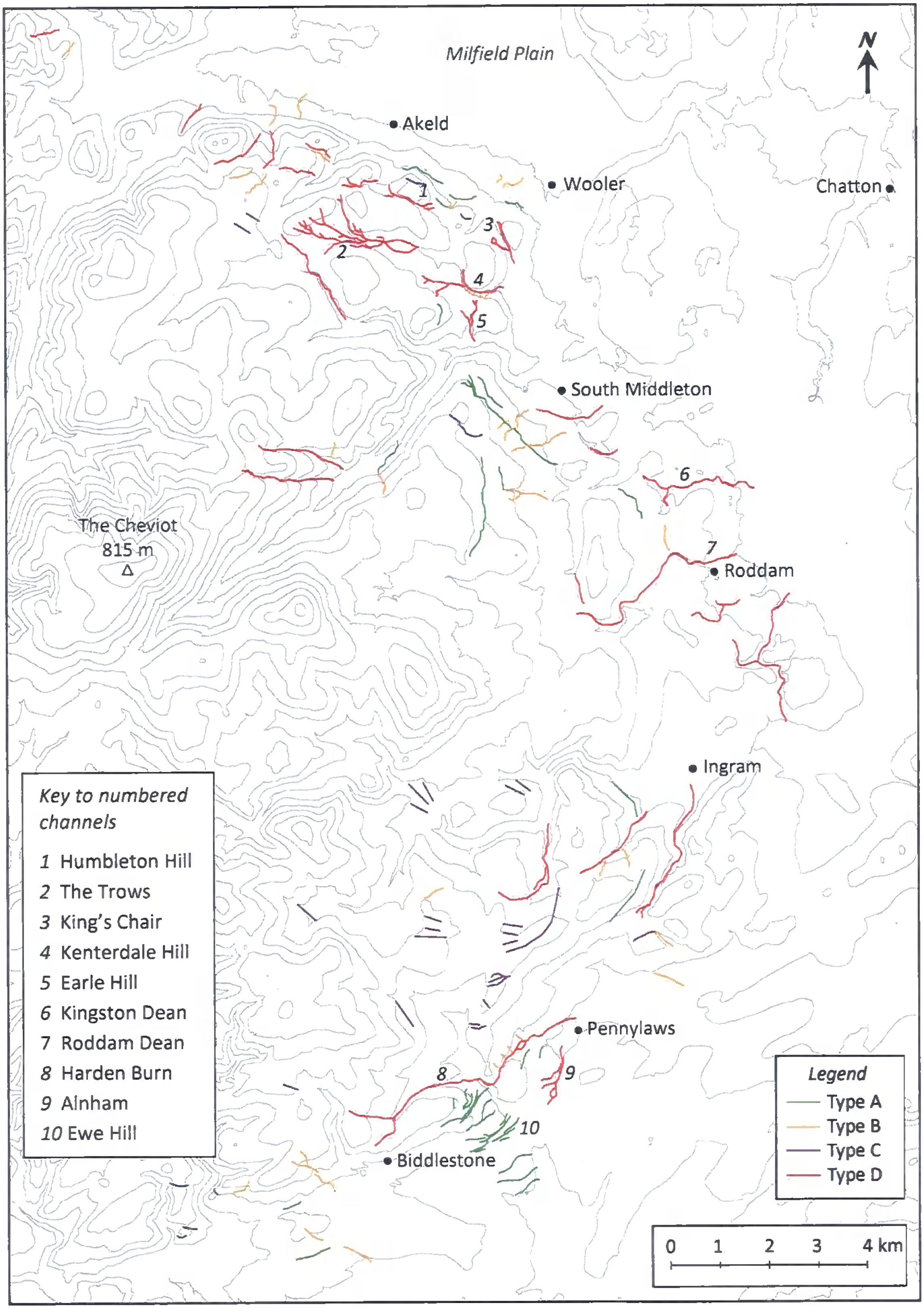


Figure 46: *Distribution of meltwater channels Types A-D in the west Cheviots. Contours are at 50 m intervals, constructed in ArcGIS from NEXTmap DEM.*

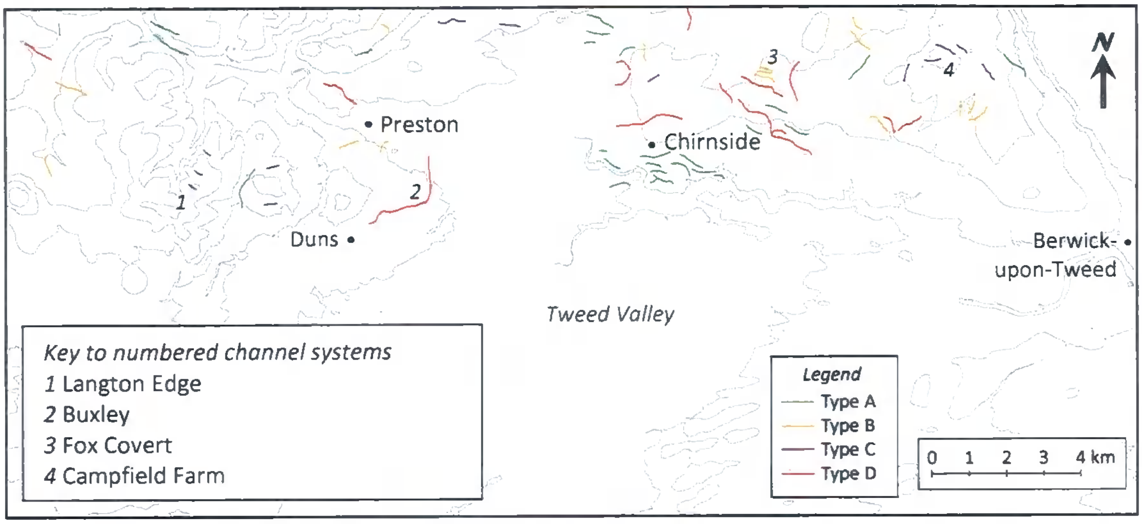


Figure 47: *Distribution of meltwater channels Types A-D in the southern Lammermuir Hills.*

Channels are predominantly cut into bedrock (andesite, basalt, rhyolite in the Cheviots, Llandovery, Lower Old Red Sandstone and Tuff and Aglomerate in the Lammermuir Hills (BGS Geology maps, NERC 2008)). In the Cheviots, channels are identified predominantly in the east. Between Roddam and Akeld, channel outlets are generally orientated towards the southeast. Between Roddam and Ingram, few channels are identified. South of Ingram, channel outlets are predominantly orientated towards the northeast, with those west of Biddlestone orientated east or southeast. In the Lammermuir Hills, the channels are predominantly orientated NW-SE, and appear to be aligned with the topography; the majority of channels are found in present day valleys.

The anastomosing channels are interpreted to have been mechanically eroded by meltwater flowing in subglacial, ice-marginal and submarginal systems. This is inferred from the form of the channels, their alignment with the topography and relationship with each another. The scale and number of these channels imply large volumes of water – and therefore ice – were present in the Cheviots at the time of formation (Sugden *et al.*, 1991). The association of these channels with glacial meltwater is in line with previous researchers (e.g. Carruthers *et al.*, 1930; Common, 1957; Price, 1960; Derbyshire, 1961; Clapperton, 1968; Lunn, 1995). Based on their morphology, four groups of channels are identified. All four are found in close association with each another in the Cheviots, although in the Lammermuir hills, only types A and D are widely identified.

4.3.1 Type A Channels: Ice-marginal channels

4.3.1.1 Description

These channels are aligned parallel to the contours and frequently start and end abruptly. In cross-section, they have the appearance of a bench rather than a complete channel. Often these channels are found orientated parallel to each another at successive heights down the hillsides. At Ewe Hill, such channels are orientated oblique to the contours, gently running down hillsides (system 10, Fig. 46, page 63). Well-defined examples of Type A channels in the Cheviots are found between Pennylaws and Biddlestone at Ewe Hill (system 7, Fig. 46, page 63) and west of South Middleton. Channels range in depth from around 5 to 25 m and vary in length from around 350 m to over 2.7 km. In several cases, such as at Akeld, these channels exhibit gently undulating long profiles. Channels are cut through the bedrock and also through Devensian Till. Occasionally multiple-inlets are identified, such as at Middleton Hall. In the Lammermuir Hills, type A channels are numerous south and east of Chirnside.

4.3.1.2 Interpretation

The alignment of these channels parallel to the contours indicates these were formed ice-marginally. As ice downwastes, marginal drainage channels form between the newly revealed hill-slopes and the stagnating ice mass (Mannerfelt, 1949; Dyke, 1993). These channels therefore most likely formed as the TIS retreated from the flanks of the Cheviots. The ice formed one half of the meltwater channel, explaining the bench-shape of these channels (Glasser and Bennett, 2004) (Fig. 48). Channels eroded at increasingly lower elevations, such as around Pennylaws and Ewe Hill represent the position of the ice margin during retreat (Glasser and Bennett, 2004). The low gradient of these channels indicate that the ice surface was relatively shallow. Since these channels would have been open to the atmosphere, water flowed with gravity downhill. However, due to the propensity of meltwater to drain into subglacial chutes (channel Type B), such channels may not fully represent the ice margin position (Mannerfelt, 1949). It is proposed that those channels with gently undulating long-profiles, flowed beneath the ice at the margins. These channels are referred to as submarginal channels and represent erosion by subglacial waters flowing more or less parallel to the hillside. The abrupt inlets and outlets of ice-marginal channels are interpreted to be the result of ice-marginal and submarginal waters plunging directly beneath the ice, possibly through subglacial or englacial channels (Benn and Evans, 1998).

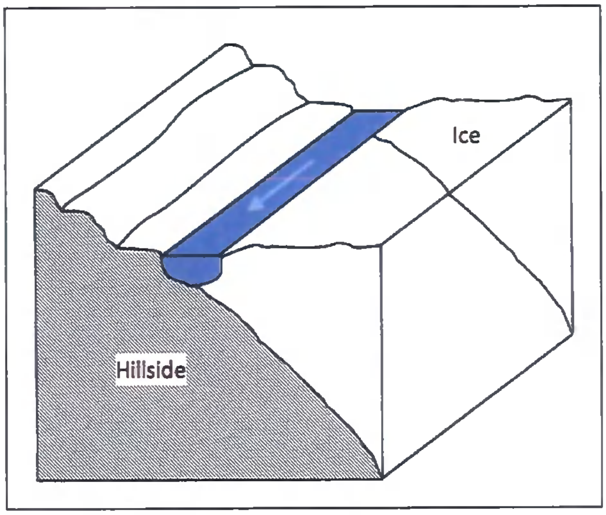


Figure 48: *Model of ice-marginal channel formation. Ice margin is retreating, cutting channels at progressively lower altitudes.*

4.3.2 Type B Channels: proglacial channels and submarginal chutes

4.3.2.1 Description

Channels in this group run straight down the hillsides, approximately at right angles to the contours and are cut through bedrock, till and glaciofluvial deposits that mantle the lowest slopes of the Cheviot massif. They are often short (< 300 m long) and start and end abruptly. In a few cases, these channels have short sections that are undulating in long profile. Several channels terminate in larger channel systems, changing direction and running obliquely downhill. In cross-section, a few channels have more deeply incised upper sections than lower. In the Cheviots, such channels are observed at Middleton Hall, at the Earl Hill system and at Pennylaws. Type B channels are infrequently identified in the Lammermuir Hills, although a few are found between Berwick and Chirnside.

4.3.2.2 Interpretation

These channels are interpreted as proglacial channels and/or submarginal chutes (e.g. Mannerfelt, 1949). The morphologic criteria used to identify these channels are primarily their position on the hillside and also their relationship with other channel systems. These channels may be proglacial channels, formed by meltwater flowing downhill from the Cheviot Ice Cap (Clapperton, 1970b) over the ice-free land between the Tweed and Cheviot ice (Fig. 49). As these ice masses separated during deglaciation, meltwater would have been focussed at the margins of the TIS and in the proglacial zone of the Cheviots. Alternatively, these channels may be submarginal chutes, formed where submarginal and subaerial waters plunged beneath the TIS margin (Fig. 49). It is considered likely that both these situations could have occurred.

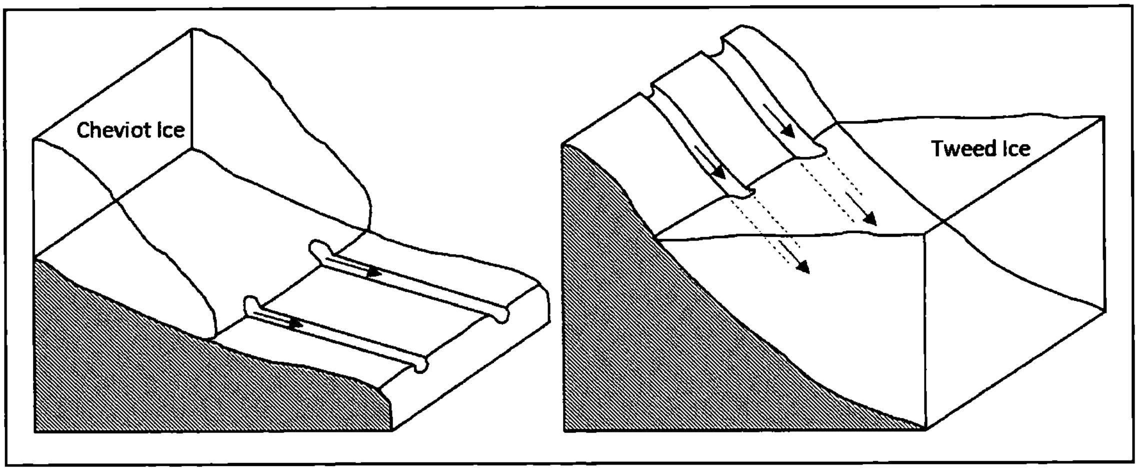


Figure 49: *Model of Type B Channel formation. (a) Channels formed in a proglacial setting by runoff from the Cheviot Ice Cap, (b) Channels Formed as submarginal and subaerial water (and possibly Cheviot ice cap runoff) plunge directly beneath the TIS margin.*

Submarginal chutes frequently exhibit steep gradients, changes in directions as they join marginal channels and abrupt inlets and 'hanging' outlets (Menzies and Shilts, 1996; Clark *et* *al.*, 2006). Several of the channels mapped in the Cheviots do exhibit seed points than appear to start out of nowhere on the hillside. Such channels are visible southwest of South Middleton (see Fig. 46, page 63). These channels have been previously interpreted as subglacial chutes (e.g. Clapperton, 1968; Lunn, 1995), a theory proposed to explain morphologically similar channels elsewhere. The lengths of submarginal chutes in the Cheviots and Lammermuir Hills allow inferences to be made on ice thickness. The distance that these channels were able to penetrate through Tweed and Cheviot ice is likely to be determined by the englacial piezometric water level (water table). Glenn (1954, cited in Clapperton 1970) proposed that such channels should only be able to penetrate through ice that was 600 ft (182 m) thick. Given that several of these channels run down hillsides for distances up to 300 m, it is suggested that meltwaters penetrated through greater thicknesses of ice. As the ice retreated from the Cheviots, it is envisaged that ice-marginal meltwaters plunged rapidly beneath the stagnating ice (c.f. Glasser *et al.*, 1999). However, Glasser *et al.* (2004) argue that a subglacial origin for such channels is tenuous as there are limited recorded modern processes of subglacial chute formation. Instead, they invoke a subaerial fluvial origin, with channels incised during eustatic lowstands. Despite this, a submarginal chute origin is favoured here. Subaerial modification of these channels is considered to have been limited, as all observed channels were dry.

4.3.3 Type C Channels: Superimposed englacial channels

4.3.3.1 Description

Channels in this group are characterized by their short (~250-600 m long), shallow and poorly defined form. They appear to have no apparent orientation to the underlying topography, instead cutting across contour lines, spurs and hill summits. They are frequently short and poorly defined. The vast majority of such channels are carved into bedrock. Such channels are found south of Akeld and also on the higher sloped between Ingram and Pennylaws, where they are found parallel to one another (see Fig. 46, page 63. Northwest of Duns, these channels cut directly across the Langton Edge spur separating Wellrig Burn from Mill Burn (see Fig. 47, page 64).

4.3.3.2 Interpretation

These channels cutting across spurs, with no apparent alignment with the relief, are interpreted as superimposed englacial channels. Clapperton (1968) proposed that where ice downwastes, supra- and englacial channels are brought to successively lower elevations until they become subglacial and erode through the highest points of the landscape (Fig. 50, page 68). These channels have been referred to as severed spur channels. It is unlikely that severed spur channels represent superimposition of supraglacial channels, as by nature, these channels are ephemeral and frequently drain through crevasses and moulins into en- and subglacial channels (Benn and Evans, 1998). The distribution of these channels throughout the Cheviots can be explained in terms of ice flow direction. Clapperton (1968) proposed that the orientation of spurs and valleys at right angles to ice flow (and therefore meltwater flow) resulted in the superimposition of channels directly across these topographic high points. The majority of these channels south of Roddam are orientated towards the southwest, which differs from subglacial (Type D) and ice-marginal (Type A) channels, the orientation of which is towards the southeast north of Roddam (TIS) and northeast south of Roddam (SIS). This implies that ice flow was from the northwest, i.e. from the central regions of the Cheviots, which adds weight to the argument that the Cheviots were occupied by ice, although it is difficult to say whether it was an independent ice cap or an overriding ice mass. Lunn (1980; 1995) highlights that, in glaciers at pressure melting point, meltwater will descend to the glacier bed and flow subglacially and that to explain superimposed channels, a thermally layered ice sheet is invoked.

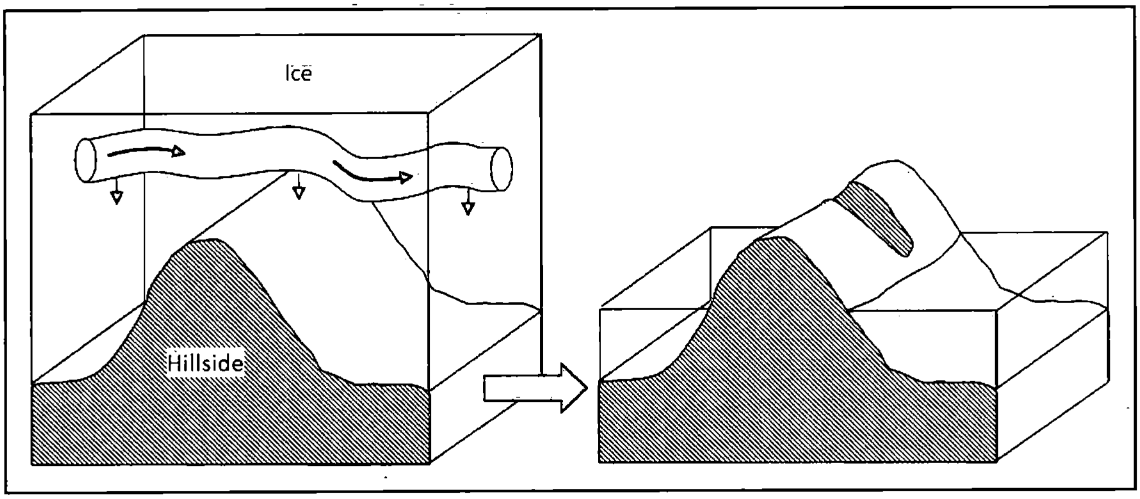


Figure 50: Model of englacial channel superimposition. Englacial conduit is lowered onto a spur or hilltop as the ice downwastes (left). The result is a 'severed spur channel' (right)

4.3.4 Type D Channels: Subglacial channels

4.3.4.1 Description

These channels mainly follow present day valleys. These channels are largest of the four types and most deeply incised (Fig. 51b, page 69), although some exhibit more shallow cross-sections (Fig. 51a, page 69). Cross sections are V- to U-shaped and predominantly symmetrical, although several exhibit sections that are asymmetric. They have anastomosing, undulating long profiles (Fig.52, page 69) and are up 5 km to in length. Channels are, in places, occupied by peat and till. Multiple inlets, frequently in cols, and multiple outlets are identified south of Pennylaws, in The Trows (Fig. 53, page 70) and the Earl Whin system south of Wooler. In several cases, channel tributaries are type B channels (see above). In the Lammermuir Hills, these channels are less well defined, with no multiple inlets nor outlets, nor bifurcating sections. These channels are however longer than those in the Cheviots, are highly sinuous and exhibit undulating long profiles.

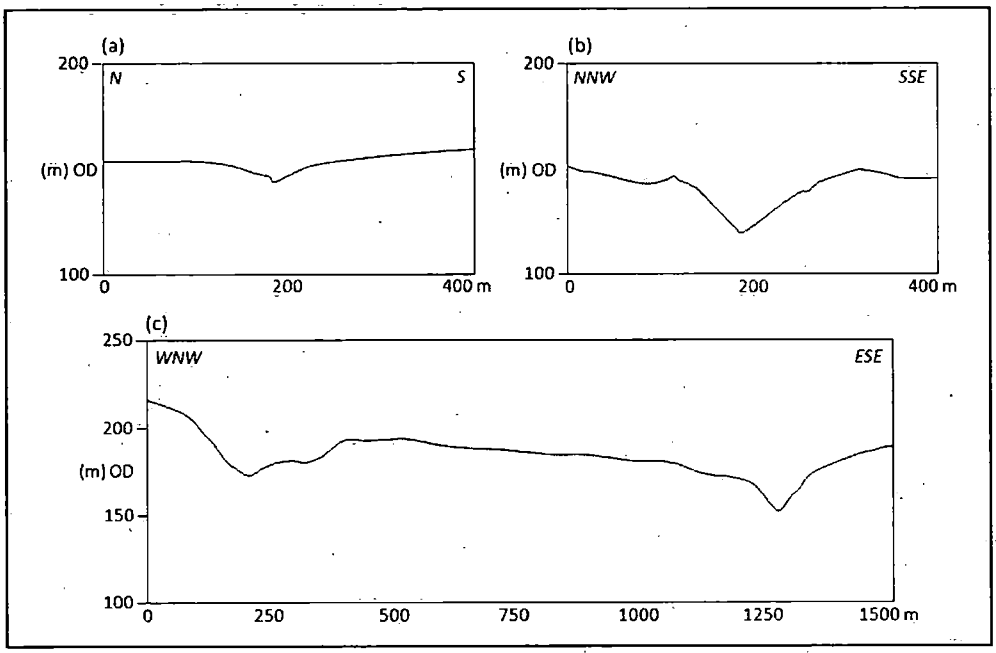


Figure 51: *Type D Channel cross sections in the Cheviots. (a) cross-section of Kingston Dean Channel (channel number 6, Fig.46, page 63); (b) cross-section of Roddam Dean Channel (channel no. 7, Fig.46, page 63); (c) cross section of the Harden Burn channel (channel no. 8, Fig.46, page 63) and the Alnham Channel (channel no. 9, Fig.46, page 63) to the south of Pennylaws.*

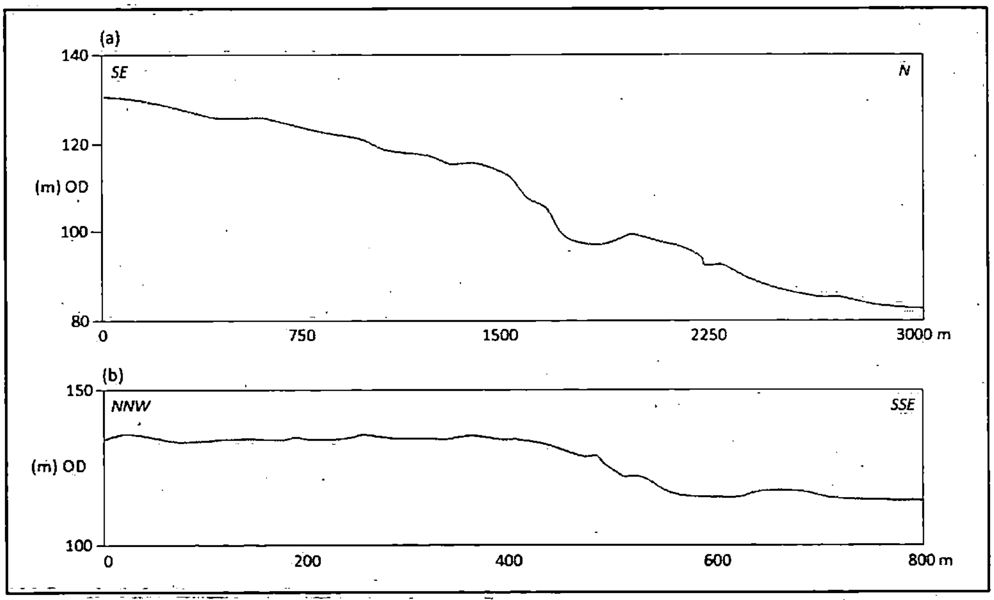


Figure 52: *Channel long profiles. (a) Buxley Channel in the Lammermuir Hills (channel no. 2 in Fig.47, page 64); (b) Kings Chair Channel in the Cheviots (channel no.3 in Fig. 46, page 63). Note undulating sections.*

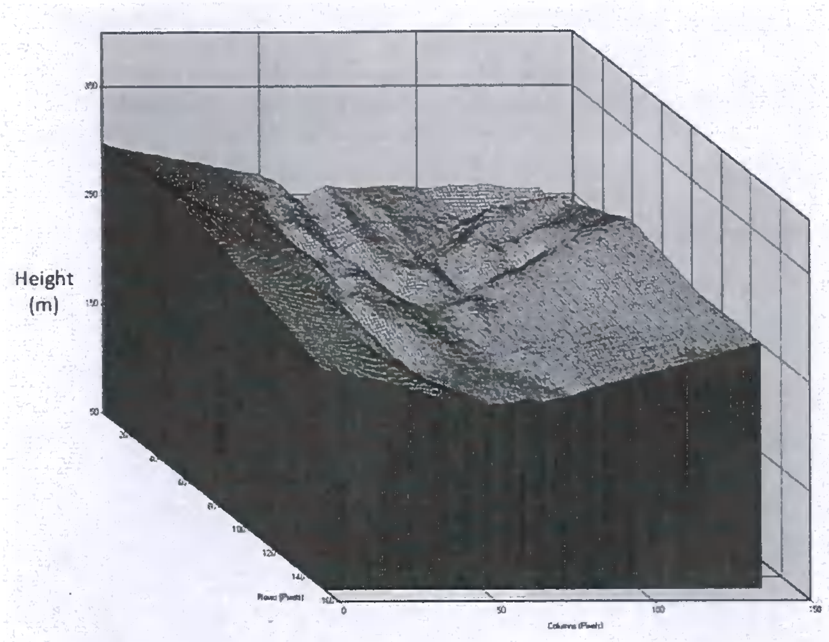


Figure 53: *DEM of The Trows system south of Wooler. View is towards the west. Note the anastomosing channels and multiple inlets. (DTM from NEXTmap® Britain, Intermap Technologies, 2008).*

4.3.4.2 Interpretation

These channels are interpreted as subglacial meltwater channels. This interpretation is based on their undulating, anastomosing long-profiles, abrupt inlets and outlets and relationship with other channels. The anastomosing and bifurcating nature of the channels in the Cheviots south of Wooler and south of Pennylaws can provide an insight into the organisation of the subglacial drainage system. Clapperton (1968) proposed that during deglaciation, as ice was thinning, meltwater drainage was concentrated in the cols, which lends an explanation to those channels with inlets in cols. Meltwater that penetrates through the glacier at the margins or through moulins, crevasses, cavities and fractures can flow in one of two ways on reaching the bed. Water either flows slowly and inefficiently in distributed, linked cavity systems or relatively turbulently and efficiently through conduits (Boulton *et al.*, 2007). Anastomosing erosional channels are thought to represent the former, with meltwater flowing in linked cavity systems. Other sections of these channels are also dendritic, which represents discrete canalised flow (Benn and Evans, 1998). This implies a shift between conduit and cavity flow during channel formation, which may happen where the glacier flows from a hard bed onto unlithified sediment (Boulton *et al.*, 2007). The Trows system is anabranched where its lower section cuts through Devensian till. It may be that the flow of ice from bedrock to unconsolidated sediment triggered a switch from channelized flow to a distributed flow system. Alternatively, anabranched sections may form over time where englacial channels are superimposed over subglacial channels (Menzies and Shilts, 1996) or as subglacial channels switch their course (Benn and Evans, 1998). Given that the majority of channels are cut into the bedrock in the Cheviots and Lammermuir Hills only very few exhibit changes in substrate along their long profiles, it can be suggested that anabranched sections form in this way.

Arguably the most diagnostic feature of subglacial channels is the undulating long profile. This characteristic is observed in a vast majority of channels identified in the Lammermuir and Cheviot Hills, such as at Humbleton Hill and the Kings Chair Channel (Fig. 54). Subglacial channels are often referred to as 'ice-directed channels' (Benn and Evans, 1998), as their orientation is controlled by the hydraulic potential gradient, which is determined by the glacier slope and the subglacial topography (Shreve, 1972). Water flowing in subglacial conduits (i.e. a closed system) is under hydrostatic pressure and can consequently flow uphill (Sugden *et al.*, 1991; Menzies and Shilts, 1996), which accounts for the occurrence of undulating long profiles. Since these channels predominantly follow present day valleys, it can be suggested that their formation was influenced by the pre-existing topography. Subglacial channels are normally associated with warm-based ice, where meltwater can penetrate to the bed (c.f. Sugden *et al*, 1991). However, in the Cheviots, the lack of glacial erosion indicates a cold-based ice cap was present (Clapperton, 1970), which is potentially incompatible with the meltwater evidence presented here. However, given this lack of evidence for glacial erosion and the fact that the channels in the Cheviots are only likely to have formed where meltwater was abundant (c.f. Clapperton, 1968), it is suggested that the subglacial channels formed during deglaciation.

Figure 54: *Steeply incised, v-shaped Kings Chair Channel in the Cheviots. (Channel no.3 in Fig. 46, page 53). View is down-channel towards the south. Note rising long-profile at centre of picture. Photo: KEHS.*

5

DISCUSSION

Geomorphological mapping of the Tweed Valley and surrounding area has identified three distinct glacial landform assemblages; subglacial streamlined bedforms, meltwater channels and hummocky glaciofluvial deposits. The identification of similar landforms in other formerly glaciated regions has facilitated the reconstruction of the glacial processes, subglacial thermal conditions, configuration and dynamics of palaeo-ice sheets (e.g. Warren and Ashley, 1994; Punkari, 1997; Kleman and Glasser, 2007). In this chapter, the implications of the data generated in this study for the palaeo-ice sheet dynamics of the Tweed Valley and surrounding area will be discussed.

5.1 The case for a Tweed Ice Stream

Tracks of numerous, closely-spaced highly attenuated bedforms have been attributed with ice streaming in a number of formerly glaciated and contemporary settings (e.g. Laymon, 1992; Hodgson, 1994; Everest *et al.*, 2005; Golledge and Stoker, 2006; Stokes *et al.*, 2006). Previous researchers have invoked fast ice flow to explain the highly attenuated bedforms in the Tweed Valley (e.g. Clapperton, 1971a; Clark *et al.*, 2004). This led Everest *et al.* (2005) to propose that these bedforms are the clear signature of a palaeo-ice stream. In this section, the evidence to support the presence of the Tweed Ice Stream will be discussed. Stokes and Clark (1999) constructed a set of criteria for the identification of palaeo-ice streams from the geomorphological record (Table 2, page 37). Based on these criteria, the existence of the TIS is strongly supported by the following five criteria:

(1) The shape and dimensions of the TIS pathway are characteristic of both palaeo-ice stream tracks and contemporary active ice streams. The TIS track is, at its widest, 35 km across. Contemporary ice streams are frequently wider than 20 km (Stokes and Clark, 1999). The Amundsen Gulf palaeo-ice stream track is very large, at over 150 km wide and 1000 km long (c.f. Stokes *et al.*, 2006). The onshore section of the TIS is approximately 65 km long in the Tweed Valley and 40 km long along the NNCP, however its total length is unknown as the terminus is inferred to have been offshore (Everest *et al.*, 2005). The shape of the TIS track is

also characteristic of ice streams; ice streams of the Antarctic and Greenland ice sheets characteristically have funnel-shaped upstream onset zone that lead into a main trunk (Jezek, 1999).

(2) The arrangement of streamlined subglacial bedforms reveals converging ice flow in the upper Tweed Valley, leading into a narrower trunk between the Cheviot and Lammermuir Hills. It is inferred that velocities increased through the zone of convergent flow into the central trunk of the TIS. This is based on a comparison with the WAIS ice streams, from which geophysical data reveal zones of flow convergence are characterised by increasing flow velocities (c.f. Jacobel *et al.*, 1996; Rignot, 2006; Joughin *et al.*, 2006). It is inferred that this convergent bedform pattern in the upper Tweed Valley represents the onset zone of the TIS. Modern ice stream onset zones are identified by convergent surface flow patterns and high velocity gradients from areas of featureless ice to zones of heavily crevassed ice (Bell *et al.*, 1998).

(3) The highly attenuated, closely-spaced streamlined bedforms with elongation ratios <1:12 are indicative of rapid ice flow (Stokes and Clark, 2002; Briner, 2007). It was thought initially that constancy of flow controls bedform elongation (Boyce and Eyles, 1991). However, the link between fast velocities and bedform elongation is now recognised; geomorphological evidence (drumlins and MSGL) from the Antarctic Continental Shelf has been associated with ice streams that retreated during the Holocene (c.f. Shipp *et al.*, 1999; O'Cofaigh *et al.*, 2008). Flow velocities of these ice streams are observed to exceed 0.8 km yr^{-1} (Bennett, 2003).

(4) The spatial variation in streamlined bedform size and elongation ratios is a product of the distinct velocity field of ice streams (Stokes and Clark, 2002). The most elongate bedforms are identified in the centre of drumlin fields; in the New York drumlin field, elongation ratios increase along a central flowline (Briner, 2007). In contemporary ice streams, flow velocities along the central regions of ice streams have been observed to exceed 800 m yr^{-1} (Paterson, 1994). At the lateral margins, drag is exerted by the slower-flowing ice (< 10 m yr^{-1}) (Anandakrishnan *et al.*, 1998; Bennett, 2003). These spatial velocity variations are reflected at the ice surface as crevasses and longitudinal ridges and troughs (Whillans and Merry, 2001). It has been proposed that the presence of a subglacial deforming layer permits rapid flow under the low driving stresses characteristic of ice streams (Blankenship et al., 2001; Tulaczyk et al., 2001). It is therefore inferred that flow velocities were greatest towards the centre of TIS track, as the most elongate bedforms (>1:10) have been identified here. Flow was slowest towards the ice margins, where shorter, less elongate bedforms have been mapped.

(5) The lateral margins through the main trunk of the TIS track are abrupt and are delineated by the higher ground of the Lammermuir Hills to the north and the Cheviots to the south. This sharp boundary between streamlined terrain and non-streamlined terrain represents the transition zone between two zones of ice with differing thermal and dynamic regimes (Raymond *et al.*, 2001; Hall and Glasser, 2003; Golledge and Stoker, 2006). Contemporary ice stream lateral margins are visible at the ice surface as zones of intensely crevassed ice up to 2 km wide (Bell *et al.*, 1998; Stokes and Clark, 1999). In contemporary ice streams, ice stream lateral margins are visible as a sharp boundary between crevassed ice exhibiting flow-parallel banding and 'featureless' ice (Bell *et al.*, 1998). Across this boundary, high velocity gradients are observed, with velocities often several degrees of magnitude higher within the ice stream than outside it. This explains the presence of streamlined bedforms inside but not outside this boundary.

5.1.1 Tweed Ice Stream Dynamics

Based on above geomorphological criteria, the existence of the TIS is strongly supported. Previous researchers have proposed that fast ice flow was responsible for the formation of the northeast trending drumlins in the Tweed Valley (e.g. Clapperton, 1971a; Clark *et al.*, 2004; Everest *et al.*, 2005). However, new data presented in this study suggest that the TIS was more dynamic than previously thought. Cross-cutting bedforms southwest of Berwick-upon-Tweed (see Fig. 18, page 34) have led to the identification of two superimposed flow-sets (see Fig.25, page 40). Both flow-sets fulfil 5 of the 8 geomorphological criteria for ice stream activity (see table 2, page 41). The first flow-set, situated in a broad arc around the Cheviot massif (shown in red in Fig. 25, page 40), shows a stronger signature in the landscape than the second flow-set, which trends towards the northeast in the lower Tweed Valley (see section 4.1, shown in blue in Fig. 25, page 40). Highly complex, cross-cutting flow-sets associated with ice streaming in the Canadian Arctic Archipelago and mainland Canada represent a reorganisation of the ice margin or a shift in the ice divide during retreat growth and retreat phases (Stokes *et al.*, 2006). The presence of two flow-sets in the Tweed Valley and surrounding area are interpreted to represent a shift in the axis of flow of the TIS. Owing to the subtle nature of the NE trending flow-set, it is proposed that localised cross-cutting or flow-switching occurred, possibly during deglaciation. The southwards diversion of the Tweed, Solway and Cheviot ice is understood to have occurred as ice advanced into the North Sea Basin (NSB) from the Scottish Highlands during the Devensian (Sissons, 1964; Lunn, 1980; Huddart and Glasser, 2002). Occupation of the NSB was episodic, with advance and retreat phases occurring throughout the Devensian (Sejrup *et al.*, 1994; Carr *et al.*, 2006; Nygard *et al.*, 2007). The NE orientated bedforms southwest of Berwick-upon-Tweed suggest ice flow was directly offshore and therefore must have formed when this sector of the NSB was ice free (shown in blue in Fig. 55). However, the timing and duration of NSB ice advances are debated, which makes any inferences on the timing of the TIS flow-switching event speculative.

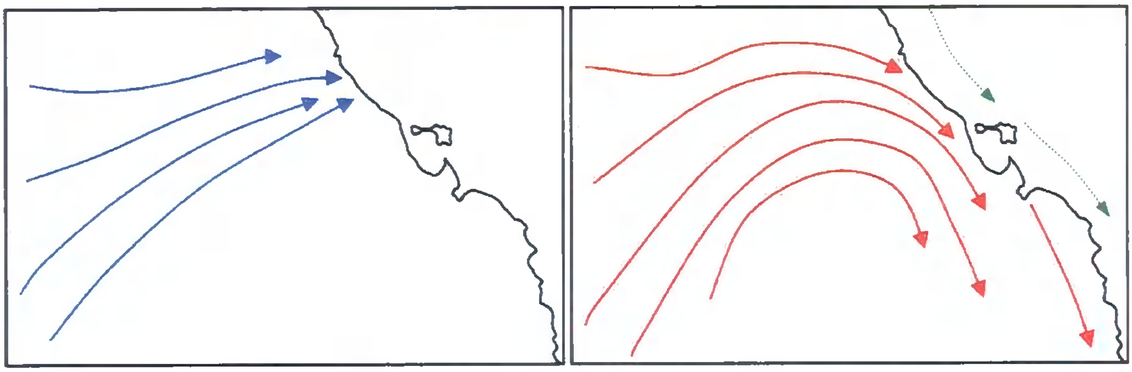


Figure 55: *Flow trajectories of the TIS inferred from orientation of streamlined bedforms.* ***LEFT:*** *flow direction NE offshore during periods where North Sea was ice free.* ***RIGHT:*** *Diversion of the TIS with growth of the North Sea lobe (flow direction in green dashed arrows). Based on data collected in this study and Carr et al., 2006; Nygard et al, 2007.*

Lambeck (1993) proposed that the NSB was largely ice free from ~26 ka BP, whereas more recent reconstructions show ice advanced later in the Devensian (c.f. Carr *et al.*, 2006; Nygard *et al.*, 2007). Carr *et al.* (2006) proposed the BIS advanced into the NSB between 18-16 ka BP, during which time an ice free corridor existed between the BIS and FIS. Nygard *et al.* (2007) proposed that the BIS and FIS were confluent until ~24 ka BP and that subsequent to this, two oscillations of the BIS occurred between ~17-15.5 ka BP. During these more recent oscillations, the BIS was not coalescent with the FIS and flowed southwards along the east coast of Britain to the southern NSB (Nygard *et al.*, 2007). It is therefore considered likely that flow of the TIS switched from onshore to offshore several times as ice advanced and retreated in the NSB. The NE trending flow-set southwest of Berwick-upon-Tweed shows flow was directly offshore, and thus implies that this sector of the NSB was ice free at this time. Thus, the NE trending flow-set may have either formed prior to or immediately following the development of the North Sea Lobe. It is proposed here that these bedforms are the product of the most recent flow stage of the TIS, perhaps during deglaciation. If these bedforms were formed prior to ice developing in the NSB, it is expected they would have been reworked or obliterated by the later flow event. As it is, the subtle nature of these features are interpreted to reflect localised flow switching during deglaciation.

5.1.2 Controls on ice streaming

The onset of ice streaming in both contemporary and palaeo-environments is frequently attributed to the nature of the subglacial topography and the presence of a subglacial deforming bed (e.g. Anandakrishnan *et al.*, 1998; Bell *et al*, 1998; Joughin *et al.*, 2002; Peters *et* al., 2006). More recently, De Angelis and Kleman (2008) proposed that the location of palaeo-ice stream onset zones were by a basal thermal boundary and glaciological factors. The onset zone of the TIS in the upper Tweed Valley is just to the west of this study site and has not been mapped. However, from studying the relationship of the mapped streamlined bedforms with the surrounding topography (Fig. 56), it can be seen that flow converges in the upper Tweed Valley. Surrounding the upper Tweed Valley is a ring of higher ground, extending from the Lammermuir Hills round to the Cheviots, with tributary valleys flowing into the Tweed Valley (black dashed arrows in Fig. 56). It is suggested these tributary valleys may have acted to channel ice into the Tweed Valley. In West Antarctica, it has been observed that the position of subglacial valleys coincide well with ice stream tributaries in the ice stream onset zones (Joughin *et al.*, 2002). In the Cairngorm Mountains in Scotland, the location of preglacial valleys is suggested to have played a dominant role in the development of ice streams by acting to channel flow, promoting strain heating and facilitating flow acceleration (Hall and Glasser, 2003).

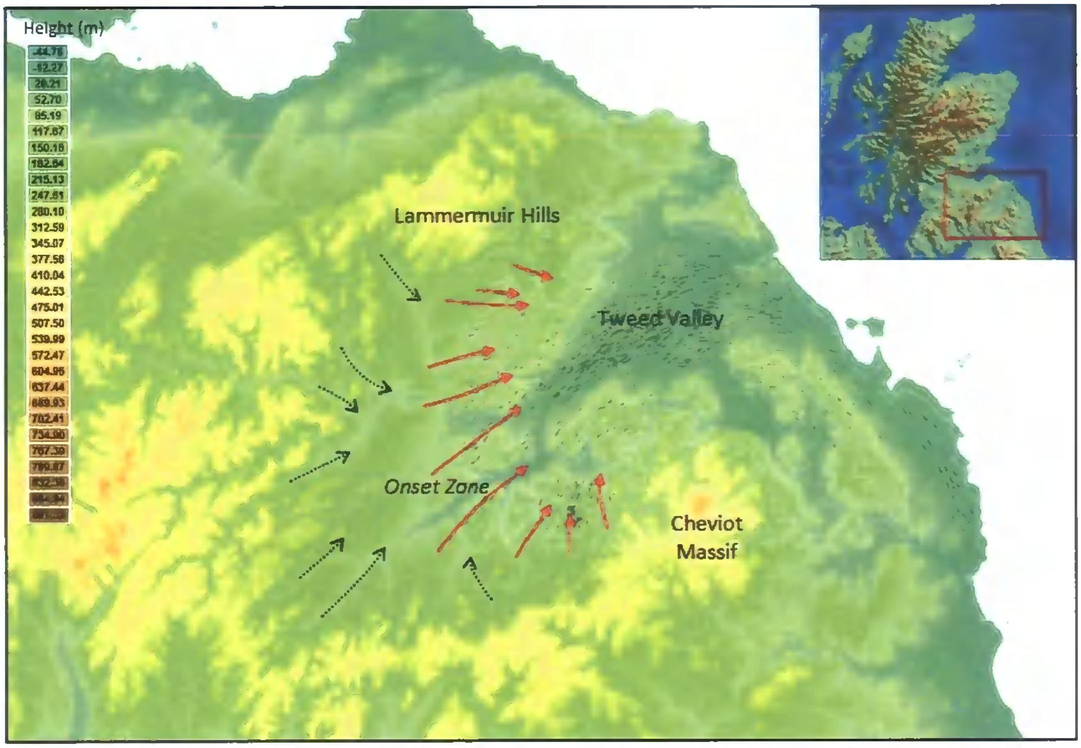


Figure 56: *Ice flow directions (shown in red) in the upper Tweed Valley as inferred from mapped bedforms (shown in black). Note that ice flow converges in the upper Tweed Valley, which is surrounded by higher ground. Black dashed arrows show orientation of tributary valleys. Inset map shows location in relation to the northern British Isles. Basemap SRTM relief-shaded DEM (CIAT, 2004).*

In terms of the flow of the TIS along the Tweed Valley, it is likely that the subglacial topography acted to channel the ice. The sharp lateral margins of the TIS are delineated by the higher non-streamlined ground of the Lammermuir and Cheviot Hills to the north and south of the main trunk of the ice stream (see Fig. 56 and Fig. 57, page 77). A similar situation is observed at Recovery Ice Stream of the Filchner-Ronne Ice Shelf in Antarctica, where mountains to the north and south exert a topographic control (Joughin *et al.*, 2006). In the valleys of the Cairngorms, there is evidence of intense glacial erosion, whereas the higher ground between have been relatively unmodified (Hall and Glasser, 2003). Similar zones of selective glacial activity are also documented in Marie Byrd Land, Antarctica and reflect spatial contrasts in the subglacial thermal organisation of the ice sheet (Kleman and Glasser, 2007). The large scale physiography and glacial geomorphology of the Tweed Valley and surrounding area bears similarities to such regions of selective glacial activity. The Cheviots show little evidence of glacial activity and are proposed to have been occupied by a cold based ice cap (Clapperton, 1970; Everest *et al.*, 2005, discussed in section 5.4), whereas the Tweed Valley and NNCP show evidence of intense ice streaming (i.e. warm-based ice). Whilst the development of the preglacial topography is unknown, it is suggested that the lateral margins of the TIS were fixed by the higher ground of the Cheviots and Lammermuir Hills. If the TIS was confluent with Cheviot ice (as was proposed by Clapperton, 1970) and southern upland ice in the Lammermuir Hills, the jump in elevation occurring at the bed of the TIS would have resulted in the development of a shear margin between these ice masses (Raymond *et al.*, 2001). This is explored more fully in section 5.3. It is also considered likely that the extensive till sheet across the Tweed Valley and along the NNCP (see Fig. 23, page 37) facilitated rapid basal movement through subglacial deformation (Everest *et al.*, 2005). It is therefore proposed that the onset of ice streaming was initiated by the topographic jump between the higher terrain surrounding the upper Tweed Valley and the lower ground of the main trunk of the Tweed Valley. Flow was facilitated in the Tweed Valley by basal sliding over the deformable substrate, with the position of the ice stream being fixed by the higher ground to the north and south.

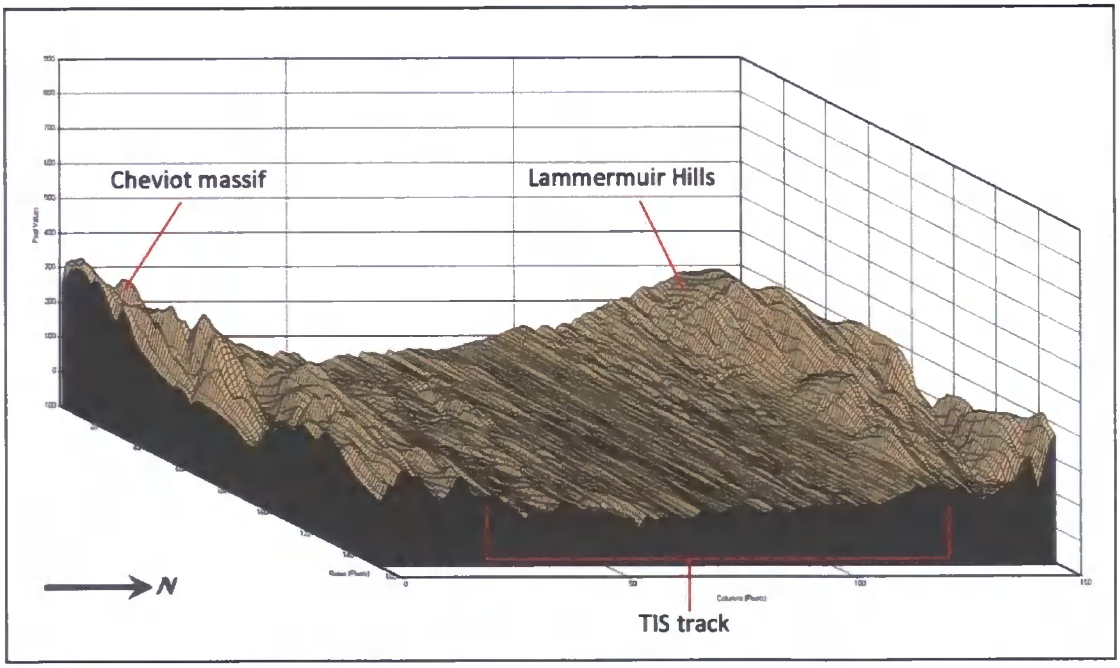


Figure 57: *DEM of the TIS track showing sharp transition between streamlined ground of the TIS track and higher non-streamlined ground of the Lammermuir and Cheviot Hills to the north and south.*

5.1.3 Duration of ice streaming

From the morphology and spatial characteristics of the subglacial streamlined bedforms of the TIS, and their relationships with other landforms, it is possible to infer whether they were formed time transgressively or isochronously (Clark, 1999). Establishing this is important for determining the history and dynamics of the ice stream (Stokes and Clark, 2001). The synchroneity of bedforms within flow-sets reveals information on when and over what timescale they formed (Clark, 1999). Time transgressive ice stream imprints are identified by overprinted and modified bedforms and reflect formation over successive glaciations during advance and retreat phases (Stokes and Clark, 2001). The bedforms of both flow-sets of the TIS are more characteristic of an isochronous ice stream imprint, in that bedforms show little sign of reworking or overprinting (c.f. Clark, 1999; Stokes and Clark, 2001). Similar, isochronous ice stream imprints are observed in the Amundsen Gulf in the Canadian Arctic Archipelago, where systematic highly parallel bedforms within individual flow-sets are interpreted as evidence of rapid flow during a single flow event (Stokes *et al.*, 2006). Subglacial bedforms along the TIS path are highly parallel, highly attenuated and are therefore interpreted to have formed relatively fast (Stokes and Clark, 2002). As with other isochronous palaeo-ice streams, it is suggested here that the observed bedforms represent the imprint of the ice stream immediately prior to ice stream shut-down (c.f. Stokes and Clark, 2003).

5.1.4 Retreat of the TIS

From the geomorphology of the TIS track, it is inferred that ice stream retreat was continuous and relatively rapid. This is based on a comparison with the geomorphological imprint of palaeo-ice streams in northern Canada (c.f. Stokes *et al.*, 2006) and the imprint of ice streams that retreated from the Antarctic Continental shelf post-LGM (c.f. Shipp *et al.*, Lowe and Anderson, 2002; Ó Cofaigh, 2008). Landforms superimposed transverse to flow over streamlined bedforms, such as terminal moraines and grounding zone wedges are evidence of stillstands of the margin during slow or episodic retreat (Lowe and Anderson, 2002; Ó Cofaigh *et al.*, 2008). Stokes *et al* (2006) speculated that a stillstand in the margin of the Amundsen Gulf Ice Stream in the Canadian Arctic Archipelago was recorded by a morainal bank. Rapid, continuous ice stream retreat, such as that recorded across Marguerite Bay of the Antarctic Peninsula, is characterised by consistently parallel bedforms and an absence of transverse moraines and cross-cutting lineations (Ó Cofaigh *et al.*, 2008). The glacial geomorphology of the Tweed Valley and NNCP show similarities to this, in that bedforms are consistently parallel and display a low degree of modification. There is also no geomorphological evidence to support episodic retreat; no terminal moraines or transverse features are identified at any distance along the TIS track. There is however, as mentioned above, evidence of localised cross-cutting, implying flow switching during deglaciation. Ó Cofaigh *et al.* (2008) suggest this is indicative of slower retreat. Despite this, it is proposed that retreat of the TIS was relatively fast, as the streamlined subglacial bedforms show little sign of modification, which is expected during slower retreat. Full retreat of the TIS is considered likely to have occurred by 15 ka BP, as the onshore margin is recorded north of the Scottish Border at this time (Huddart, 2002a).

5.2 The Bradford Interlobate Complex

The Bradford complex on the NNCP is comprised of eskers, crevasse fills, kames and subaqueous fans, which together form an elongate complex of sand and gravel over 10 km long (see section 4.2.3). The complex is located approximately parallel to regional ice flow indicators, as is inferred from the streamlined bedforms to the west and east. To the west, streamlined bedforms reveal that ice flow was oblique to the main N-S trend of the complex, whereas to the east it was more parallel, implying ice flow was broadly convergent here. It is argued here that this relationship between these landform assemblages, which not previously been recognized, is key in determining the origin of the Bradford Complex. In this section, the implications of the geomorphological data for the mode of genesis of this complex and the style of deglaciation will be discussed.

5.2.1 Genesis of the Bradford Complex

In terms of the origin of the Bradford Complex, it is proposed that deposition occurred in a series of subglacial feeder conduits and proglacial lakes that formed during ice retreat (see section 4.2.3.2). During deglaciation, kame and kettle topography was formed amongst the eskers and subaqueous fans, resulting in the distinct hummocky appearance of the complex. Similar palaeo-environmental settings have been invoked at the Bradford Complex by other researchers (c.f. Parsons, 1966, cited in Huddart, 2002b). Parsons (1966) concluded that the series of eskers formed in closed conduits as ice retreated towards the north. This was developed by Huddart (2002b), who proposed that esker formation was initiated in subglacial conduits discharging into ice-dammed proglacial lakes. Whilst the broad palaeo-environmental setting is agreed on in this study (i.e. proglacial lakes and subglacial feeder conduits), the glacial configuration and dynamics are suggested to have differed. Both Huddart (2002b) and Parsons (1966) proposed that the Bradford Complex formed during stillstands of the ice margin as ice retreated to the north through the Warren Gap. However, evidence presented in this study show that retreat of the TIS along the NNCP was continuous; there is no evidence that retreat was punctuated by stillstands (see section 5.1.4). As mentioned above, the

Bradford Complex is found in close association with adjacent streamlined bedforms, the orientation of which suggest flow on the complex was convergent. It is proposed in this study that the TIS and NSL were convergent and that the growth of the NSL diverted the TIS southwards along the NNCP (see section 5.1.1). Given the suggestion that the NSL flowed for a short distance onshore in Northumberland and further south in County Durham (Teasdale and Hughes, 1999, Davies *et al.*, 2009), it is proposed here that the Bradford Complex was formed in the convergence zone between these two ice streams.

Morphologically similar elongate glaciofluvial complexes have been identified in the convergence zones of palaeo-ice streams of the Fennoscandian Ice Sheet (c.f. Punkari, 1908, 1997; Mäkinen, 2003), the Cordilleran and Laurentide Ice Sheets (c.f. Fraser, 1993; Brennand and Shaw, 1996; Brennand *et al.*, 1996; Hicock and Fuller, 1996) and the British-Irish Ice Sheet (c.f. Warren and Ashley, 1994, 1997; Thomas and Montague, 1997; Huddart and Bennett, 1997; Delaney, 2001, 2002). These complexes, known as interlobate complexes or interlobate moraines (c.f. Punkari, 1980), are the product of focussed glacial drainage and glaciofluvial ice-contact deposition between two converging ice masses within polycentric ice sheets (Brennand and Shaw, 1996; Punkari, 1997; Huddart and Bennett, 1997; Thomas and Montague, 1997). In terms of morphology, these complexes are similar to the Bradford Complex. For example, the Tampere interlobate zone in southern Finland is an elongate complex formed of a central esker ridge flanked by kettle holes, glaciofluvial hummocks and deltas (Punkari, 1997). Flow indicators also converge on this complex (Punkari, 1997). Based on a comparison with these complexes, insights into the genesis of the Bradford Complex are provided.

It is proposed that glacial drainage and glaciofluvial deposition were focussed in the zone of convergence between the TIS and NSL on the basis that with convergent flow, ice surface profiles are approximately concave (Hughes, 1981). As glacial drainage within ice sheets is in large parts controlled by the ice surface slope (Shreve, 1972), meltwater flow within the TIS and NSL would have been towards the confluence zone. Meltwater would also have been focussed towards the bed through surface crevasses. The high velocity gradients that exist across interlobate zones would also have promoted melting in the shear-heated converging ice (Punkari, 1997), leading to a well-established subglacial drainage network (Warren and Ashley, 1994) (Fig. 58, page 81). As the TIS and NSL separated along their confluence zone, it is proposed that deposition was initiated in these subglacial conduits as lakes formed proglacially (Fig. 58, page 81).

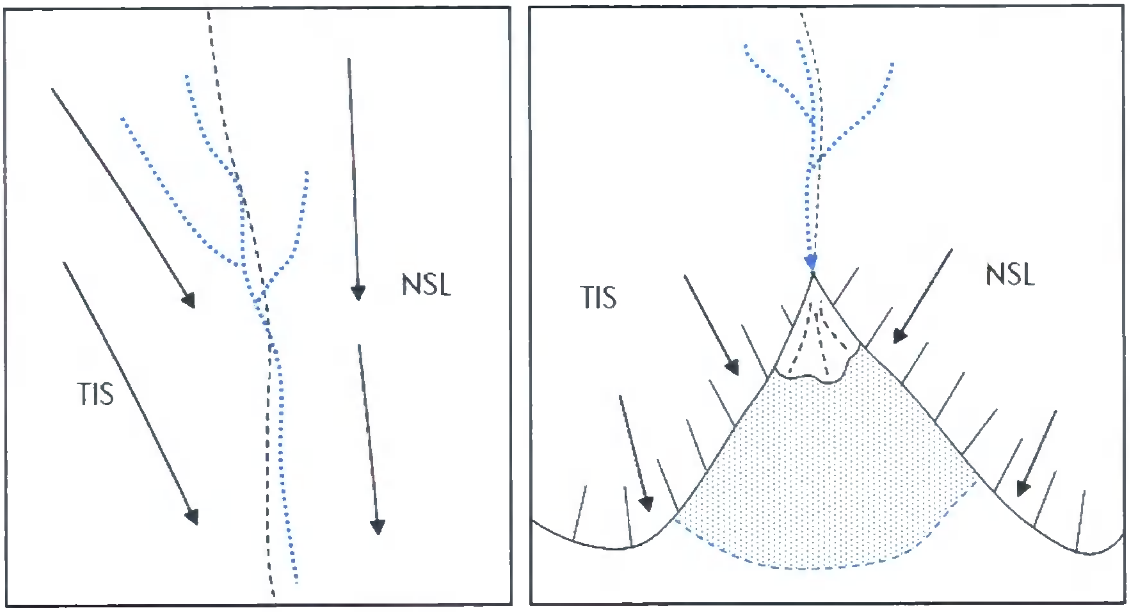


Figure 58: *Proposed formation of the subaqueous fans of the Bradford Complex.* ***LEFT:*** *convergent flow of the Tweed Ice Stream and North Sea Lobe, with subglacial drainage system (blue dashed line) developing in strain-softened convergent zone.* ***RIGHT:*** *Development of interlobate lake as TIS and NSL separate. Subaqueous fan is indicated to have formed here. Flow trajectories are shown as black lines.*

A similar depositional setting is invoked for the eskers of Irish midlands, where a large interlobate lake is suggested to have formed between two retreating ice centres (c.f. Warren and Ashley, 1994). Delaney (2002) alternatively proposed that the Irish Eskers formed beneath a narrow marginal zone of stagnating, 'sluggish' ice that terminated in a large lake where water pressures were high enough to keep the subglacial conduits open. In both scenarios, a time-transgressive genesis is proposed (Warren and Ashley, 1994; Delaney, 2002). A time-transgressive mode of deposition is also favoured for the Bradford Complex. Unlike at the Harricana glaciofluvial Complex (Brennand and Shaw, 1996), the expected geomorphic characteristics of synchronous deposition are not observed at Bradford. The sub-aqueous fans and associated discontinuous eskers at Bradford have been interpreted as evidence of deposition in subglacial conduits, discharging into proglacial lakes (see section 4.2.3.2). If deposition of the Bradford Complex had occurred synchronously, it is difficult to account for the fans, as (based on the assumption that they formed sub-aqueously) standing bodies of water are required at the ice-margin. Ice-contact fans also form periodically (c.f. McCabe and Eyles, 1988; Warren and Ashley, 1994), and therefore suggest that synchronous deposition is unlikely as a formative mechanism for the Bradford Complex.

The distribution and morphology of the eskers, kames and subaqueous fans of the Bradford Complex provide evidence on the style of deglaciation. The discontinuous nature of the eskers throughout the complex supports the proposition that glacial retreat in this region was relatively rapid (see section 5.1.4). This is based on morphological comparison with the discontinuous, beaded eskers in central Ireland, the length of which was limited by rapid marginal retreat (Warren and Ashley, 1994). The lack of ice-transverse features here, as along the TIS track, are evidence of continuous retreat (c.f. Ó Cofaigh *et al.*, 2008). It is noted, however, that the degree of post-depositional erosion of the Bradford eskers is unknown, and their discontinuous form may be a result of subaerial stream erosion, slumping or human activity. As it is proposed that eskers form within active ice (Warren and Ashley, 1994; Punkari, 1997), and based on the above, it is likely that the Bradford complex formed within relatively rapid, continuously northwards retreating active ice. The presence of kames and kettle holes in close association with the eskers are evidence that stagnating, sediment laden ice was left as the margin retreated.

There is strong evidence to suggest that the Bradford Complex was formed in the interlobate zone between the TIS and NSL. Based on the above, the formation of this glaciofluvial and glaciolacustrine complex is inferred to be as follows:

(1) In the convergence zone of the TIS and NSL, glacial flow and meltwater drainage converged, leading to the formation of an subglacial drainage system.
(2) As the TIS and NSL separated during deglaciation, lakes ponded in the interlobate zone between these two ice masses into which the subglacial conduits discharged. Deposition was initiated within the conduits and subaqueous fans developed (c.f. Warren and Ashley, 1994). This complex formed time transgressively (c.f. Brennand and Shaw, 1996), with deposition occurring from south to north as the ice margin retreated. Retreat was continuous and relatively rapid, as is inferred from a lack of flow-transverse features along the TIS track (c.f. Ó Cofaigh *et al.*, 2008) and the discontinuous eskers at the northern end of the complex.
(3) As the ice retreated to the north, sediment-covered stagnating ice melted, forming the kettle holes and hummocky moraine.

5.3 Geomorphology and Sedimentology of the TIS lateral margins

Extensive zones of hummocky terrain have been mapped around Cornhill and Wooler (see sections 4.2.1, 4.2.2). These two complexes exhibit stark similarities; both are comprised of subglacial eskers, crevasse fills, supraglacial kames, kame terraces, ice-contact deltas and subaqueous fans and are found on the lower flanks of the Cheviots in close association with the mapped meltwater channels. Morphologically, these landforms are similar to those of the Bradford Interlobate Complex, however, they differ markedly from the latter with respect to their location situated along the lateral margins of the TIS track. The location of these glaciofluvial complexes and their relationship with the surrounding landforms is key for determining their origin.

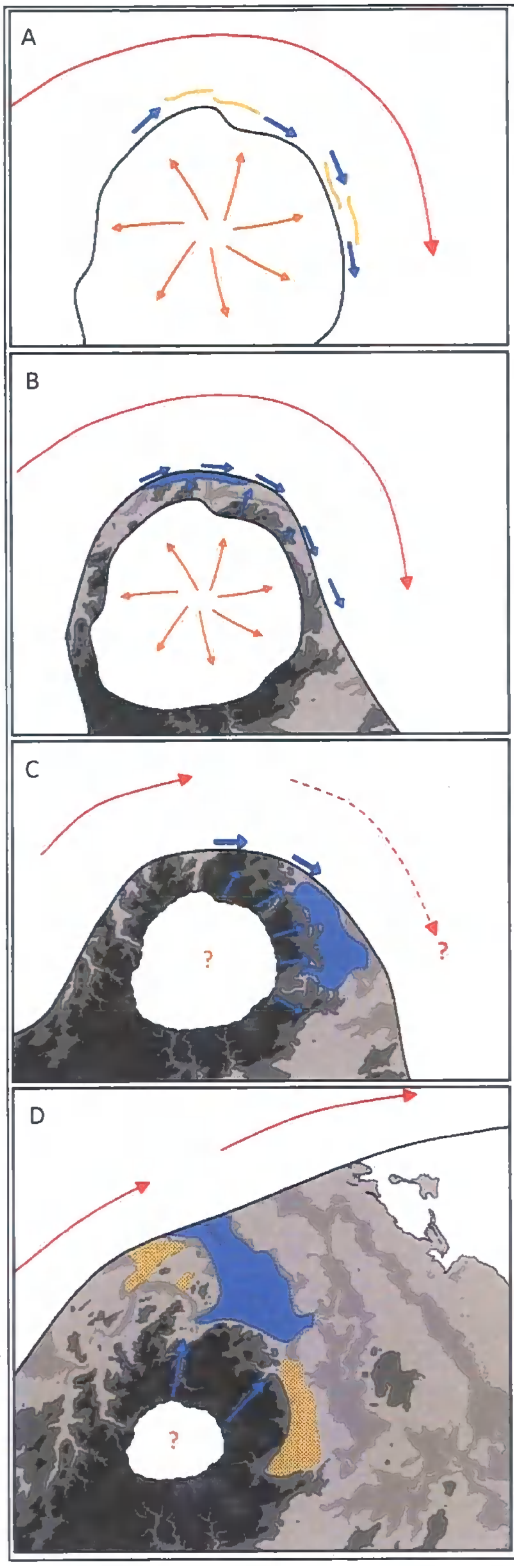
The Wooler and Cornhill complexes are interpreted to have formed during deglaciation. This is based on several lines of evidence. Firstly, the Cornhill sands and gravels are superimposed over drumlins towards the southwest of the complex (as observed by Gunn and Clough, 1895 and Carruthers *et al.*, 1932), which is evidence the Cornhill Complex formed after ice streaming and drumlin formation. This is in agreement with Clapperton (1971a). As the TIS is proposed to have flowed during a late stage of deglaciation (Everest *et al.*, 2005), this supports the deglacial origin of the Cornhill Complex. Secondly, the extensive spreads of glaciofluvial deposits and landforms suggests high volumes of meltwater were present (Sissons, 1973). Thirdly, the abundance of meltwater channels in the Cheviots is also evidence that a well-organised subglacial drainage system existed here (Golledge and Stoker, 2006). The association of these complexes with deglaciation is in agreement with previous researchers, who have linked the glaciofluvial complexes to ice-directed meltwater drainage from the Cheviots and deposition in a stagnating ice body (c.f. Carruthers *et al* ., 1927, 1932; Clapperton, 1968, 1970, 1971a, 1971b).

Also in agreement with previous researchers is the assertion that the Wooler and Cornhill Complexes are part of the same glaciofluvial system (c.f. Carruthers *et al.*, 1932). This is based on their comparative morphology, sedimentology and location. The complexes are separated from each other by the Milfield Plain, which was occupied by Glacial Lake Ewart that formed during deglaciation as the Tweed ice had retreated from the Cheviots (Clapperton, 1971; Evans *et al.*, 2005). The former presence of this lake is recorded by glaciolacustrine sediments, lake shorelines and glaciolacustrine deltas (Butcher, 1967). Based on the distribution of glaciofluvial deposits through the Glenn Valley and at the southernmost limit of the Milfield Plain, it is speculated that the Wooler and Cornhill complexes were connected prior to the development of Lake Ewart. If indeed these Cornhill-Wooler Complex (as it shall be called) was initially more extensive than at present, it implies that a major drainage system was in operation in this area during deglaciation.

Based on the location of the Cornhill-Wooler Complex at the lateral margins of the TIS, it is proposed that this drainage system developed in the lateral shear zone at the margin of the TIS. A shear zone would have formed between the differentially flowing TIS and the cold-based Cheviot Ice Cap, the latter of which is inferred to flowed radially from the highest summits to coalesce with Tweed ice (Clapperton, 1970). This interpretation is based on analogy with the lateral margins of many modern ice streams which are characterised by heavy en-échelon crevassing (LeB. Hooke, 2005). The shear margins of Jakobshavn Isbræ South Ice Stream in western Greenland are characterised by open, extensional crevasses with rounded edges and highly degraded surfaces (Mayer and Herzfeld, 2000). Meltwater will penetrate to the bed through these crevasses, which, together with high basal melt rates, leads to the development of extensive subglacial drainage networks (Paterson, 1994). It is considered likely that such a situation would have occurred between the TIS and the Cheviot Ice Cap. The association of this drainage system with the TIS is supported by the orientation of the eskers of the Cornhill-Wooler Complex. As flow of water within subglacial conduits is controlled by the ice surface slope and the pressure gradient within the ice (Warren and Ashley, 1994; Johansson, 2003) the orientation of eskers can be used to reconstruct ice flow directions. The orientation of eskers within these two complexes indicate that ice flow was E-W in the Cornhill Complex and N-S in the Wooler Complex. This is in broad agreement with regional ice flow indicators, which suggest flow of the TIS was around the Cheviots in a broad arc (see section 4.1.2, also Clapperton, 1970).

It is proposed that the Cornhill-Wooler Complex formed time-transgressively. As esker deposition is only likely within active ice (Warren and Ashley, 1994; Punkari, 1997), and the eskers mapped here are predominantly orientated parallel to flow of the TIS, deposition of these eskers must have been initiated prior to ice separation and stagnation (Fig. 59a, page 85). As the TIS and the Cheviot ice cap separated, the suture zone would have acted as a focus for glacial and subaerial drainage. The well-established subglacial drainage network, as evident from the eskers at both Cornhill and Wooler, would likely have discharged into this ice-free area, further focussing glaciofluvial drainage. At this time, the Cornhill kame terraces at Moneylaws Hill are proposed to have formed (Fig. 59b, page 85), along with the ice-contact fans at Roddam and the ice-contact delta at Kilham. The latter of these, as it is formed in a proglacial water body, is only likely to have formed once Cheviot ice had retreated from the Glenn Valley. Hummocky moraine and supraglacial kames would have been deposited as the ice downwasted (Fig. 59c, page 85), with kettle holes forming as debris-covered ice stagnated (Fig. 59d, page 85).

It is proposed that the majority of Cheviot meltwater channels also formed during deglaciation as the TIS retreated and the Cheviot Ice Cap downwasted. Clark *et al.* (2006) proposed that meltwater channels must develop in the ablation zones of ice sheets during retreat phases as large volumes of meltwater are required for their formation. The abundance of meltwater channels is, as mentioned above, indicative of a well-organised sub-glacial drainage system (Golledge and Stoker, 2006). Four channel types have been identified: ice-marginal channels,



A: Period of ice-directed drainage and esker formation

The TIS and Cheviot Ice Cap at this stage are inferred to have been confluent around the lower flanks of the Cheviots. The lateral margin of the ice stream is delineated by the boundary between streamlined and non-streamlined ground and the location of eskers, inferred to have formed in the lateral shear zone between these two ice masses.

B: Period of ice separation and ice-marginal channel and kame terrace formation

The Cheviot and TIS separate along their zone of confluence, leading to the formation of kame terraces at Moneylaws Hill and ice-marginal meltwater channels in the Cheviots (Type A Channels, see Fig. 63). Submarginal chutes form as subaerial and glacial meltwaters plunge beneath the margin of the TIS. Erosion of subglacial channels and superimposed meltwater channels, initiated by the downwasting of Cheviot and Tweed ice.

C: Period of proglacial channel erosion, glaciofluvial deposition and ice-dammed lake formation

As the TIS retreats further, meltwater and subaerial waters flow from the Cheviots. Deposition of the Roddam Bog ice-contact fan and the ice-contact delta in the Glenn Valley are proposed to have occurred during this stage, and also supraglacial kames. Glacial lakes form in the ice-free gap between the Cheviot Ice Cap and the retreating TIS (c.f. Clapperton 1971b).

D: Final stage of deglaciation

It is hypothesised that following retreat of the NSL (~15 ka BP, see section 2.1) resulting in flow of the TIS offshore into the NSB. The Cheviot Ice Cap has most likely fully retreated by this time (e.g. Clapperton, 1971a). The formation of Glacial Lake Ewart during this late stage of deglaciation on the Milfield Plain is evidence that the TIS was blocking regional drainage routes (c.f. Lunn, 1980). The Cornhill-Wooler Complex is suggested to have been partly destroyed by the development of this lake (c.f. Carruthers et al., 1930). Kettle holes formed as debris-covered stagnating ice melted.

***Figure 59:** Time-Transgressive formation of the Cornhill-Wooler Glaciofluvial complex. Red arrows denote flow of the TIS, orange arrows the flow of the Cheviot Ice Cap. Glacial lakes are shown in sold blue, beige areas show glaciofluvial deposits. Blue arrows show direction of meltwater and subaerial water flow.*

proglacial or submarginal chutes, superimposed englacial channels and subglacial channels (see section 4.3). The distribution of these and the relationship between them provide information on their formation and also the style of deglaciation. As meltwater flow within ice will broadly flow parallel to the steepest ice surface profile (Shreve, 1972; Clark *et al.*, 2006), subglacial and superimposed channels can be used to reconstruct ice flow directions. In the north of the Cheviots, superimposed channels are predominantly orientated towards the southeast, which is broadly parallel with flow of the TIS. It is therefore speculated that these ice-directed channels were formed by the superimposition of englacial conduits that had formed in the strain-softened region between the TIS and Cheviot Ice Cap. Subglacial channel systems indicate close ice control (Lunn, 1980) and, together with proglacial channels form a radial flow pattern, indicating flow was directly off the Cheviots. It is possible that the largest, most well developed subglacial channel systems, such as The Trows, formed during several episodes of downwasting and were enlarged by subaerial waters during interglacials. As the ice-marginal and proglacial channels would have required an ice-free slope on which to form, these channels most likely represent the most recent stage of channel formation during deglaciation when an ice-free zone existed between the TIS and Cheviot Ice Cap.

5.4 Regional Glaciological Implications

Geomorphological mapping of the Tweed Valley and the surrounding area has shown that three landform assemblages dominate the glacial geomorphological record; streamlined subglacial bedforms, meltwater channels and glaciofluvial complexes (see chapter 4). The identification of these has led to the proposition that an ice stream flowed along the Tweed Valley and coalesced with the North Sea Lobe along the North Northumberland Coastal Plain. In the interlobate zone between these two ice masses, the Bradford Complex was formed during deglaciation. At the southern lateral margin of the TIS, the Cornhill-Wooler Complex formed as the TIS retreated from the lower flanks of the Cheviots. In this section, the implications of these data for the regional ice dynamics, the dynamics of the BIS and the wider glaciological implications will be discussed.

Regarding the Devensian ice configuration of the region, there is debate as to whether the Cheviots supported an independent ice cap. Early theories suggested foreign ice had overridden the summits (Geikie, 1876, cited in Clapperton, 1970), whereas Clapperton (1970) proposed the existence of an independent ice cap based on glacial erratics, tills and striations throughout the Cheviots. More recent work by Harrison *et al.* (2006) argued that periglacial landforms (e.g. solifluction sheets, tors and scree-slopes) in the Cheviots were evidence that the Cheviots were not covered by an ice cap during the Devensian. This latter view is rejected on the basis that geomorphological data presented in this study provide strong evidence that the Cheviots were occupied by a radially-flowing ice cap that coalesced with the TIS on its northern flanks. The strongest evidence to support this ice configuration pattern comes from the organisation of meltwater channels (see section 4.3) and the sedimentological data from Roddam in the Wooler Complex (section 4.2.1.2). As mentioned in section 5.3, the organisation of subglacial and proglacial meltwater channels suggest ice flow was radial from the summits of the Cheviots. Palaeocurrent data from the ice-contact fan at Roddam also support this theory, as ripple and clast orientation suggests meltwater flow was from the NW, i.e. directly off the Cheviots (see section 4.2.1.2). The complex interbedded sand and gravel troughs and prograded channel fills are indicative of abundant meltwater and sediment flow from the Cheviots. The lack of extensive glacial erosion in the Cheviots was suggested by Clapperton (1970) to indicate that the Cheviot ice cap was cold-based (Clapperton, 1970). Whilst the presence of subglacial meltwater channels in the Cheviots would appear to contradict this, the subglacial channels here are interpreted to have formed during a late stage of glaciation (see section 4.3.4.2, page 70). Furthermore, the majority of channels in the Cheviots are associated with flow of the TIS around the Cheviot Massif (see sections 4.3.1., page 64 and 4.3.2, page 66).

Streamlined bedforms along the Tweed Valley and NNCP implies fast-flowing, warm based ice (see section 5.1). These spatially varying zones of glacial activity are evidence that the BIS in this region was polythermal, with a thawed-bed (warm-based) corridor of ice in the Tweed Valley bordered by slower-flowing, frozen-bed (cold-based) ice over the Cheviots. Similar, spatially varying zones of glacial activity (erosion) have been observed in the Cairngorm Mountains in Scotland (Hall and Glasser, 2003), in NW Canada (Kleman and Glasser, 2007) and in Greenland (Sugden, 1974). In these regions, the distribution of frozen and thawed bed zones appears to be topographically controlled, with the majority of frozen-bed patches found on topographic highs (Hall and Glasser, 2003; Kleman and Glasser, 2007). It has been proposed in section 5.1.2 that the TIS was topographically controlled, with the lateral margins of the ice stream controlled by the topographic jump where the Tweed Valley meets the Cheviots and Lammermuir Hills. Across these thermal boundaries between the TIS and the slower-flowing Cheviot ice, the predicted high velocity gradients are inferred to have resulted in strain heating and the development of extensive subglacial drainage networks (see section 5.3). The Cornhill-Wooler Complex formed during deglaciation in this zone. The interpretation of this complex as the TIS lateral margin suture-zone signature has implications for the interpretation of glaciofluvial complexes found at palaeo-ice stream lateral margins. The interpretation of the

Bradford Kame as an interlobate complex formed the convergence zone between the TIS and NSL (see section 5.2) has highlighted the importance of a landform assemblage approach; the close spatial association of the streamlined bedforms and glaciofluvial and glaciolacustrine deposits of the Bradford complex has not previously been accounted for.

This study has provided a new insight into the late Devensian history of the Tweed Valley and north-east Northumberland. The highly attenuated streamlined bedforms along the NNCP and cross-cutting flow-sets southwest of Berwick-upon-Tweed have been interpreted as evidence of the diversion of the TIS by the NSL. This diversion of Tweed ice along the North Sea coast has been previously suggested at (c.f. Lunn, 1980; Teasdale and Hughes, 1999), although evidence to support it was limited; flow trajectories were largely based on erratic distribution along the NNCP. The distribution and orientation of bedforms mapped in this study provide evidence of the interaction of two competing ice masses along the coast – the TIS and the NSL. The Bradford Interlobate Complex is interpreted as the onshore signature of their confluence and subsequent separation during deglaciation.

Dates generated from studies into the dynamics of the NSL suggest that ice advanced into the NSB around 15 ka BP (and possibly earlier) (e.g. Nygard *et al.*, 2007; McCabe *et al.*, 2005) in response to widespread cooling in the Northern Atlantic brought about by Heinrich 1 (e.g. McCabe *et al.*, 1998; 2005). It is therefore proposed that the Bradford Interlobate Complex formed after 15 ka BP and prior to 13 ka BP, when the southerly limit of the BIS is inferred to have been north of the Scottish Border (Lunn, 1980; 1995). The interaction between the NSL and ice flowing from the Cheviots, Southern Uplands and the Scottish Highlands has been observed in County Durham, where lodgement tills are interpreted as evidence of a highly dynamic, multi-lobate ice sheet (Davies *et al.*, 2009). Combined with the findings of this study, which has provided substantial evidence of the diversion of the TIS in northeast Northumberland, our knowledge of the interactions of the NSL and ice flowing from onshore accumulation centres in the northeast of England is greatly improved.

Following the retreat of the NSL, the TIS is suggested to have flown directly offshore for a short period of time, as is inferred from the orientation of bedforms southwest of Berwick-upon-Tweed. Through comparison of the geomorphological imprint of the TIS with other palaeo-ice streams (e.g. Stokes and Clark, 2003; Stokes *et al.*, 2006), it is suggested that the mechanism that led to ice-stream shut down was relatively rapid. This is based on a lack of geomorphological evidence to support a narrowing and a decrease in velocity of the TIS, which has been similarly observed at palaeo-ice stream locations in eastern Canada (e.g. Stokes and

Clark, 2003). Everest *et al.* (2005) proposed that the TIS may have propagated deglaciation of this region through the draw-down of ice from the higher ground in the upper reaches of the Tweed Valley. If this situation had occurred, it is envisaged that the TIS would have switched off once the volume of ice upstream was greatly reduced, i.e. when the ice stream had 'run out' of ice (c.f. Stokes and Clark, 2003). Everest *et al.* (2005) also estimated that the TIS drained approximately 3500 km^2 of the BIS. The data presented in this study, however, show the TIS was more spatially extensive than initially thought (c.f. Everest *et al.*, 2005), and therefore, it is assumed the TIS drained a larger area of this sector of the BIS. This is somewhat speculative and there are still many questions surrounding the timing and dynamics of the TIS, particularly the mechanisms triggering ice streaming and subsequent shut down. Furthermore, few dates exist to constrain these events in the Tweed Valley and northeast Northumberland. This study has, however, improved our knowledge on a region that has been relatively neglected in terms of glaciological research, and opens the way for more detailed investigation of this sector of the British-Irish Ice Sheet.

6

CONCLUSION

6.1 Summary

The aim of this project was to investigate the glacial geomorphology of the Tweed Valley and surrounding area with a view to reconstructing the regional glacial history. To achieve this aim, the four objectives set in section 1.3 were met. A review of the existing literature on the field site (objective 1) showed a heavy focus on the description of the glacial geomorphology and sedimentology of selected sites throughout the region, with little emphasis on reconstructing the regional ice dynamics. In order to reconstruct the regional glacial history (objective 4), geomorphological mapping of the Tweed Valley and the surrounding area from NEXTMap DEMs (objective 2) and sedimentological surveys at selected sites (objective 3) were carried out. The results of these revealed the dominance of three glacial landform assemblages:

(1) Highly attenuated, closely-spaced subglacial streamlined bedforms in a broad track around the Cheviot massif;
(2) Glaciofluvial complexes at Wooler, Cornhill and Bradford comprised of eskers, ice-contact (kame) terraces, subaqueous deltas and ice-contact fans and supraglacial kames.
(3) Subglacial, superimposed, ice-marginal, proglacial and submarginal meltwater channels in the eastern Cheviot massif and Lammermuir Hills.

The spatial relationships and distribution of these landforms has enabled several inferences to be made on the glacial history of the Tweed Valley and surrounding area.

6.2 Conclusions

On the basis of the glacial geomorphological record of the Tweed Valley and surrounding area, the following conclusions are reached regarding the regional ice dynamics.

(1) Streamlined subglacial bedforms along the Tweed Valley and NNCP have been associated with fast ice flow. Supportive evidence of a Tweed Ice Stream from the geomorphological record include highly convergent flow patterns, highly attenuated,

closely-spaced streamlined bedforms, sharp lateral margins and characteristic shape and dimensions. The high parallel conformity of the mapped bedforms within each of the two flow-sets and lack of reworking of bedforms are indicative of isochronous (rapid) formation. The cross-cutting flow-sets southwest of Berwick-upon-Tweed has led to suggestion that the TIS was diverted along the NNCP by the growth of the NSL in the North Sea Basin during the Late Devensian. Following retreat of the NSL, the TIS flowed directly offshore in the lower Tweed Valley.

(2) The onshore signature of the confluence zone between the NSL and the TIS is recorded by the Bradford Interlobate Complex. Ice flow at this elongate sand and gravel complex was broadly convergent, as is implied by the orientation of adjacent streamlined bedforms. Focussed glacial drainage at this zone resulted in the development of an interlobate subglacial drainage network. Deposition of the Bradford eskers was time-transgressive as the ice masses separated along their zone of confluence and was most likely initiated in subglacial conduits discharging into proglacial lakes.

(3) The TIS is interpreted to have been largely topographically-controlled. This is evident from the location of the TIS lateral margins in the Tweed Valley, which coincide with the higher ground of the Cheviots and Lammermuir Hills. Retreat of the TIS is believed to have been continuous and relatively rapid, as is evident from the lack of flow-transverse features along the TIS and NNCP.

(4) At the TIS lateral margins, high rates of shear between the streaming ice and cold-based ice of the Cheviots resulted in high rates of strain heating and the development of an extensive subglacial drainage system. During deglaciation, as these ice masses separated, glaciofluvial deposition occurred time-transgressively in the newly ice-free region between the TIS and Cheviot Ice Cap, leading to the formation of the Cornhill-Wooler Complex.

(5) During deglaciation, meltwater channels formed as the TIS and Cheviot Ice Cap downwasted/retreated and a well-organised subglacial drainage system developed. The radial organisation of subglacial and proglacial meltwater channels in the Cheviots have been used to support the existence of a Cheviot Ice Cap. This interpretation is in agreement with previous researchers who proposed the Cheviots were covered by a cold-based ice cap that coalesced with faster-flowing ice on its lower flanks (e.g.

Clapperton, 1970). Channels orientated parallel to flow of the TIS are evidence of englacial channel superimposition in a down-wasting ice sheet.

(6) Spatial varying zones of glacial activity in the Tweed Valley, NNCP and Cheviot massif suggest that a polythermal ice sheet existed during the Late Devensian in northeast England, with a warm-based, fast flowing ice stream in the Tweed Valley and NNCP and a cold-based ice cap over the Cheviots. This is evident from the sharp boundary between streamlined terrain in the Tweed Valley and NNCP (indicative of warm-based, fast-flowing ice) and the non-streamlined terrain of the Cheviot massif and Lammermuir Hills.

(7) Geomorphological mapping from NEXTMap Britain DEMs has proved an invaluable tool for the reconstruction of the ice dynamics of the Devensian Ice Sheet in the Tweed Valley and surrounding areas. It has been shown that this sector of the BIS was far more dynamic than originally thought, which has wider implications for the entire BIS.

6.3 Recommendations for further research

One of the main limitations of this study was the lack of sedimentary exposures throughout the region. Although this has not prevented inferences being made on the genesis of landform assemblages nor the regional ice dynamics, further sedimentological investigations would improve the robustness of these reconstructions. As several landforms of the Bradford Interlobate Complex and the Cornhill-Wooler Complex are listed as SSSI's (e.g. the Bradford Kames, Campfield Kettle Holes, The Trows, Roddam Dene), non-invasive methods are required to establish the internal structure (stratigraphy and sedimentology) of these landform assemblages. The use of Ground Penetrating Radar (GPR) is recommended, as radar reflection profiles can be used to identify bedding structures and deposits of varying densities (Neal, 2004). GPR surveys of the streamlined bedforms of the TIS track would also be invaluable to future reconstructions, as the results from this may shed light on whether the drumlins and flutings are rock-cored or show evidence of subglacial sediment deformation.

The TIS is inferred to have terminated an unknown distance offshore in the North Sea Basin (see section 5.1). Geophysical surveys of bathymetric troughs of the Antarctic Continental Shelf have provided information on the submarine imprints of ice streams that retreated during the Holocene (Ó Cofaigh *et al.*, 2008). It is suggested that similar geophysical surveys of the North Sea Basin offshore of Berwick-upon-Tweed would provide important information on the configuration of the TIS (i.e. its offshore extent and location of its terminus) and the style of retreat, strengthening the reconstructions of this sector of the BIS. It is also proposed that in order to fully understand the regional ice dynamics, the signature of the Solway Ice Stream, which is inferred to have been confluent with the Cheviot Ice Cap towards the south of the Cheviot massif (Clapperton, 1970b; Lunn, 1995), should be investigated. It has been proposed that Solway ice was diverted by the North Sea Lobe (c.f. Sissons, 1964, Huddart and Glasser, 2002) as in the case of the TIS. Given the value of the glacial geomorphological record for the reconstruction of the TIS, an assessment of the glacial geomorphology of the Solway Ice Stream would undoubtedly strengthen reconstructions of the configuration and dynamics of the BIS in northeast England.

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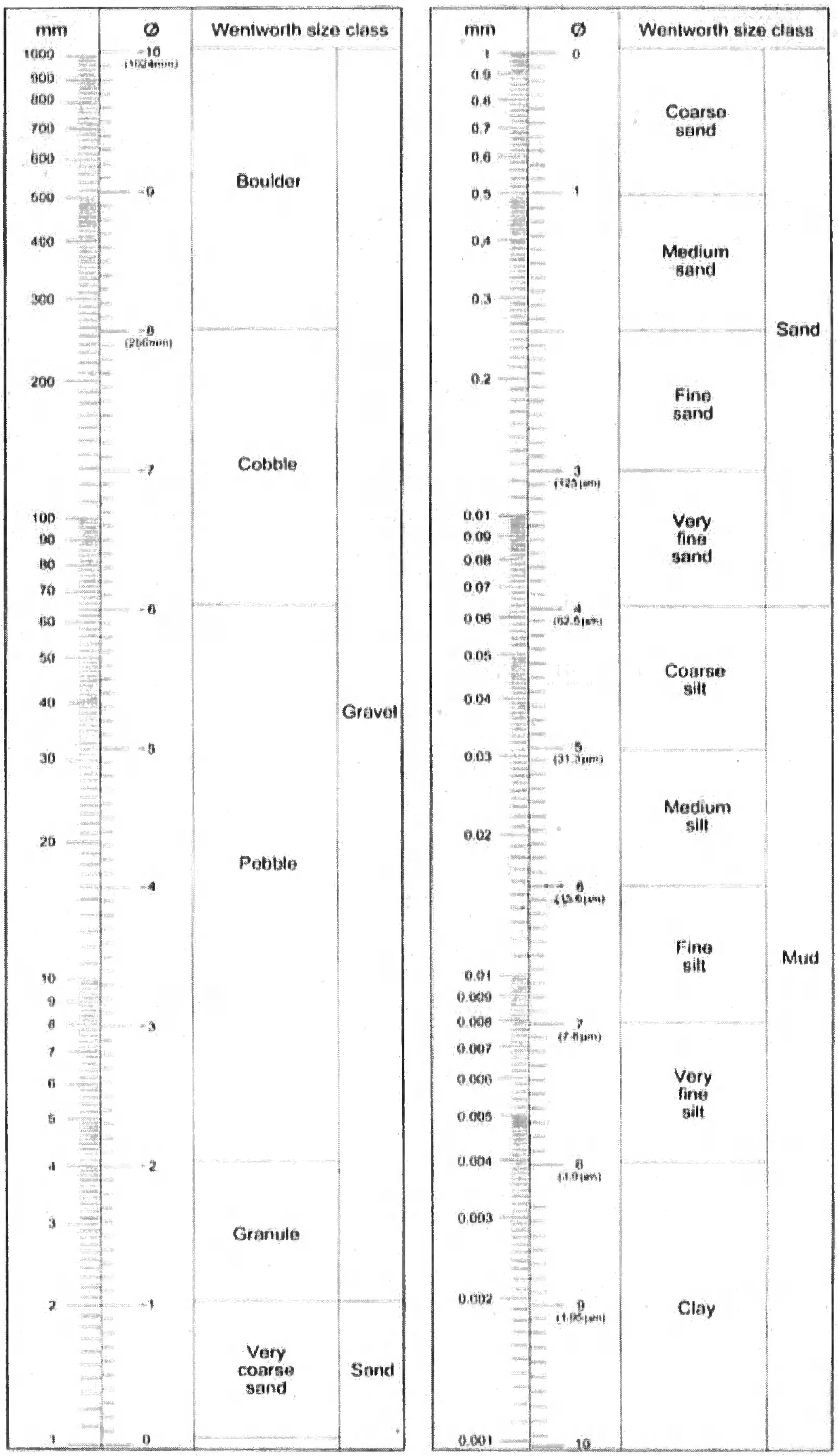
Appendix A: Lithofacies Codes (from Evans and Benn, 2004)

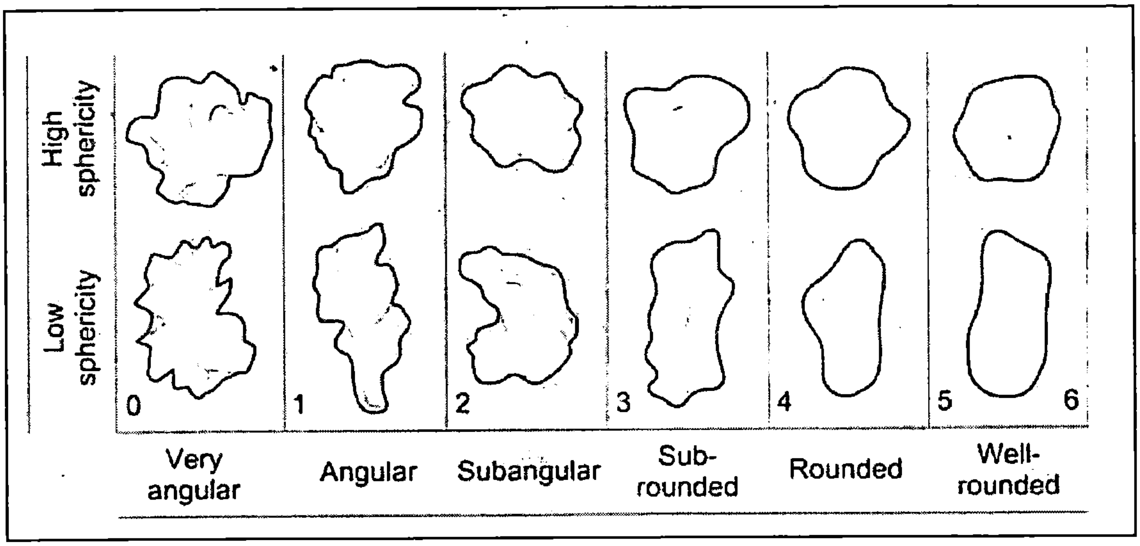
Code	Description
Diamictons	*Very poorly sorted admixture of wide grain-size range*
Dmm	Matrix-supported, massive
Dcm	Clast-supported, massive
Dcs	Clast-supported, stratified
Dms	Matrix-supported, massive
Dml	Matrix-supported, laminated
--- (c)	Evidence of current reworking
--- (r)	Evidence of re-sedimentation
--- (s)	Sheared
--- (p)	Includes clast pavement(s)
Boulders	*Particles > 256mm (b-axis)*
Bms	Matrix-supported, massive
Bmg	Matrix-supported, graded
Bcm	Clast-supported, massive
Bcg	Clast-supported, graded
Bfo	Deltaic foresets
BL	Boulder lag or pavement
Gravels	*Particles of 8-256mm*
Gms	Matrix-supported, massive
Gm	Clast-supported, massive
Gsi	Matrix-supported, imbricated
Gmi	Clast-supported, massive (imbricated)
Gfo	Deltaic foresets
Gh	Horizontally bedded
Gt	Trough cross-bedded
Gp	Planar cross-bedded
Gfu	Upward-fining (normal grading)
Gcu	Upward-coarsening (inverse grading)
Go	Openwork gravels
Gd	Deformed bedding
Glg	Palimpsest (marine) or bedload lag
Granules	*Particles of 2-8mm*
GRcl	Massive with clay laminae
GRch	Massive and infilling channels
GRh	Horizontally bedded
GRm	Massive and homogeneous
GRmb	Massive and pseudo-bedded
GRmc	Massive with isolated outsize clasts
GRmi	Massive with isolated, imbricated clasts
GRmp	Massive with pebble stringers
GRo	Open-work structure
GRruc	Repeating upward-coarsening cycles
GRruf	Repeating upward-fining cycles
GRt	Trough cross-bedded
GRcu	Upward coarsening
GRfu	Upward fining
GRp	Cross-bedded
GRfo	Deltaic foresets

Code	Description
Sands	*Particles of 0.063-2mm*
St	Medium to very coarse and trough cross-bedded
Sp	Medium to very coarse and planar cross-bedded
St (A)	Ripple cross-laminated (type A)
Sr (B)	Ripple cross-laminated (type B)
Sr (S)	Ripple cross-laminated (type S)
Scr	Climbing ripples
Ssr	Starved ripples
Sh	Very fine to very coarse and horizontally/plane bedded or low angle cross-lamination
Sl	horizontal and draped lamination
Sfo	Deltaic foresets
Sfl	Flasar bedded
Se	Erosional scours with intraclasts and crudely cross-bedded
Su	Fine to coarse with broad shallow scours and cross-stratification
Sm	Massive
Sc	Steeply dipping planar cross-bedding (non deltaic foresets)
Sd	Deformed bedding
Suc	Upward coarsening
Suf	Upward fining
Srg	Graded cross-laminations
SB	Bouma sequence
Scps	Cyclopsams
--- (d)	With dropstones
--- (w)	Wtih dewatering structures
Silts & clays	*Particles of <0.063mm*
Fl	Fine lamination often with minor fine sand and very small ripples
Flv	Fine lamination with rhythmites or varves
Fm	Massive
Frg	Graded and climbing ripple cross-laminations
Fcpl	Cyclopels
Fp	Intraclast or lens
--- (d)	With dropstones
--- (w)	Wtih dewatering structures

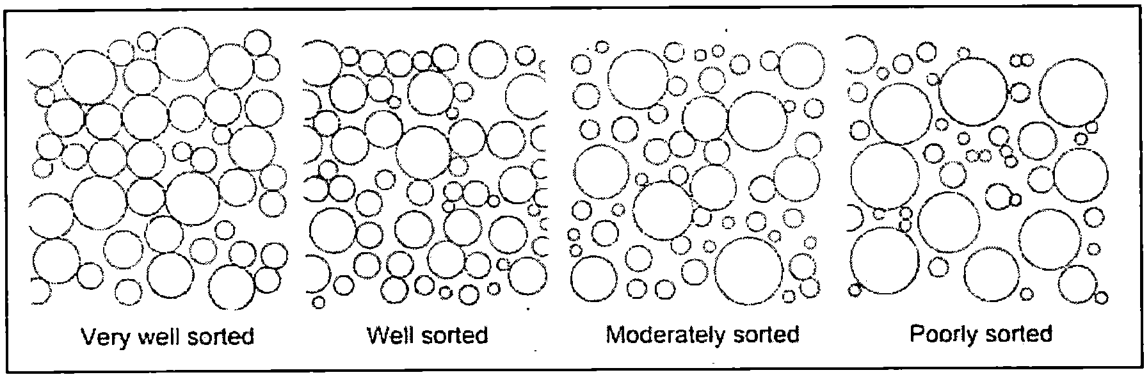
Appendix B: Wentworth Particle Size Chart (after Wentworth, 1922 in Evans and Benn, 2004)



Appendix C: Powers' Roundness Index (after Powers, 1953, in Evans and Benn, 2004)



Appendix D: Sorting Chart (from Evans and Benn, 2004)



Appendix E: Particle Size Results

Roddam Bog Logs 1 and 3 Particle size results:

		RB1.6	RB1.5	RB1.4	RB1.3	RB3.6	RB3.5
	ANALYST AND DATE:						
	SIEVING ERROR:	-99.2%	-99.1%	-99.3%	-99.2%		
	SAMPLE TYPE:	Polymodal, Very Poorly Sorted	Polymodal, Very Poorly Sorted	Unimodal, Moderately Sorted	Polymodal, Poorly Sorted	Polymodal, Moderately Sorted	Bimodal, Moderately Sorted
	TEXTURAL GROUP:	Muddy Sand	Muddy Sand	Sand	Sand	Sand	Sand
	SEDIMENT NAME:	Medium Silty Coarse Sand	Very Coarse Silty Very Coarse Sand	Moderately Sorted Very Coarse Sand	Poorly Sorted Coarse Sand	Moderately Sorted Coarse Sand	Moderately Sorted Medium Sand
METHOD OF MOMENTS Arithmetic (µm)	MEAN	299.3	636.3	1129.5	747.9	810.3	473.6
	SORTING	375.4	574.5	456.9	468.2	459.7	342.4
	SKEWNESS	1.190	0.655	-0.043	0.787	0.521	2.508
	KURTOSIS	3.276	2.221	2.085	2.587	2.356	10.13
METHOD OF MOMENTS Geometric (µm)	MEAN	62.42	269.5	1015.0	606.0	567.9	394.1
	SORTING	9.738	5.954	1.643	1.959	1.946	1.786
	SKEWNESS	-0.341	-1.212	-0.991	-0.227	-0.586	0.380
	KURTOSIS	1.775	3.900	3.576	2.303	2.737	3.547
METHOD OF MOMENTS Logarithmic (ϕ)	MEAN	4.002	1.892	-0.023	0.723	0.582	1.343
	SORTING	3.284	2.574	0.715	0.970	0.960	0.837
	SKEWNESS	0.341	1.212	0.991	0.227	0.586	-0.380
	KURTOSIS	1.775	3.900	3.576	2.303	2.737	3.547
FOLK AND WARD METHOD (µm)	MEAN	69.86	300.3	1048.0	624.1	680.9	389.1
	SORTING	10.01	5.539	1.630	2.005	1.955	1.753
	SKEWNESS	-0.271	-0.479	-0.297	-0.056	-0.176	0.071
	KURTOSIS	0.670	1.030	1.024	0.818	0.876	1.114
FOLK AND WARD METHOD (ϕ)	MEAN	3.839	1.735	-0.068	0.680	0.554	1.362
	SORTING	3.324	2.470	0.705	1.003	0.967	0.810
	SKEWNESS	0.271	0.479	0.297	0.056	0.176	-0.071
	KURTOSIS	0.670	1.030	1.024	0.818	0.876	1.114
FOLK AND WARD METHOD (Description)	MEAN:	Very Fine Sand	Medium Sand	Very Coarse Sand	Coarse Sand	Coarse Sand	Medium Sand
	SORTING:	Very Poorly Sorted	Very Poorly Sorted	Moderately Sorted	Poorly Sorted	Moderately Sorted	Moderately Sorted
	SKEWNESS:	Fine Skewed	Very Fine Skewed	Fine Skewed	Symmetrical	Fine Skewed	Symmetrical
	KURTOSIS:	Very Platykurtic	Mesokurtic	Mesokurtic	Platykurtic	Platykurtic	Leptokurtic

		RB3.4a	RB3.4b	RB3.5a	RB3.5b
	ANALYST AND DATE:				
	SIEVING ERROR:				
	SAMPLE TYPE:	Unimodal, Moderately Well Sorted	Unimodal, Moderately Sorted	Trimodal, Poorly Sorted	Unimodal, Poorly Sorted
	TEXTURAL GROUP:	Sand	Muddy Sand	Sandy Mud	Muddy Sand
	SEDIMENT NAME:	Moderately Well Sorted Fine Sand	Very Coarse Silty Fine Sand	Very Fine Sandy Very Coarse Silt	Very Coarse Silty Fine Sand
METHOD OF MOMENTS Arithmetic (um)	MEAN	143.9	132.8	34.65	118.8
	SORTING	62.20	56.58	35.23	54.01
	SKEWNESS	0.283	0.204	1.832	-0.105
	KURTOSIS	2.880	2.872	6.819	2.376
METHOD OF MOMENTS Geometric (um)	MEAN	125.7	105.0	18.67	83.49
	SORTING	1.881	2.491	3.687	3.298
	SKEWNESS	-3.058	-2.762	-0.767	-2.485
	KURTOSIS	20.83	12.88	3.115	9.853
METHOD OF MOMENTS Logarithmic (ϕ)	MEAN	2.992	3.251	5.743	3.582
	SORTING	0.912	1.317	1.882	1.722
	SKEWNESS	3.058	2.762	0.767	2.485
	KURTOSIS	20.83	12.88	3.115	9.853
FOLK AND WARD METHOD (µm)	MEAN	132.5	118.4	19.66	98.49
	SORTING	1.623	1.958	3.679	2.495
	SKEWNESS	-0.248	-0.397	-0.298	-0.568
	KURTOSIS	1.057	1.473	1.057	1.923
FOLK AND WARD METHOD (ϕ)	MEAN	2.916	3.078	5.669	3.344
	SORTING	0.699	0.969	1.879	1.319
	SKEWNESS	0.248	0.397	0.298	0.568
	KURTOSIS	1.057	1.473	1.057	1.923
FOLK AND WARD METHOD (Description)	MEAN:	Fine Sand	Very Fine Sand	Coarse Silt	Very Fine Sand
	SORTING:	Moderately Well Sorted	Moderately Sorted	Poorly Sorted	Poorly Sorted
	SKEWNESS:	Fine Skewed	Very Fine Skewed	Fine Skewed	Very Fine Skewed
	KURTOSIS:	Mesokurtic	Leptokurtic	Mesokurtic	Very Leptokurtic

Kilham Log 4 Particle Size results

		Kilham 4.9	Kilham 4.8	Kilham 4.7	Kilham 4.6	Kilham 4.5
	ANALYST AND DATE:					
	SIEVING ERROR:	-99.1%	-99.3%	-99.4%	-99.2%	-99.1%
	SAMPLE TYPE:	Trimodal, Very Poorly Sorted	Polymodal, Poorly Sorted	Bimodal, Very Poorly Sorted	Unimodal, Very Poorly Sorted	Unimodal, Poorly Sorted
	TEXTURAL GROUP:	Muddy Sand	Sand	Muddy Sand	Muddy Sand	Sand
	SEDIMENT NAME:	Medium Silty Very Coarse Sand	Poorly Sorted Very Coarse Sand	Medium Silty Very Coarse Sand	Very Coarse Silty Very Coarse Sand	Poorly Sorted Medium Sand
METHOD OF MOMENTS Arithmetic (um)	MEAN	731.4	985.7	845.1	802.8	449.9
	SORTING	625.6	593.7	604.3	630.5	315.9
	SKEWNESS	0.455	-0.024	0.176	0.280	1.299
	KURTOSIS	1.847	1.848	1.806	1.738	5.413
METHOD OF MOMENTS Geometric (um)	MEAN	281.9	586.9	412.5	356.5	309.2
	SORTING	7.622	4.872	5.038	6.271	3.240
	SKEWNESS	-1.387	-2.452	-1.781	-1.435	-2.560
	KURTOSIS	3.999	8.796	5.436	4.234	12.01
METHOD OF MOMENTS Logarithmic (ɸ)	MEAN	1.827	0.769	1.277	1.488	1.693
	SORTING	2.930	2.284	2.594	2.649	1.696
	SKEWNESS	1.387	2.452	1.781	1.435	2.560
	KURTOSIS	3.999	8.796	5.436	4.234	12.01
FOLK AND WARD METHOD (um)	MEAN	311.1	788.8	528.4	365.5	360.2
	SORTING	7.126	3.490	4.675	5.834	2.407
	SKEWNESS	-0.581	-0.553	-0.611	-0.615	-0.277
	KURTOSIS	1.272	2.008	1.566	1.205	1.415
FOLK AND WARD METHOD (ɸ)	MEAN	1.684	0.342	0.920	1.452	1.473
	SORTING	2.833	1.803	2.225	2.544	1.267
	SKEWNESS	0.581	0.553	0.611	0.615	0.277
	KURTOSIS	1.272	2.008	1.566	1.205	1.415
FOLK AND WARD METHOD (Description)	MEAN:	Medium Sand	Coarse Sand	Coarse Sand	Medium Sand	Medium Sand
	SORTING:	Very Poorly Sorted	Poorly Sorted	Very Poorly Sorted	Very Poorly Sorted	Poorly Sorted
	SKEWNESS:	Very Fine Skewed	Very Fine Skewed	Very Fine Skewed	Very Fine Skewed	Fine Skewed
	KURTOSIS:	Leptokurtic	Very Leptokurtic	Very Leptokurtic	Leptokurtic	Leptokurtic

		Kilham 4.4a	Kilham 4.4b	Kilham 4.3	Kilham 4.2
	ANALYST AND DATE:				
	SIEVING ERROR:		-99.4%		-99.3%
	SAMPLE TYPE:	Trimodal, Moderately Sorted	Unimodal, Moderately Sorted	Unimodal, Poorly Sorted	Unimodal, Moderately Well Sorted
	TEXTURAL GROUP:	Sand	Sand	Muddy Sand	Sand
	SEDIMENT NAME:	Moderately Sorted Medium Sand	Moderately Sorted Very Coarse Sand	Very Coarse Silty Very Fine Sand	Moderately Well Sorted Very Coarse Sand
METHOD OF MOMENTS Arithmetic (um)	MEAN	458.7	1120.3	98.57	1262.4
	SORTING	330.3	489.8	110.3	471.9
	SKEWNESS	1.532	0.034	7.243	-0.293
	KURTOSIS	4.901	1.868	77.12	1.932
METHOD OF MOMENTS Geometric (um)	MEAN	364.7	992.6	66.05	1151.6
	SORTING	1.977	1.694	2.990	1.583
	SKEWNESS	-0.050	-0.801	-2.016	-0.954
	KURTOSIS	3.010	3.008	9.656	3.027
METHOD OF MOMENTS Logarithmic (φ)	MEAN	1.455	0.011	3.920	-0.204
	SORTING	0.983	0.760	1.580	0.663
	SKEWNESS	0.050	0.801	2.015	0.954
	KURTOSIS	3.010	3.008	9.656	3.027
FOLK AND WARD METHOD (um)	MEAN	363.0	1019.9	75.15	1174.2
	SORTING	1.982	1.696	2.339	1.588
	SKEWNESS	0.068	-0.272	-0.235	-0.408
	KURTOSIS	1.069	0.881	1.445	0.963
FOLK AND WARD METHOD (φ)	MEAN	1.462	-0.028	3.734	-0.232
	SORTING	0.987	0.762	1.226	0.667
	SKEWNESS	-0.068	0.272	0.235	0.408
	KURTOSIS	1.069	0.881	1.445	0.963
FOLK AND WARD METHOD (Description)	MEAN:	Medium Sand	Very Coarse Sand	Very Fine Sand	Very Coarse Sand
	SORTING:	Moderately Sorted	Moderately Sorted	Poorly Sorted	Moderately Well Sorted
	SKEWNESS:	Symmetrical	Fine Skewed	Fine Skewed	Very Fine Skewed
	KURTOSIS:	Mesokurtic	Platykurtic	Leptokurtic	Mesokurtic