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Glaciodynamics of the central sector of the last British-Irish Ice Sheet in Northern England

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Abstract

The central sector of the last British-Irish Ice Sheet (BIIS) was characterised by considerable complexity, both in terms of its glacial stratigraphy and geomorphological signature. This complexity is reflected by the large number and long history of papers that have attempted to decipher the glaciodynamic history of the region. Despite significant advances in our understanding, reconstructions remain hotly debated and relatively local, thereby hindering attempts to piece together BIIS dynamics. This paper seeks to address these issues by reviewing geomorphological mapping evidence of palimpsest flow signatures and providing an up-to-date stratigraphy of the region. Reconciling geomorphological and sedimentological evidence with relative and absolute dating constraints has allowed us to develop a new six-stage glacial model of ice-flow history and behaviour in the central sector of the last BIIS, with three major phases of glacial advance. This includes: I. Eastwards ice flow through prominent topographic corridors of the north Pennines; II. Cessation of the Stainmore ice flow pathway and northwards migration of the North Irish Sea Basin ice divide; III. Stagnation and retreat of the Tyne Gap Ice Stream; IV. Blackhall Wood-Gosforth Oscillation; V. Deglaciation of the Solway Lowlands; and VI. Scottish Re-advance and subsequent final retreat of ice out of the central sector of the last BIIS. The ice sheet was characterised by considerable dynamism, with flow switches, initiation (and termination) of ice streams, draw-down of ice into marine ice streams, repeated ice-marginal fluctuations and the production of large volumes of meltwater, locally impounded to form ice-dammed glacial lakes. Significantly, we tie this reconstruction to work carried out and models developed for the entire ice sheet. This therefore situates research in the central sector within contemporary understanding of how the last BIIS evolved over time.

Key Words: British-Irish Ice Sheet; ice-sheet reconstruction; Late Devensian; sedimentology; geomorphology; glaciodynamic

1. INTRODUCTION

Recent modelling studies have demonstrated that the last British-Irish Ice Sheet (BIIS) was highly dynamic, drained by a number of oscillating, fast-flowing ice streams (Boulton & Hagdorn, 2006; Hubbard et al., 2009) and associated with rapid switches in ice-flow direction driven by shifting ice-dispersal centres and ice divides (Salt & Evans, 2004; Greenwood & Clark, 2008, 2009a,b; Hughes, 2008; Livingstone et al., 2008; Davies et al., 2009a; Evans, et al., 2009; Finlayson et al., 2010). These reconstructions are in accord with the field evidence, which is characterised by a rich diversity and complexity of Late Pleistocene sediments and landforms (Clark et al., 2004; Evans et al., 2005). The last BIIS extended as far south as the Isles of Scilly (Scourse, 1991; Hiemstra et al., 2006), covered almost all of Ireland (Ó Cofaigh & Evans, 2007; Greenwood & Clark, 2009a,b), extended to the northern and western continental shelf edges (Sejrup et al. 2005, 2009; Bradwell et al., 2007, 2008; O Cofaigh, et al., 2010a) and coalesced with Scandinavian ice in the North Sea Basin (Carr, et al., 2006; Bradwell et al., 2008; Sejrup et al., 2009; Davies et al., 2011). The ice sheet was drained by large, fast-flowing ice streams and outlet glaciers in the Celtic and Irish Sea basins (Evans & Ó Cofaigh, 2003; Ó Cofaigh & Evans, 2007; Roberts et al., 2007; Thomas & Chiverrell, 2007; Bigg et al., 2010), the north-western and north-eastern margins of Scotland (Merritt et al., 1995; Bradwell et al., 2007), and southwards down the east coast of England (Boulton & Hagdorn, 2006; Catt, 2007; Davies et al., 2009a; Boston et al., 2010; Evans & Thompson, 2010). Deglaciation was typified by periodic calving into the Atlantic Ocean (Knutz et al., 2001, 2002; Peck et al., 2007; Hibbert et al. 2010; Scourse et al. 2009), pro-glacial lake development along its southern edge (Evans et al., 2005; Bateman et al., 2008), oscillating margins (Evans & Ó Cofaigh, 2003; Thomas et al., 2004; Thomas & Chiverrell, 2007; McCabe et al., 2007; Livingstone et al. 2010a,b) and numerous regional re-advances, some of which have been correlated with Heinrich Events (McCabe & Clark, 1998; McCabe et al., 2005). Constraints on the temporal evolution of the ice-sheet, gleaned from terrestrial dating techniques (e.g. Bateman et al., 2008, 2011; Ballantyne, 2010; Telfer et al., 2009; McCarroll et al. 2010; Ó Cofaigh et al., 2010b) and offshore ice-rafted debris (IRD) records (e.g. Peck et al., 2007; Hibbert et al., 2009; Scourse et al., 2009; Bigg et al., 2010) have provided a chronological framework from which the asynchronous response of different sectors of the ice sheet and the associated ice streams can be compared and correlated to climatic fluctuations (see Chiverrell & Thomas 2010 for a review).

The central sector of the BIIS, which in this article refers to the northern Irish Sea Basin across to NE England (Fig. 1) and covers the counties of Cumbria, Dumfries-shire and Durham, had an important role to play in modulating a dynamic, non-linear and extensive ice sheet (cf. Boulton & Hagdorn 2006; Evans et al., 2009; Hubbard et al., 2009; Clark et al., In press). This is due to its central location and close proximity to the major upland ice-dispersal centres of the Southern Uplands, Lake District, Cheviots and Pennines (Figs. 1 & 2), promoting a complex geomorphological ice-flow signature reflecting multiple dynamic changes in ice-flow direction, flow reversals and oscillatory margins (e.g. Trotter, 1929; Hollingworth, 1931; Huddart, 1970, 1991, 1994; Letzer, 1978, 1987; Mitchell, 1994, 2007; Clark, R. 2002; Huddart & Glasser, 2002; Livingstone et al., 2008; Davies et al., 2009a; Evans et al., 2009; Stone et al., 2010; Davies et al., In press). It is therefore a key area in terms of reconstructing the palaeo-dynamics of the ice sheet through the last glacial cycle and the linkages between major ice-flow phases, divide shifts, ice-sheet marginal oscillations and sub-Milankovitch scale climate events. Of particular importance, given the region's connection with western and eastern England through the major topographic lows of the Stainmore and Tyne Gaps (Livingstone et al., 2008) (Figs. 1 & 2), is its potential for elucidating the asynchronicity in ice dynamics exhibited between the North Sea and Irish Sea sectors of the last BIIS (Thomas & Chiverrell, 2010; Clark et al. In press). The Solway Lowlands and Dumfries-shire are also located in the former onset zone of the Irish Sea Ice Stream (Merritt & Auton, 2000; Roberts et al. 2007) and therefore could have played an important role in regulating its drainage, whilst the glacial lobe that occupied the Vale of York was sourced from ice in the Lake District and Howgill Fells (e.g. Catt, 2007; Chiverrell & Thomas, 2010). Similarly NE England was affected by competing ice lobes from both NW England and Scotland, with ice flow through the Tyne Gap and as part of the Tweed Ice Stream thus impacting upon the dynamics of the North Sea ice lobe (cf. Davies et al., 2009a, 2011).

Given the growing body of research in this sector of the ice sheet over the last decade and the long-standing history of glaciological investigation in this area (e.g. Goodchild, 1875, 1887; Trenchmann, 1915; Trotter, 1929; Hollingworth, 1931; Huddart, 1970; Letzer, 1978, 1981, 1987), it is now pertinent to review and synthesize the evidence to provide a holistic reconstruction of ice sheet dynamics in the central sector of the BIIS, and contextualise this with regard to the overall pattern of ice-flow. This paper therefore has two aims: (a) to compile and review the stratigraphical and geomorphological evidence for ice-flow in the central sector of the last BIIS; and (b) to produce an ice-sheet reconstruction for the central sector of the BIIS that conforms to glacial theory and best fits the available geological evidence (both at a regional and national scale).

2. TRADITIONAL MODELS OF DYNAMIC ICE-FLOW IN THE CENTRAL SECTOR OF THE LAST BIIS

Over the last decade, the glacial landscape inherited from the last BIIS has been recognised to exhibit a signature of dynamic ice flow (e.g. Salt, 2001; Salt & Evans, 2004; Greenwood & Clark, 2008). The paradigm shift from a static model of ice-dispersal centres (e.g. Flint 1943, 1971; Bowen, 2002) to a dynamic model of rapid flow phases caused by migrating ice divides and dispersal centres was driven by conceptual breakthroughs and glacial geomorphological mapping of palaeo-ice sheets in the northern hemisphere (Dyke et al. 1982; Punkari, 1982, 1993; Dyke & Prest 1987; Boulton & Clark, 1990a,b; Clark, 1997; Clark et al., 2000; Clark & Meehan, 2001; Hughes, 2008; Greenwood & Clark, 2009a,b), reverting to an early conceptual model proposed by Tyrrell (1898). Indeed, it is now recognised that several generations of subglacial lineations or other glacial features can cross-cut each other, resulting in palimpsest flow signatures which can document changing ice flow directions and record a relative chronology of flow phasing (Fig. 3) (e.g. Boulton & Clark, 1990a,b; Clark, 1999; Clark & Meehan, 2001; Livingstone et al., 2008; Hughes, 2008; Greenwood & Clark, 2009a,b).

Early attempts to reconstruct palaeo-ice flows in the central sector of the BIIS proposed “basal ice-sheds” to explain the complex drumlin orientations, erratic pathways (see Fig. 4) and stratigraphy in the Vale of Eden (Harmer 1928; Hollingworth, 1931); a concept that is now incompatible with our modern knowledge of glacier physics (cf. van der Veen 1999; Cuffey & Paterson 2010). Rose and Letzer (1977) were the first to suggest that the region contained overprinted subglacial bedforms that recorded migrating ice-dispersal centres, a concept that was revisited by Mitchell (1991a, b, 1994) and Mitchell and Riley (2006) and used to prove ice-flow reversal in the Vale of Eden due to ice-divide migration associated with a Dales Ice Centre over the uplands of the western Yorkshire Dales extending across the Howgill Fells to the Lake District (Mitchell, 1994; Mitchell & Riley, 2006). The mobility of the local northern Pennine ice divide has similarly been reconstructed by Mitchell (2007) based upon superimposed drumlins in the Cow Green area, whilst Livingstone et al., (2008) provided a relative chronology of ice-flow phases from cross-cutting landforms throughout the Vale of Eden, Solway Lowlands and through the Tyne and Stainmore Gaps (Fig. 3b). To the north of the Solway, Salt and Evans (2004) provided evidence of four early, topographically unconstrained ice flow stages and three later topographically constrained stages for SW Scotland. Similarly Roberts et al. (2007) identified two distinct phases of ice flow on the Isle of Man, with the first phase sourced from Scottish ice-dispersal centres and the second phase from the Solway Lowlands. In NE England, several phases of ice flow are recognised in the stratigraphic record. They reflect the oscillation of the Tyne Gap Ice Stream in response to shifting ice divides and drawdown in the Irish Sea Basin, before the increasing dominance of the Scottish-sourced North Sea Lobe during the Dimlington Stadial (Davies et al., In press).

The landform evidence in the lowlands to the west of the Pennines has been used traditionally to identify three distinct phases (Fig. 3a) during the last glaciation, including: an early 'Scottish Advance', followed by a 'Main Glaciation' and then a 'Scottish Re-advance' (cf. Trotter & Hollingworth, 1932b). An additional 'Gosforth Oscillation', between the 'Main Glaciation' and the 'Scottish Readvance', was proposed by Trotter (1937) and has gained recent support from Merritt & Auton (2000) and Livingstone et al. (2010a).

The early 'Scottish Advance' is delimited by erratic trains, sourced in the Southern Uplands and traced up the Vale of Eden and across the Stainmore Gap, and in the lower Derwent valley (Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932b). Erratic trains also indicate the flow of Lake District ice over the Stainmore Gap during this phase.

Erratic trains (Fig. 4) and drumlins in the Vale of Eden were used to constrain the 'Main Glaciation' (Harmer, 1928; Trotter, 1929), which was characterised by northerly flowing Lake District ice coalescing with Scottish ice in the Solway Lowlands, before streaming eastwards through the Tyne Gap (Fig. 3) (Harmer, 1928; Trotter, 1929). Westerly flowing ice was also recorded in this phase by drumlins trending around the northern margin of the Lake District into the Irish Sea Basin (Fig. 3) (Trotter, 1929; Hollingworth, 1931).

The Scottish Readvance is recorded by a thin till, eskers and deltaic deposits, marking a temporary re-advance of Scottish ice into the Solway Lowlands during deglaciation (Trotter, 1922, 1923, 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1971a, b, 1991, 1994; Huddart & Tooley, 1972; Huddart et al., 1977, Huddart & Clark, 1994; Livingstone et al., 2010). The Scottish Readvance also impacted on the Cumbrian coast as least as far south as Annaside and Gutterby, where Scottish Readvance ice laid down tills that overlie glacigenic deposits attributed to the Gosforth Oscillation (Trotter, 1937; Merritt & Auton 2000; see Huddart & Glasser 2002 for review).

Livingstone and co-authors (2008) expounded upon the three-stage reconstruction of dynamic ice flow by producing a regional map of the glacial bedforms and flow sets. Seven major phases of ice flow were recorded in the region, based on cross-cutting relationships (Figs. 3b & 5) (Livingstone et al., 2008, 2010c). Flow phases LT1-3, ST1-2 and ES1 are associated with the Main Glaciation (Stage 1: Table 1) when ice flowed eastwards, across the Pennines, through the Tyne and Stainmore Gaps (Figs. 3b & 5) (Livingstone et al., 2008, 2010c). The subsequent migration of ice-divides back towards major upland massifs meant the major ice drainage arteries of the Tyne and Stainmore Gaps diminished (Stage II: Table 1), resulting in flow phases LT4 and ST3-4 down the North Tyne and out of Teesdale respectively (Figs. 3b, 5c, d). A major flow switch occurred when ice drained westwards into the Irish Sea Basin (LT5) (Figs. 3b & 5a, b) as a fast-flowing tributary of the Irish Sea Ice Stream (Stage IV: Table 1) (Livingstone et al., 2008, 2010a). This stage (IV) is also correlated with flow phase EC1 (Figs. 3b & 5a), which records the inland advance and southerly expansion of the North Sea Lobe. The youngest flow phases are thought to record the final re-advance of ice into the Solway Lowlands (LT6 and SF1) (Fig. 3b) and the subsequent retreat back into upland dispersal centres (Stage VI: Table 1) (Livingstone et al. 2008, 2010b).

The till sequence in NE England has been traditionally attributed to two-phases, with initial ice-flow from NW Britain moving through the Tyne Gap, followed by a southerly ice flow from the Cheviots/Tweed region (Lunn, 1995; Teasdale & Hughes, 1999; Davies et al. 2009a). The later southerly flow is interpreted as surging in response to the activation of different ice dispersal centres (Eyles et al., 1982, 1994; Catt, 1991a,b; Douglas, 1991; Evans et al., 1995; Teasdale & Hughes, 1999; Boulton & Hagdorn, 2006; Davies et al. 2009a; Boston et al., 2010). This North Sea Lobe may have been deflected by Fennoscandian ice (Boulton et al., 1991; Carr et al., 2004; Davies et al., 2011) or a relict ice dome in the central North Sea (Clark et al., In press).

3. STRATIGRAPHIC FRAMEWORK

The lithostratigraphy used here mainly follows a new, formally ratified, 'top down' framework covering all onshore Quaternary deposits in Great Britain (McMillan et al., 2011). All glacigenic deposits (glacial, glaciofluvial, glaciolacustrine, glaciomarine) younger than the Ipswichian (MIS 5e) are assigned in this framework to the Caledonia Glacigenic Group, which embraces subgroups containing formations and members with common lithological characteristics and properties. Older deposits are assigned in a similar manner to the Albion Glacigenic Group. Descriptor epithets have been formally reintroduced to aid the user. The subgroups in northern England (Fig. 6) broadly distinguish tills laid down by ice emanating from local ice dispersion centres from tills deposited by coastal ice streams and their onset zones. The major till units of the central sector of the BIIS are summarised in Table 2. All units mentioned here have been defined in the BGS Lexicon of Named Rock Units and the Index of Computer Codes (http://www.bgs.ac.uk/lexicon/).

The Quaternary stratigraphic succession of northern England falls largely within the glacial limits of the Main Late Devensian (MLD) Glaciation, with glacial erosion having been effective in removing most evidence of older glaciations (Stone et al. 2010). This is supported by the regional chronostratigraphy, with well-dated events such as the Dimlington Stadial (Rose, 1985; Bateman et al., 2008) and the extension of the Irish Sea Ice Stream to its maximum extent in the Celtic Sea (e.g. Ó Cofaigh & Evans, 2007) being used to provide constraints on key flow phases (Table 1). This is exemplified by the flow of ice through the Tyne Gap (see Livingstone et al. 2010a), which has been correlated with the Blackhall Till at Whitburn Bay and assigned to the Dimlington Stadial (Davies et al., 2009a). Similarly, ice flow through the Stainmore Gap formed the Escrick moraine in the Vale of York (Table 1), which has been dated to the Late Devensian, and formed when the North Sea Lobe blocked the Humber Estuary, resulting in the formation of Glacial-Lake Humber (Catt, 2007; Bateman et al., 2008).

3.1. Pre-Devensian Glaciations

Fragmentary evidence of pre-Devensian deposits in central and western Cumbria that survived the last glaciation provide additional constraints on the glacigenic chronostratigraphy (Stone et al., 2010; McMillan et al., 2011) (Table 1). At Thornsgill (National Grid Reference (NGR) NY 381 242) and Mosedale (NY 356 239), a sequence comprising two till units separated by a prominent palaeosol and peat bed provides evidence of multiple glacial cycles, with the lowermost Thornsgill Till thought to correlate with a pre-Devensian (MIS 6 or older) advance (Table 1), and the peat layer with either the last (Ipswichian) interglacial (Huddart & Glasser, 2002); or with an Early Devensian interstadial

(Boardman, 2002). The basal unit from a borehole at Wigton (NY 2532 4866) (Eastwood et al. 1968, p227), which comprised red clay containing ostracods, foraminifera and Turritella communis (Wigton Marine Bed: Table 1) is thought to be Ipswichian in age. A pre-Devensian till has also been observed in a glaciotectonically disturbed section at Scandal Beck (NY 742 024) below organic deposits that have been ascribed to the Ipswichian Interglacial on the basis of pollen analysis (Carter et al., 1978; Letzer, 1978; Mitchell, 2002). Merritt and Auton (2000) conclude that the products of at least one major pre-Devensian glaciation in Western Cumbria are represented in cored boreholes around Drigg and Lower Wasdale (Drigg Till) (Fig. 7a). Ice possibly flowed out of Wasdale during MIS 4 to lay down the Maudsyke Till. Varved glaciolacustrine deposits (Carleton Silt) were subsequently deposited in proglacial lakes when the glacier retreated, possibly on two occasions. The Carleton Silt passes upwards into cold-water marine deposits (Glannoventia Formation) recording a marine transgression and contemporary sea level of at least 20 m below a.s.l., probably during MIS 3.

In County Durham, a buried valley at Warren House Gill contains three diamictons interbedded with silts, sands and gravels. The basal diamicton, the Ash Gill Member of the Warren House Till Formation, is a glaciomarine to subglacial glaciotectonite and traction till with erratics from the North Sea Basin, Scotland and, rarely, Scandinavia (Davies et al., 2010b). This diamicton is older than the MIS-7 Easington Raised Beach Formation (Davies et al., 2009b), which contains erratics interpreted as being derived from the Ash Gill Member.

3.2. Devensian multiple till and stratified sediment sequences

3.2.1. Background

The complex sequences of glacigenic deposits that characterise the central sector of the BIIS have been subject to ongoing debate since the 19th Century, featuring in the development of conceptual models outlining how glacial sediments can be deposited (cf. Huddart & Glasser, 2002). Tripartite successions comprising clay, silt and sand sandwiched between till units are common throughout northern England and have traditionally been assigned to the waxing and waning of glaciers over multiple glaciations (cf. Huddart & Glasser, 2002). However, Goodchild (1875, 1887) fitted the deposits of Cumbria into a model of englacial and subglacial meltout during in situ downwasting of stagnant ice during a single glaciation. In a seminal paper, Carruthers (1953) used the laminated clay and silt deposits of the Vale of Eden and NE England to propose glacial "undermelt" as an important process. This idea, that multiple and complex sequences can be deposited during one glaciation, was further developed in the 1960s and 70s (e.g. Boulton, 1972, 1977) and applied to some of the Cumbrian deposits by the work of Huddart (1970). Research undertaken by the British Geological Survey in the 1920s and 1930s had taken a different view, favouring a model of ice-frontal advance and retreat associated with proglacial deposition (e.g. Trotter, 1929; Hollingworth, 1931).

The tripartite sequence that characterises much of the field area is further complicated by the identification of an additional thin till cap in the Solway Lowlands, correlated to the Scottish Re-advance (cf. Livingstone et al. 2010b for a review). This interpretation of the capping till has been challenged however, with some authors dismissing it as largely illusory (Evans & Arthurton, 1973; Pennington, 1978). Eyles and McCabe (1989, 1991) interpreted the till as a glaciomarine mud drape in line with their 'glaciomarine model' for the Irish Sea Basin. The implication of this model is that rising sea level in an isostatically depressed marine basin caused the rapid retreat of the Irish Sea Ice Stream and resulted in marine limits up to 140 m a.s.l (see Huddart and Glasser 2002 for discussion).

The following section provides a comprehensive account of the regional glacial stratigraphy of the last BIIS in northern England and Dumfries-shire (see Tables 1 & 2 and Figs. 4-7). By compiling this information we are able to highlight current controversies and relate our stratigraphical framework to the glacial geomorphology.

3.2.2. Central and Western Cumbria

Early in the twentieth century a 'tripartite' stratigraphy was recognised in central west Cumbria (Trotter 1929; Hollingworth 1931; Trotter & Hollingworth 1932b). This comprised a 'Lower Boulder Clay', 'Middle Sands and Gravels' and an 'Upper Boulder Clay', although additional units were known to occur locally and the two tills could not be distinguished where the intervening sand and gravel was absent. The lower till was linked to the Main (Late Devensian) Glaciation whereas tills higher in the sequence were linked to readvances of Scottish ice, either the Gosforth Oscillation or the succeeding Scottish Readvance (Trotter et al., 1937). Subsequent work along the coast by Huddart (1971) broadly reconfirmed the tripartite stratigraphic model, but dismissed evidence for the Gosforth Oscillation proposed by Trotter et al., (1937). The lower (Selker) till was interpreted to be the basal till deposited by the Main Late Devensian ice sheet, the middle (Annaside) sands and gravels were retreat stage sandur sediments, and the upper ‘Gutterby Spa Complex’, which included tills, was the product of the Scottish Readvance (Huddart and Tooley, 1972; Huddart et al., 1977).

Following extensive investigations in western Cumbria undertaken in the 1990s on behalf of UK Nirex Ltd (Bowden et al., 1998), a more comprehensive lithostratigraphical framework was set up (Nirex, 1997; Merritt and Auton, 2000) (Fig. 7a). This scheme, which embraces some units named by Huddart and Tooley (1972) and formalised by Thomas (1999), has been updated by McMillan et al., (2011); it includes three subgroups, 13 formations and 46 members. The Central Cumbria Glacigenic Subgroup includes erratics derived predominantly from the Lake District and Shap Fell, whereas the West Cumbria Glacigenic Subgroup includes material from southern Scotland, the Solway Lowlands, the west Cumbrian Coalfield and the northern Cumbrian mountains (Fig. 7a). The deposits of the latter subgroup were laid down by ice flowing through the Solway Lowlands, swinging around the north-western side of the Lake District and feeding into the Irish Sea Ice Stream (Flow-set LT5 of Figs. 3b & 5a-c). Formations and members within the two subgroups interdigitate locally, notably in lower Wasdale (Fig. 7a), where they reveal former interactions between locally-sourced and far-travelled ice. An increase in the proportions of Scottish erratics in tills occurs toward the top of several multi-till sequences in the district, indicating that Scottish ice gradually became dominant over ice emanating from the mountains to the east (Eastwood et al., 1931; Trotter et al., 1937; Nirex 1997).

Glaciers advanced from the western valleys of the Lake District, particularly Wasdale, to lay down the very stony Holmrook Till (Fig. 7a). This occurred during the build-up of the Late Devensian ice sheet, which probably reached its maximum extent early in MIS 2, when the whole region was glaciated. Scottish ice within the Irish Sea ice stream eventually deflected Lake District ice southwards and flowed across Lower Wasdale, laying down the Ravenglass Till (Table 2) (Merritt and Auton, 2000).

The Irish Sea Ice Stream became dominant during several readvances, as recorded by thinly interbedded tills and sands and gravels within the Gosforth Glacigenic Formation (Fig. 7a, Table 2). The first major readvance (Gosforth Oscillation) followed significant deglaciation. As the Irish Sea Ice Stream thickened and encroached inland, thick sequences of fine-grained, laminated glaciolacustrine sediment accumulated within Lower Wasdale (Whinneyhill

Coppice Clay) and other major valleys of the district (e.g. Ehen Valley Silt) (Fig. 7a). During the Gosforth Oscillation, ice over-rode most of the coastal plain up to a height of about 100 m a.s.l., a scenario previously proposed by Trotter et al., (1937). Local ice expanded across Lower Wasdale to lay down the Green Croft Till (Table 2). On its retreat an ice marginal lake reformed, within which the glaciolacustrine Holmeside Clay and deltaic Mainsgate Wood Sand and Gravel were deposited (Fig. 7a). The lake was impounded by the Irish Sea Ice Stream, the margin of which oscillated across the estuary of the River Irt several times, laying down the Drigg Beach, Fishgarth Wood and other tills (Fig. 7a & Table 2). These minor oscillations were probably contemporaneous with the creation of the well-known glacitectonically stacked moraine at St. Bees (Williams et al., 2001), 30 km to the north-west, widely attributed to the Scottish Readvance (Huddart and Glasser, 2002, and references therein).

3.2.3. Solway Lowlands and Vale of Eden

The oldest glacigenic deposits associated with the Main Late Devensian Glaciation in the Vale of Eden and Solway Lowlands are the Gillcambon and Chapelknowe Till Formations (cf. Stone et al., 2010; McMillan et al., 2011; Tables 1 & 2; Figs. 5a,b, 7b). These till units contain an extensive suite of Scottish Southern Upland erratics including Criffel and Dalbeattie granite and greywacke (Table 2). The Gillcambon Till also contains erratics from the Lake District and Vale of Eden and it is therefore difficult to differentiate its erratic and geochemical provenance from other glacigenic sediments (e.g. Greystoke Till Formation) in the region (e.g. Dixon et al., 1926; Livingstone et al., 2010a). The Gillcambon Till has been assigned to the ‘Early Advance' of Scottish ice up the Vale of Eden across the Stainmore Gap and into Eastern England (Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932b; Raistrick, 1934; Huddart, 1970; Catt, 1991, 2007). However, at the entrance to the Vale of Eden, it has been reconciled with ice-flow convergence on the Solway Lowlands from the Southern Uplands and the Lake District, before streaming eastwards through the Tyne Gap (Livingstone et al., 2010c) (Fig. 5a,c). The easterly flow of ice through the Tyne Gap during deposition of the Gillcambon Formation is equivocal given the identification of this till at Willowford and the corresponding till fabric orientations (NE to NNE) (Trotter, 1929; Huddart, 1970; Livingstone et al. 2010c).

Overlying the Gillcambon Till in the Vale of Eden and Solway Lowlands are a discontinuous series of sands, gravels and laminated clays and silts termed the 'Middle Sands' (cf. Trotter and Hollingworth, 1932b) (Fig. 7b). Goodchild (1875, 1887) and Huddart (1970) attributed the deposits to the subglacial melting of a single, stagnant ice mass, while Trotter (1929) and Hollingworth (1931) surmised that they were formed proglacially, thus delimiting a period of partial deglaciation, followed by re-advance. The interpretation of Hollingworth (1931) has recently been reinforced by the identification of a sequence of clastic varves at Blackhall Wood, which were deposited in a pro-glacial lake formed in the Solway Basin (Livingstone et al., 2010b). These varved sediments and the associated debris flow and outwash sand, gravel and diamicton deposits exposed in the Caldew Valley and identified in borehole records between Carlisle and Penrith (see Livingstone et al. 2010a), delimit a period of retreat and ice-free conditions; the associated lithostratigraphic unit has been named the Blackhall Wood Glaciolacustrine Formation (Fig. 7b; Table 1).

An upper till in the Vale of Eden and Solway Lowlands (Edenside Till Member of Greystoke Till Formation) is assigned to the ‘Main Glaciation’ (MLD) (Hollingworth 1931; McMillan et al., 2011) (Table 2). This red-brown till is generally between 5-20 m thick and forms the drumlinoid landforms that cover the Vale of Eden and Solway Lowlands (Figs. 7b, 5a,b;

Table 1) (Livingstone et al., 2010b). It is characterised by a mixed provenance of Scottish, Lake District and local erratics (Trotter, 1929; Hollingworth, 1931; Huddart, 1970) and due to its position above the Blackhall Wood Glaciolacustrine Formation is now believed to have been deposited during the Blackhall Wood/Gosforth Re-advance (Livingstone et al. 2010b; Table 1).

In the Solway Lowlands and SE Dumfries-shire, the Chapelknowe and Greystoke Till formations are overlain locally by gravels, sands and clays (Plumpe Farm Sand and Gravel & Great Easby Clay formations) and then capped by a thin (<5 m) upper red till (Gretna Till Formation) (Livingstone et al., 2010b; McMillan et al., in press; Fig. 7b; Tables 1 & 2). The Great Easby Clay and Plumpe Farm Sand and Gravel Formation relate to glaciolacustrine and glaciofluvial deposition respectively, including that associated with Glacial-Lake Carlisle (Dixon et al., 1926; Huddart, 1970), during the partial deglaciation of the region, probably following the Blackhall Wood/Gosforth Oscillation. The Great Easby Clay Formation consists of dark reddish brown clays, silts and very fine-grained sands that are generally thinly laminated and locally varved. Both formations show evidence for being partially deformed. These sediments have been traced as far west as Annan, north to Langholm (McMillan et al., in press), east as far as Great Easby (with turbidite structures, dropstones and thick proximal varves, Huddart, 1970) and just to the south of Carlisle (Livingstone et al., 2010b). Also overlying the Greystoke Till is the Brampton sand and gravel complex (Baronwood Sand and Gravel Formation), deposited in the lee of Pennines (cf. Livingstone et al. 2010d; Table 1; Fig. 7). This landform-sediment assemblage, which is 44 km² and up to 25 m thick, comprises a series of kettle holes, discontinuous ridges (eskers) and flat-topped hills (ice-walled lake plains) (Huddart, 1970, 1983; Livingstone et al., 2010d).

The Gretna Till (Fig. 7b, Table 2) has a western Southern Upland provenance and pinches out east and south of Carlisle, and has therefore been associated with a late-stage re-advance of Scottish ice into the Solway Lowlands (Fig. 5a) (e.g. Dixon et al., 1926; Trotter & Hollingworth 1932; Huddart, 1970, 1994). Phillips et al., (2007) suggest that the thin till sheet is a function of a short-lived pulse moving rapidly across water-saturated sediment and into proglacial lakes dammed up against higher ground to the east of the ice. The short-lived nature of the re-advance coupled with the buffering effects provided by water-saturated sediment at the ice-bed interface resulted in the re-advance exerting negligible depositional, erosional or deformational influence (Livingstone et al., 2010b). Also associated with the Scottish Re-advance, and overlying drumlins comprising the Greystoke Till, is the 9 km² Holme St. Cuthbert sand and gravel complex (Kilblane Sand and Gravel Formation: Figs. 3 & 5a; Table 1) (Huddart, 1970); interpreted as an ice-contact delta (Huddart, 1970, 1994; Huddart & Tooley, 1972; Huddart et al., 1977; Huddart & Glasser, 2002; Livingstone et al., 2010b). Palaeocurrents are from the west and north-west, with an erratic assemblage from the Southern Scottish coast indicated by high percentages of Criffel granite and a distinctive Lower Calciferous Sandstone conglomerate.

3.2.4. Dumfries-shire-Langholm area

It is apparent both from the composition of the Gretna and Chapelknowe tills and the predominant orientation of drumlins that the lowlands north of the inner Solway Firth were crossed by ice flowing from the west (Figure 3b, flow-sets LT3). This ice was mainly sourced in western and central parts of the Southern Uplands, specifically in the Galloway Hills (including Criffel) and the Moffat Hills, and it flowed into the Solway Lowlands via the valleys of the Nith and Annan respectively (Figs. 5a & 7). Relatively little ice flowed into the Solway Lowlands from the Langholm Hills, north of Gretna, apart from when flow-sets LT5 and LT7 were created following the LGM (Fig. 3b). This scenario is supported by the distribution of red, sandstone-rich (Gretna) and yellowish-brown, greywacke-rich (Langholm) tills in the district (BGS 2005, 2006) (Fig. 7c). Furthermore, the widespread presence of red, granite-bearing till passing upwards into rubbly, greywacke-rich till north and west of Langholm indicates that ice overwhelmed southern parts of the Langholm Hills from the west during an early phase of the last glaciation (Lumsden et al., 1967; McMillan et al., in press), probably during the LGM. NEXTMap images of the Langholm Hills reveal that this region of the Southern Uplands has been relatively little modified by glacial erosion (Merritt & Phillips 2010, Fig. 46), probably because it was a minor ice dispersal centre and overlain by mainly sluggish, cold-based ice. A minor, southward re-advance of ice from the Langholm Hills occurred following the separation of ice masses during deglaciation (Lumsden et al., 1967), laying down the Mouldy Hills Formation (BGS, 2006).

Although the Gretna and Chapelknowe till formations have not been mapped out to the north-east of Langholm, it is clear that red, sandstone and granite-bearing tills occupy much of Liddesdale (Day, 1970) and a suite of far-travelled glacial erratics sourced in the Galloway Hills have been recorded high on the catchment divide north of Kielder (Clough, 1889). This suggests that some transfluence of ice may have occurred into the catchment of the Tweed at the LGM (see Fig. 9: dotted arrows). This is supported by the results of satellite image interpretation undertaken by BGS (McMillan et al., 2011, in press), which identifies elongate, glacially streamlined landforms arcing north-eastwards towards the divide, but is at odds with the conclusions of Livingstone et al. (2008), who identified no flow sets consistent with such flow (see Fig. 5a). The features identified by the BGS are interpreted instead as having formed by ice flowing in the opposite direction, towards the Irish Sea, after the LGM (Figure 3b, flow-set LT5). The blunt ends of some drumlins in Liddelsdale do indeed indicate south-westward flow (Day, 1970), but it is likely that this flow set has a more complicated history and results from more than one flow phase of the last glaciation.

3.2.5. Tyne Gap

Only one major till unit has been recorded in the Tyne Gap (Wear Till Formation) (Table 2), which ranges in thickness up to 90 m in concealed channels, principally on the east coast (Lunn, 2004; Hughes et al., 1998 (Figure 7d). Lineations representing both erosional and depositional features relate to ice flow through the Tyne Gap as a topographically-controlled ice stream during the LGM (Livingstone et al., 2010c). The Wear Till is characterised by mixed provenance with both Lake District (e.g. Borrowdale Volcanic Group, Carrock Fell gabbro, Penrith sandstone and Threlkeld grey quartz porphyry) and Southern Upland (e.g. Dalbeattie and Criffel granite, greywacke and Silurian grits) erratics (Fig. 4), and probably correlates with the Gillcambon Till (Tables 1 & 2). This is supported by Trotter (1929), who correlated the till at Willowford with the lower-most till of the Vale of Eden. The boundary separating two distinct erratic trains associated with Lake District and Southern Upland provenances is indistinct, with a general southerly increase in Lake District erratics towards the north Pennines (Fig. 4). This diffuse boundary is indicative of competing ice flows, with both Scottish and Lake District ice-dispersal centres becoming dominant at different times (Fig. 5c) (Lunn, 2004). From west to east the till in the Tyne gap changes from a red colour, inherited from the Permo-Triassic sandstones of the Solway Lowlands and Vale of Eden, to a grey colour, reflecting the Carboniferous rocks of the Tyne. Till in the lower South Tyne Valley is typically a hard, stiff, grey-brown clay. Clasts are striated, and comprise sandstone, limestone, greywacke, granite and dolerite, which accords with the eastwards decrease of Lake District erratics noted by Trotter (1929), and which in turn has been associated with a later phase of south-easterly, Scottish-sourced ice down the North Tyne Valley (Dwerryhouse, 1902; Livingstone et al., 2010c).

The Wear Till is overlain by a sequence of silts, clayey sands and gravels, both intercalated with, and locally capped by diamicton (Lovell, 1981; Allen and Rose, 1986). This sequence (Ebchester Sand and Gravel Formation) appears chaotically deposited and is now thought to have formed as the ice retreated westward into the Tyne Gap and stagnated, allowing glaciofluvial outwash and glaciolacustrine deposits (due to localised ponding) to infill the valley as meltwaters drained from the western margin of the ice (Yorke et al., 2007). Diamicton is discontinuously exposed above the sands and gravels in the South Tyne Valley, representing either a re-advance subglacial till (Huddart and Glasser, 2002), or a series of debris flows or re-worked deposits (Yorke et al., 2007). Kamiform deposits are particularly extensive along the flanks of the present course of the Tyne (Yorke et al., 2007) and there is evidence for a major lake (Glacial Lake Wear) which extended up the lower Tyne Valley (Raistrick, 1931; Smith, 1994; Teasdale and Hughes, 1999).

3.2.6. Pennines and Stainmore Gap:

In contrast to the detailed information now known from the complex lithostratigraphic sequences exposed in the lowlands of northwest and northeast England, there are few recent investigations of the superficial deposits of the intervening Pennine uplands (Mitchell, 1991a,b; 2007). This has restricted the development of a formal lithostratigraphy (cf. Stone et al., 2010) for the uplands that would enhance correlation across northern England. The Pennines do however feature prominently in early studies of ‘drift’ and it was such investigations on Pennine tills that formed the basis for understanding the mechanics of till depositional processes within the paradigm of land-based ice sheets (Dakyns, 1872; Tiddeman, 1872; Goodchild, 1875,1887).

These early workers and the subsequent mapping by the Geological Survey (Aveline and Hughes, 1888; Dakyns, et al., 1890; 1891) confirmed that the superficial deposit sequence is dominated by widespread, compact diamicton (Yorkshire Dales Till Formation of McMillan et al., 2011) (Table 2). This till is notable for its lack of facies variability and the dominance of local Carboniferous rock types (Fig. 4) characteristic of the Pennines, thereby indicating that the western Pennine uplands had been covered by local ice centred over Baugh Fell (Dakyns, et al., 1891; Raistrick, 1926), with a separate centre over Cross Fell in the northern Pennines (Dwerryhouse, 1902). Only in the lower ground associated with the Stainmore Gap is there another compact diamicton (Stainmore Forest Till Formation of McMillan et al., 2011) dominated by local Carboniferous lithologies but with notable Shap granite and Lake District erratics which can be traced eastwards into the lowlands of eastern England (Figs. 4, 5d & Table 2) (Madgett & Catt, 1978; Burgess and Holliday, 1979; Mills and Hull, 1976).

3.2.7. North-east coast of England:

Along the east coast of northern England, two subglacial tills have been recognised within the North Sea Coast Glacigenic Subgroup (Figure 7d). The lower, greyish-blue Blackhall Till is characterised by abundant Carboniferous rocks from the Tyne Gap and a component of Southern Upland erratics (Table 2) (Francis, 1970; Davies et al., 2009a). The Blackhall Till was deposited by ice flowing eastwards through the Tyne Gap (Fig. 5c) (Raistrick, 1931; Catt, 1991; Davies et al., 2009a; Table 1) and is correlated with the Wear Till of the Tyne Gap (Figure 7d & Table 2). It was probably laid down when ice extended southwards down the Vale of York (Clark et al., In press), to the shelf-edge in northern Britain (Bradwell et al 2008), and was confluent with the Fennoscandian ice sheet in the North Sea Basin (Carr et al., 2006; Sejrup et al., 2009; Davies et al., 2009a, 2011, in press). Streaming occurred in the suture zone between the two ice sheets in the North Sea Basin (Graham et al. 2007). The topographically controlled Tyne Gap Ice Stream potentially retreated in response to drawdown in the Irish Sea, resulting in an ice-free enclave in eastern Co. Durham. At Warren House Gill, the Blackhall Till contains evidence of deposition of sands and gravels in cavities at the ice-bed interface, as well as periods of ice-marginal oscillation (Davies et al., in press).

Overlying the Blackhall Till is the Peterlee Sand & Gravel Formation (Francis, 1970), a series of sands and gravels deposited as a sandur in front of the expanding ‘North Sea Lobe’ (Davies et al., in press) (Table 1). Inland, the Wear Till, which is laterally equivalent to the Blackhall Till (Figure 7d), is widely overlain by glaciolacustrine silts and clays of the Tyne and Wear Glaciolacustrine Formation, which were deposited in Glacial Lake Wear, when it was dammed by the North Sea Lobe (Raistrick, 1931; Smith, 1994, Stone et al., 2010). These deposits are believed to have been deposited between the southwards-flowing North Sea Lobe off the east coast (Teasdale & Hughes, 1999) and westerly retreat of ice from the Tyne Gap (Livingstone et al., 2010c).

The North Sea Lobe was formed from the coalescence of several ice streams emanating from the Forth, Tweed and Grampian Highlands. It periodically surged in response to shifting ice divides, resulting in the dominance of different ice-accumulation centres and producing a thrust and stacked series of tills at Skipsea in Yorkshire (Boston et al., 2010; Evans & Thomson, 2010). During the Dimlington Stadial, the North Sea Lobe became more dominant on the Durham coast, pushing inland (cf. Catt, 1991) and depositing the upper, red Horden Till (Table 2) (Smythe, 1912; Francis, 1970; Davies et al., 2009a). A boulder pavement, comprising well-orientated, striated and faceted limestone and sandstone boulders, occurs locally at the contact between the Blackhall and Horden tills, for example, at Whitburn Bay. The pavement, together with associated sands and gravels, is interpreted as a depositional lag created at the ice-bed interface of the North Sea Lobe as it overrode earlier glacigenic sediments. Meltwater winnowing removed finer grained material and locally deposited the coarser fractions of the reworked glacigenic sediments, whereas ploughing and lodgement processes orientated and striated the boulders (Davies et al., 2009a).

Based on the provenance, the Horden Till is correlative with the Skipsea Member of the Holderness Formation in Yorkshire and the Bolders Bank Formation offshore (Table 2) (Francis, 1970; Madgett & Catt, 1978; Carr et al., 2006; Davies et al., 2009a; Davies et al., 2011). It contains erratics derived from Northumberland, the Cheviots, the Southern Uplands, and the Grampian Highlands. Further south, the North Sea Lobe impinged on the coast of north Norfolk (Straw, 1960; Pawley et al. 2006) and dammed a series of glacial lakes (Evans et al., 2005; Clark et al., 2004), the largest being Glacial-Lake Humber, which was in existence from at least 22 ka BP until after Heinrich Event 1, at ~16 ka BP (Bateman et al., 2008, 2011; Murton et al., 2009).

The eastern limits of the North Sea Lobe are usually taken to be the limits of the Bolders Bank Formation. This gives the lobe the characteristic shape defined by Boulton et al (2002) during the Last Glacial Maximum. The sediments onshore in eastern England record local ice flow to the southwest (Davies et al., 2009a), but in the offshore region near The Wash they appear to record a radial flow towards the east (Cameron et al. 1992; Carr et al. 2006). The North Sea Lobe was probably constrained in the North Sea Basin by the Fennoscandian Ice Sheet, with which it was confluent at the LGM (29-22 ka BP). This provides a mechanism to force the flow of the lobe southwards over the deformable marine sediments of the North Sea bed. Dates on Glacial Lake Humber indicate that the North Sea Lobe was in existence until at least ~16 ka BP. However, there is evidence of ice-free conditions in the northern and central

North Sea by 25 ka BP (Sejrup et al., 2005), indicating that this ice stream remained in existence after the removal of confining stresses. An alternative explanation, proposed by Clark and co-authors (In press), is that, following maximum glaciation of the North Sea Basin during the Last Glacial Maximum, a stagnant dome of ice remained in the southern-central North Sea Basin, thus deflecting the North Sea Lobe.

4. RECONSTRUCTION OF THE CENTRAL SECTOR OF THE LAST BIIS

Given the complex array of glacial landforms and deposits in Northern England and the relevance this has for resolving the glaciodynamics of the last BIIS, it is important that we elucidate its glacial history. The following section provides an updated glacial reconstruction by combining glacial geomorphological mapping (Section 2; Figs. 3b & 5) with the stratigraphic framework outlined above (Section 3; Tables 1 & 2).

Clark et al., (In press) have attempted to reconstruct the pattern and timing of retreat of the BIIS based on glacial landform evidence and deglacial dates (Fig. 8). This chronological control provides a contextual template against which we can ‘fit’ and tune the relative flow-phases, derived from the geomorphology (Fig. 3b; Section 2), and stratigraphic framework outlined in Section 3 into our six-stage reconstruction (below), and gives the model greater pertinence at the ice-sheet scale.

4.1. Constraints on the onset and termination of glaciation in the central sector of the last BIIS

Temporal constraints on the onset and recession of ice in the central sector of the last BIIS are poor. However, there is now a general consensus that the major expansion of Late Devensian ice began sometime between 29-26 ka cal. BP (Fig. 8) (Huddart & Glasser, 2002; Hall et al., 2003; Bos et al., 2004; Brown et al., 2006; Telfer et al., 2009; Clark et al. In press). The deglacial history is more difficult to determine due to the asynchronous nature of retreat (cf. Clark et al., In press and Fig. 8). However, a date at Windermere places deglaciation at ca. 17.7 ka cal. BP (Coope & Pennington, 1977), while loess deposited in karstic depressions at Morecambe Bay and the Yorkshire Dales have been dated to 19 ±2 ka cal. BP and 17 ±2 ka cal. BP respectively (Telfer et al. 2009); and erratic boulders dated at Norber in the Yorkshire Dales suggest that deglaciation was as late as 18 ±1.6 ka BP (Vincent et al., 2010). However, this probably does not reflect the dynamic multiphase retreat of ice in the Solway region (Fig. 8). Moreover, the proximity of three key dispersal centres (Lake District, Southern Uplands and Pennines) and the identification of multiple re-advances (cf. Huddart & Glasser, 2002) preclude a rapid return to permanent ice-free conditions over the region. This is exemplified by a date of 14.7 ±0.6 ka cal. BP on peat reported at St. Bees in western Cumbria (Coope & Joachim, 1980), whilst exposure ages from the Southern Uplands and Lake District suggest that ice persisted in source areas after ~14.3 ka BP (McCarroll et al., 2010). Eastern England meanwhile was thought to have become ice free sometime after 15.5 ka BP (Bateman et al., 2011).

4.2. Initial expansion of ice out of upland areas:

Implicit within previous glacial reconstructions has been the advance of Scottish ice up the Vale of Eden and into the Stainmore Gap (533 m a.s.l.) (Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932a, b; Huddart, 1970; Catt, 1991). Evidence for this flow phase is limited to a suite of Scottish erratics and several exposures of the lower, red-brown to grey-brown Gillcambon Till interpreted to have been deposited during this advance (Fig. 4) (Trotter, 1929; Hollingworth, 1931; Trotter & Hollingworth, 1932; Huddart, 1970) (red dashed arrows of Stage I, Fig. 9). An alternative explanation proposed by Evans et al., (2009) is that Scottish ice encroached into NW England during a previous glaciation, and/or an early stage of the Late Devensian glaciation, and provided a ready supply of Southern Upland erratics which were subsequently dispersed by localised ice flows. If this strong flow of Scottish ice into NW England indeed occurred during the last glaciation then the prevalence of Scottish erratics throughout the stratigraphic sequence demand that it be associated with an early ice movement. This could relate to the initial expansion of ice out of upland dispersal centres. Such a movement has been identified in the NE Irish Sea Basin, with Salt and Evans (2004) and Roberts et al., (2007) both recognising that initial ice flow into the region was driven by Scottish Highlands ice flowing out of the Firth of Clyde and down the west coast (also see Boulton & Hagdorn, 2006). Further inland across southern-central Scotland the subglacial bedform evidence and dispersal trains indicate that Scottish Highlands ice did not overwhelm this region, instead reaching as far south as the upper Nith Valley, Sanquhar Gap, before subsequently being forced back as the Southern Upland dispersal centre increased in dominance (Charlesworth, 1926; Eyles et al., 1949; Sutherland, 1984; Finlayson et al., 2010). Eventually the Southern Uplands ice centre extended across the northern Irish Sea Basin into Irenland (Greenwood & Clark, 2009a). Within the central sector of the BIIS, this ice-flow stage was likely characterised by congestion and thickening of ice throughout the Vale of Eden and Solway Lowlands, reinforced by the expansion of ice out of the Lake District, Howgill Fells and Pennines. This interpretation is based upon a priori knowledge governing the general growth and expansion of ice sheets (instantaneous glacierization: Ives et al., 1975).

4.3. Stage I

Eastwards ice flow through prominent topographic corridors of the north Pennines

During stage I the ice divide straddling the northern sector of the Irish Sea Basin reached its southern-most point, pinning Galloway ice against the Cumbrian coast and forcing Lake District ice eastwards (cf. Evans et al., 2009; Fig. 9). Ice in the central sector of the BIIS had also reached its maximum observed thickness, overwhelming the entire region and moving independently of topography over cols in the northern Pennines (see Fig. 9). The highest mountains in the north Pennines, Cold Fell and Cross Fell, maintained local ice caps that fed into the main body of ice (Dakyns et al., 1891; Dwerryhouse, 1902; Trotter, 1929; Vincent, 1969; Lunn, 1995; Mitchell, 2007). The local ice-caps on the highest massifs are interpreted as cold-based plateau icefields, as inferred from the survival of pre-glacial interfluve geomorphology (Kleman & Glasser, 2007) and the paucity of subglacial bedforms indicative of temperate ice flow (Mitchell, 2007). Stage I therefore comprises the greatest recorded mass of ice in the central sector of the BIIS and is thus correlated with maximum ice sheet expansion at ~29-23 cal. ka BP (Table 1). It was at this time that the grounded Irish Sea Ice Stream reached its maximum extent, advancing into the Celtic Basin and reaching as far south as the Isles of Scilly (Scourse et al., 1990; Scourse, 1991; Hiemstra et al., 2005; Ó Cofaigh & Evans, 2007; McCarroll et al., 2010). This is supported by reworked shells from Irish Sea till from the southern coast of Ireland which provided AMS ^{14}C dates of 25-24 cal. ka BP (Ó Cofaigh & Evans, 2007; Ó Cofaigh et al., 2010a), OSL dates of 24.4-21.6 ka from deglacial outwash above the Irish Sea Till (Ó Cofaigh et al., 2010) and cosmogenic nuclide dates from Scilly of 22.1 ±2.8 to 20.9 ± 2.2 ka BP (McCarroll et al., 2010). In addition, the off-shore record indicates a peak in ice-rafted debris from Goban Spur, off the Celtic Basin, at

~25-24 ka BP (Scourse et al., 2009; Haapaniemi et al., 2010), whilst Greenwood & Clark (2009b) constrain the maximum limits of the Irish Ice Sheet to between ~28-24 ka BP (stage III-IV of Greenwood & Clark, 2009b). Similarly, the northern sector of the BIIS is thought to have extended to the continental shelf and become confluent with Scandinavian ice between ~30-25 cal. ka BP (Bradwell et al., 2008). As ice-flow during stage I was generally easterly, it is therefore inferred that large parts of the central sector of the BIIS did not function as a tributary of the Irish Sea Ice Stream during its maximum expansion to the Scilly Isles (see Fig. 9). The glacial model for the Irish Ice Sheet proposed by Greenwood and Clark (2009b) envisaged a broad ice divide centred over the northern Irish Sea Basin, extending towards Galloway during maximum ice extent, and this can be reconciled with the evidence from the central sector of the ice sheet during this phase that shows ice-flow being forced eastwards through mountain passes (Fig. 9).

The build-up of ice in the central sector of the BIIS coincided with either the advance of Scottish ice through the Stainmore Gap (see red dashed arrows, Fig. 9) and/or the development of an ice divide across the Vale of Eden/Solway Lowlands (Fig. 9) (Letzer, 1978). This is reconciled with geomorphological evidence indicating convergent flow into and across Stainmore Gap (533 m) from the NW, W and SW (ice-flow phases ES1, ST1 from Figs. 3b & 5b, d) (Hollingworth, 1931; Letzer, 1978; Mitchell, 1991a,b, 1994; Mitchell & Letzer, 2006; Livingstone et al., 2008; Fig. 7). Erratic trains of the distinctive Shap Granite and Permian Brockram (Trotter, 1929; Hollingworth, 1931; Letzer, 1978) constrain the easterly flow of Lake District ice from the Howgill Fells and the eastern sector of the Lake District (Fig. 4), and the development of another ice divide extending from the Lake District to the western Dales Ice Centre (Mitchell, 1991a, 1994) (Fig. 9). The convergence and flow of ice through the Stainmore Gap suggests that it acted as a major flow artery, transferring ice to the eastern side of the country (Fig. 5d). Flow was supplemented by north Pennine ice from the Cross Fell ice cap, and in the Vale of York and Vale of Mowbray by ice flowing out of the major valleys of the Yorkshire Dales; principally Wensleydale, which acted as a major conduit of ice from the Yorkshire Dales Ice Centre (cf. Mitchell, 1994, 2007; Mitchell et al., 2010; Fig. 9).

Erratics, particularly Shap granite, demonstrate the eastwards flow of Lake District ice through the Stainmore Gap southwards into the Vale of York as well as flowing along the east coast to Holderness as part of the North Sea Lobe (Fig. 4). Lake District erratics and Shap Granite form part of the Withernsea Member in Yorkshire (Catt & Penny, 1966; Madgett & Catt, 1978), and boulders of Shap Granite are reported on the Yorkshire coast (Fig. 4, Harmer, 1928). It is perhaps puzzling that Lake District erratics have not been observed in the lower Skipsea Till on the Yorkshire coast given that it has been dated to ~21.7 – 16.2 cal. ka BP (during stages I-III) (Catt et al., 2007; Bateman et al., 2011). Differentiating the tills by erratic content alone is difficult given the mixed provenance and similar geochemical signatures of the tills. Indeed, this complex provenance signature implies that the tills have cannibalised underlying sediments and were heavily glaciotectonised (e.g. Boston et al., 2010; Evans & Thomson, 2010). However, as the Vale of York Lobe was also fed by the Yorkshire Dales Ice Centre, perhaps the Stainmore drainage outlet initially flowed eastwards, out towards the mouth of the Tees (cf. Kendall & Wroot, 1924). As the North Sea Lobe expanded and became dominant following shutdown of the Stainmore and Tyne Gap ice-drainage pathways (Stage II-III), ice was forced southwards and the previous sediments deposited by the Stainmore Ice Stream were cannibalised and transported down the east coast. This is supported by an OSL date of 18.3 ka on the formation of Glacial-Lake Tees (Platter et al., 2000) that indicates deglaciation of the Tees Estuary prior to the deposition of the Withernsea Till and therefore a northerly source of the North Sea Lobe during stage VI (also see Roberts et al., accepted). It is also worth considering the implications of the lowermost Basement Till on the Holderness coast being Late Devensian (Eyles et al., 1994). If so, the Basement Till could correlate with LGM ice flow (stage I) and the Skipsea Till could correlate with ice advance during stage III, and this would give three major phases of ice advance as suggested in this paper for the NE sector of the Irish Sea Basin. However, the idea that the Basement Till is Late Devensian has been strongly challenged (e.g. Catt & Penny, 1966; Catt, 2007).

In the Vale of York, ice coalesced with ice from the Yorkshire Dales forming a major trunk glacier that flowed southwards, depositing the reddish brown Vale of York Till Formation and the Escrick and York moraine ridges (stage I and early in stage II; Table 1) (cf. Catt, 2007). The southerly limit of this glacier lobe is contentious, with Gaunt (1976, 1981) suggesting that it may have surged beyond the Escrick moraine into Lake Humber (which was impounded by the North Sea Lobe) and to a temporary ice limit 50 km south of this moraine, marked by a discontinuous line of gravels at Wroot and Thorne (cf. Catt, 2007 for a review, and also Bateman et al., 2008; Murton et al., 2009; Fairburn, 2011). Although still the subject of debate, Glacial-Lake Humber is thought to have existed from ~24 to at least 16.6 cal. ka BP (Bateman et al., 2000, 2008, 2011; Murton et al., 2009), with the Vale of York glacier reaching its maximum limit at ~23 ka BP (Murton et al., 2009; Chiverrell & Thomas, 2010; Clark et al., In press).

Overprinted relationships of subglacial lineations in the Stainmore Gap (Figs. 3a, 5b, 5d) (Livingstone et al., 2008) reveal a shift in the influence of ice-dispersal centres, possibly related to the southwards migration of the Vale of Eden ice divide (Fig. 9). As the influence of ice from the Vale of Eden waned, the area of the Howgill Fells-Baugh Fell became the dominant ice-dispersal centre, resulting in NE ice flow through the Stainmore Gap (ice-flow phase ST2 from Fig. 3b) (Letzer, 1978, 1987; Livingstone et al., 2008; Fig. 9).

During stage I, ice flow through the Tyne Gap (152 m) consisted of an E-ENE moving topographic ice stream (Beaumont, 1971; Bouledrou, et al., 1988; Livingstone et al., 2010c; Figs. 5b & 9). It was characterised by convergent flow from the Southern Uplands and Lake District ice centres, with tributaries of locally sourced ice flowing into the main artery from the northern Pennines (Dwerryhouse, 1902; Trotter, 1929; Clark, R. 1969; Livingstone et al., 2010c) (Fig. 9). It must be noted that if Scottish ice had flowed across the Stainmore Gap (Scenario Ia; Fig. 9) then the convergence of Lake District and Scottish ice through the Tyne Gap Ice Stream must have occurred during a later phase, or that Lake District erratics in the Solway Lowlands were dispersed eastwards by Scottish ice, producing a multi-stage transport history. The sensitivity of the Tyne Gap Ice Stream to the migration of ice divides and ice-dispersal centres is demonstrated by shifts in ice-flow direction from NE to E (ice-flow phases LT1-3 from Figs. 3b & 5b) (Livingstone et al., 2008, 2010c; Fig. 9). These shifts reflect a change in dominant regional flow through the Tyne Gap, between Lake District and Scottish Southern Upland ice (Livingstone et al. 2010c), and also migration of the Irish Sea Basin ice divide back towards Scotland, as suggested by Salt (2001), Salt and Evans (2004) and Greenwood and Clark (2009b).

Advance of the North Sea Lobe across the eastern end of the Vale of Pickering resulted in the construction of moraine ridges extending from Scarborough to Flamborough Head (cf. Catt, 2007), which dammed the eastwards drainage of the North York Moors rivers leading to the formation of Glacial-Lake Pickering, which extended c. 40 km westwards before establishing an outflow at the Kirkham Gorge (Kendall, 1902; Clark, et al., 2004; Evans et al., 2005). Given the configuration of the ice sheet during its maximal expansion (e.g. Clark et al., In press) it is hypothesised that Glacial-Lake Pickering formed during this early stage (Fig. 9).

4.4. Stage II

Cessation of the Stainmore Gap ice flow pathway and northwards migration of the North Irish Sea Basin ice divide

Stage II is characterized by a gradual reduction in ice volume resulting in the migration of ice divides back into upland regions, leading to significant changes in lowland ice dynamics primarily related to the eastwards flow of ice through the Tyne and Stainmore Gaps (Table 1). These two major ice-flow pathways are treated independently during stage II, because cross-cutting relationships fail to distinguish between the relative terminations of ice flow over each of these cols. However, given the height of the Stainmore Gap (533 m a.s.l.) compared to the Tyne Gap (152 m a.s.l.), and their relative locations compared to major ice-dispersal centres and ice divides (Fig. 10), it is logical to assume that the Stainmore Gap would be more sensitive to ice-flow dynamics.

4.4.1. Stainmore Gap

As the ice divide situated across the Vale of Eden continued to migrate southwards, ice flow across Stainmore Gap from the western side of the Pennines began to weaken. This is demonstrated by the youngest set of subglacially-formed lineations of the mountain pass, which indicate SE ice flow towards the Vale of York, dominated by ice sourced in the Upper Tees (ice-flow phases ST3-4 from Figs. 3b, 5d) (Mitchell, 2007; Livingstone et al., 2008; Table 1).

Further retreat of the ice divide and surface lowering west of the Stainmore Gap eventually resulted in a switch in ice direction towards the north (Fig. 10), thereby severing the flow of ice over the Stainmore col (Table 1) (Trotter, 1929; Letzer, 1987; Mitchell & Clark, 1994; Smith, 2002; Mitchell & Riley, 2006; Livingstone et al., 2008; Evans et al., 2009). This is clearly represented by cross-cutting patterns in the Vale of Eden, with W-E orientated drumlins preserved in Stainmore becoming increasingly scarce towards the west, where the later northwards flow from the Dales ice centre (Mitchell & Riley, 2006) exerted a greater influence in re-moulding the landscape (Fig. 5b) (cf. Evans et al., 2009). Significantly, this marked the end of its utility as a major ice transfer corridor. Preservation of the earlier convergent flow set towards Stainmore Gap was probably a result of cold-based ice, possibly associated with an ice-divide situated over Stainmore's western margin.

Further changes in ice configuration were limited to localised shifts in Pennine ice-dispersal centres, probably contemporaneous with the large-scale regional re-organisation. Cross-cutting drumlins at the head of the Tees Glacier, in a region over 600 m a.s.l., indicate a shift in ice flow from a restricted N-S (ice-flow phase ST3 from Fig. 3b & Fig. 5b) movement down the valley, to an easterly trajectory (ice-flow phase ST4 from Fig. 3b) as the ice divide expanded westwards (cf. Mitchell, 2007). Part of the Pennine ice centre, that previously contributed to the southern feeder zone of the Stainmore Ice Stream, is also thought to have been captured by the upper zone of Wensleydale Ice Stream, as evidenced by cross-cutting drumlins in Grisedale and Aisgill Moor on East Baugh Fell (Fig. 10) (Mitchell, 1991, 1994).

It remains a challenge correlating the subsequent retreat of Stainmore ice across the Pennines. However, the Vale of York Ice Lobe is thought to have been in retreat by 20.5 ka BP (Bateman et al., 2008; Clark et al., In press), which may relate to the shutdown of the Stainmore ice drainage outlet. East of Stainmore, deglaciation of the Tees Estuary is constrained by a date of 18.3 ka BP on the formation of Glacial-Lake Tees, with the lake thought to have been in existence until 16.3 ka BP (Plater et al., 2000). The lateral margin of Stainmore ice during retreat is defined by the Feldom moraine along the edge of the meltwater channel of Gilling Beck to the north of Richmond. The large channel systems that lie outside this limit at Marske and feed into the Swale are related to this ice margin (Bridgland et al., 2011). The landforms of the Vale of York show numerous drumlins associated with ice flow southwards from Stainmore and eastern England with a number of retreat phases marked by moraines. Cross-cutting landforms in the eastern part of the original Wensleydale valley near Bedale indicate that Wensleydale ice was able to extend eastwards into the Vale of York, forming the Leeming moraine after the ice retreat within the Vale of York (Bridgland et al., 2011), possibly during stage IV (Fig. 5d). An upper till, with Cheviot erratics, exposed in gravel quarries at Catterick and Scorton indicates a readvance of ice into the upper part of the Vale of York associated with the North Sea Lobe, which may be associated with Great Smeaton moraine just south of the present Tees valley (Mitchell et al., 2010) and which predates the damming of the lower Tees by the ice dam at Teesmouth (Agar, 1954).

4.4.2. Tyne Gap:

As the ice divide in the Irish Sea Basin continued to retreat into the western Southern Uplands and Northern Ireland, so the influence of the Tyne Gap as an easterly flowing ice stream weakened. Instead, ice flow became dominated by Scottish ice flowing SE down the North Tyne Valley and out of the Bewcastle Fells (Fig. 10; ice-flow phase LT4 from Figs. 3b & 5c; Table 1 (Clark, R. 2002; Livingstone et al., 2010c [stage II])). Stage II therefore corresponds with the eastwards migration of the Southern Uplands-Scottish highlands ice divide, (Salt & Evans, 2004; Livingstone et al., 2008, 2010c; Finlayson et al., 2010) causing the central Southern Uplands to become increasingly important as an ice-dispersal centre. Finlayson et al., (2010) tentatively inferred that the eastward expansion of the ice divide may have coincided with the Clogher Head Readvance (18.5-16.7 cal. ka BP), although this seems too late for our reconstruction.

4.5. Stage III

4.5.1. Stagnation and retreat of the Tyne Gap Ice Stream

Stage III was characterised by widespread deglaciation of the central sector of the BIIS (Table 1; Fig. 11). Ice in the Tyne Gap underwent incremental recession of its eastern-most margin, as delimited by a series of transverse moraines (Fig. 11) (Smythe, 1912; Livingstone et al., 2008). This resulted in decoupling of the Tyne Gap ice from the North Sea Lobe. In the lower South Tyne and Derwent valleys further retreat was associated with widespread stagnation and ablation, as evidenced by chaotic sequences of debris flow diamictons, kamiform deposits and ice-contact glaciofluvial and glaciolacustrine sediments (Clark, 1970; Douglas, 1991; Mills & Holiday, 1998; Yorke et al., 2007) (Fig. 11). Eventually, the lower Tyne and the eastern coast of northern England formed a largely ice-free enclave with ice constrained to the higher ground of the Pennines. The development of ice-free enclaves in eastern England can be reconciled with eastward migrating ice-divides and the subsequent capture and drawdown of ice into the Irish Sea Ice Stream (e.g. Livingstone et al., 2008; Finlayson et al., 2010). At Warren House Gill on the Durham coast, a sandur (the Peterlee Sand and Gravel) was deposited at the westerly margin of the expanding North Sea Lobe (Fig. 11). Ice in the Tyne Valley supplied sediment-charged meltwaters to the valley and lowlands as part of a major proglacial drainage network (Livingstone et al. 2010c, Table 1). Ice flow down the North Tyne Valley probably persisted during this (and the following) stage(s) (Table 1), while glacial flow from local Pennine ice-dispersal centres similarly remained (Fig. 11).

4.5.2. Scenario IIIa

Ice-free enclave in NW Cumbria and the development of ice-dammed lakes along the margin of the Irish Sea Ice Stream

Further deglaciation as recorded by extensive spreads of glaciofluvial and glaciolacustrine sediments (including clastic varves) and debris-flow deposits mark the onset of an ice-free enclave within the Solway Lowlands (Livingstone et al., 2010a; Fig. 11). Its western margin, off the Cumbrian coast, was delimited by ice feeding the Irish Sea Ice Stream (Fig. 11), which, towards the end of stage III, acted to impede meltwater drainage, leading to the development of proglacial lakes in the Solway Lowlands (Livingstone et al., 2010a; Fig. 11) and lower Wasdale (Merritt and Auton, 2000). Clastic varves identified at Blackhall Wood (Blackhall Wood Clay Formation) indicate that the proglacial lake (Glacial-Lake Blackhall Wood) existed for at least 261 years, while stratigraphic correlation with other laminated sediments suggest that the lake occupied an area of at least 140 km^2 (Livingstone et al., 2010a; Fig. 11; Table 1).

4.5.3. Scenario IIIb

Downwasting, stagnating Main Late Devensian ice in the Solway Lowlands

In contrast to scenario IIIa, Huddart (1970, 1983) envisaged a much more complicated series of depositional environments associated with downwasting, stagnating Main Glaciation ice in the Eden, Petteril and Caldew valleys (i.e. Stage V, Fig. 13) and slightly younger in age than the Brampton Kame Belt. Furthermore, he did not recognise any extensive glaciolacustrine unit, or a readvance (as suggested later in Stage IV). Instead, he recognised four small lakes in the Eden valley at Baronwood and Lazonby, surrounded by downwasting, stagnant ice. In the Petteril valley, the M6 motorway boreholes and sections revealed laminated clays at various altitudes and occurring either as intrabeds in till and sand or as distinct stratigraphic units. Topographically the laminated clays are situated in basins to the lee of the higher drumlinised ground to the west and in the Petteril valley to the east. Further detailed observations at Calthwaite Beck-Low Oaks (motorway chainage 645-710 and at Moss Pool (Huddart, 1970, p86-94) indicate deposition in small, subaerial, ice-marginal lakes, with associated multiple debris flow diamictons derived from adjacent tills. Similarly, Huddart (1970, 1981) reported a motorway excavation at Carrow Hill that contained sands and gravels, laminated silts and clays and supraglacially-derived mass flow diamictons, interpreted by him as an unstable ice-walled lake plain (Clayton & Cherry, 1967) that persisted for at least 87 years (varve counts). Similar depositional environments were envisaged by Huddart (1970) for exposures in the Caldew Valley to the west. At slightly lower elevations Huddart (1970) interpreted outcrops of stratified sediments at Brisco as deltas deposited into a small ice-marginal lake and at Carleton as a kame terrace sequence related to ice downwasting in the Petteril valley (Huddart, 1970, p.165-167). Lacustrine deposits found in the Caldew Valley to the west are interpreted by Huddart (1970) as a series of marginal or ice-walled lake basins. There is thus controversy in the interpretations of these sediments at Blackhall Wood: either a result of a readvance of ice down Edenside (Livingstone et al., 2010a), or as a result of continued downwasting and stagnation of the main Glaciation ice sheet (Huddart, 1970, 1981, 1983).

4.5.4. Regional significance:

Deglaciation during stages II-III (Table 1) is thought to correspond with the rapid deglaciation of the Irish Sea Ice Stream from its maximum southern limit. However, the northern sector of the Irish Sea Basin is thought to have remained glaciated during this and the subsequent stage (IV) proposed in this paper. The most compelling evidence for this is provided by Roberts et al., (2007) who propose a two-phase ice-flow model to explain glacial geomorphology on the Isle of Man; an initial SE flow from the Southern Uplands (phase I) followed by SW flow out of the Solway Firth (phase II). Both phases are attributed to streaming associated with the Irish Sea Ice Stream and are therefore attributed to the LGM (Roberts et al., 2007). Based on the Event Stratigraphy presented in Table 1, phase I of Roberts et al., (2007) can be correlated confidently with stage I and II in line with evidence indicating drainage of ice coalescing in the Solway Lowlands eastwards across the Pennines rather than into the Irish Sea as a tributary of the ice stream (Table 1). The only significant flow of ice into the Irish Sea Basin from the Solway Lowlands occurred during stage IV (ice-flow phase LT5 from Figs. 3b 5a), during the Blackhall Wood Re-advance, and phase II must therefore correlate with this event (see below). Therefore, ice is interpreted to have overrun the Isle of Man throughout stages I to IV and stretched down the west Cumbrian coast, impounding lakes against higher ground during stage III (scenario “a”) (Merritt & Auton, 2000; Livingstone et al. 2010a).

4.6. Stage IV (following on from scenario IIIa)

Blackhall Wood-Gosforth Oscillation

Despite the reservations noted above (see Scenario IIIb) and the evidence presented in Huddart (1970, 1981, 1983, 1991), an alternative view is that the overall pattern of retreat in the central sector of the BIIS was reversed during the Blackhall Wood Re-advance phase (stage IV). This re-advance was characterised by a switch in flow direction, with ice moving down the Vale of Eden from the western Pennines/Howgill Fells before curving westwards around the northern edge of the Lake District and out into the Irish Sea Basin (ice-flow phase LT5 from Figs. 3b & 5a) (Livingstone et al., 2008, 2010a; Fig. 12). Subglacial lineations constructed by Scottish ice provide evidence for a coeval advance of Southern Uplands ice out of the Esk Valley, across the Solway Firth and down the Cumbrian coast (Livingstone et al., 2008, 2010; Fig. 12), with the SW flow direction indicative of a dominant ice divide in the central Southern Uplands maintained from the previous stage. The arcuate 'swarm' of subglacial lineations (Fig. 12) display sedimentological and morphological characteristics typical of a fast-flow signature such as high elongation ratios (12:1) and evidence of pervasive glaciotectonic deformation (Stokes & Clark, 2001). It is envisaged that ice moving rapidly into the Irish Sea Basin from the Solway Lowlands coalesced with, and functioned as a tributary of, the Irish Sea Ice Stream during this stage (Livingstone et al., 2010a). The fast flow signature of stage IV subglacial lineations in the Solway Lowlands suggests that ice was being vigorously drawn-down into the Irish Sea Ice Stream (Eyles & McCabe, 1989; Evans & Ó Cofaigh, 2003; Ó Cofaigh & Evans, 2007). A northwards transition into hummocky terrain and ribbed moraine delimits the lateral margin, where inferred ice velocities were significantly reduced (Livingstone et al., 2008, 2010a; Figs. 5a & 12).

The Blackhall Wood Re-advance (Livingstone et al., 2010a) is tentatively correlated with the Gosforth Oscillation in Cumbria (Trotter, 1937), prior to which there was a temporary retreat of Lake District valley glaciers and proglacial lakes dammed against Irish Sea Ice (stage IIIa) (Merritt & Auton, 2000). The subsequent Gosforth Re-advance (stage IV, Table 1) saw Lake

District ice again coalescent with the Irish Sea Ice Stream, leading to extensive drumlinisation of the region (Merritt & Auton, 2000). However, Huddart (1970, 1991) had suggested that there was no need for the Gosforth Oscillation and that the landforms and sediments could be accommodated into earlier phases (stage III). Masking of the arcuate drumlins associated with flow-phase LT5 (Fig. 3b) by the Holme St. Cuthbert deltaic sequence (Huddart, 1970) verifies that this re-advance occurred prior to the Scottish Re-advance (Livingstone et al., 2010a; Table 1)(see below). The influence of the central sector of the BIIS on the Irish Sea Ice Stream during this stage (IV) of glaciation is replicated in the regional geological record, for example by overprinted bedforms on the Isle of Man that reveal a late SW phase of ice flow initiated in the Solway Lowlands (Roberts et al., 2007: phase II).

A considerable part of the northern Irish Sea basin might have become deglaciated following a widespread collapse of the Irish Sea Ice Stream during the interval between the LGM (stage I) and the Gosforth Oscillation Re-advance (stage V), for if the Cardigan Bay Formation within that region is correctly correlated with the LGM (Jackson et al., 1995), then the retreat must have been considerable and prolonged. This is because the overlying Upper Western Irish Sea Formation includes a widespread discontinuity, the 'X-unconformity' (Nirex, 1997), which Merritt and Auton (2000) conclude formed subglacially during the Gosforth Oscillation (see Huddart and Glasser, 2002, Fig. 5.24). A sequence of clay, silt and mud up to about 35 m thick within the Upper Western Irish Sea Formation below the X-unconformity formed in a cold-water, low-salinity, proglacial glaciomarine environment (Pantin 1977, 1978) and, based on average accumulation rates calculated for such sediments (Boulton 1990), would have taken 3000 years or more to accumulate. These coarsely laminated deposits formed within the 'Vannin Sound', which lay to the south-east of the Isle of Man, probably connected to the open sea to the south-west (Thomas, 1985). The Vannin Sound sequence has been correlated with the Dogmills Member, some 20 km to the north-west on the Isle of Man, and hence to the glacial readvance responsible for creating of the Bride Moraine on that island (Thomas, 1985). The sediments underlying the X-unconformity within the Vannin Sound contain dropstones and ice-berg dump structures and were interpreted by Pantin (1977, 1978) to have accumulated sub-tidally by sediment-laden meltwater plumes adjacent to a floating ice shelf. These sediments may be re-interpreted to have been deposited from low-velocity hyperpycnal flows, a style of sedimention that was characteristic of conditions in the North Atlantic during H1, particularly between 16.7 and 15.1 ka BP (Stanford et al., 2011). Indeed, if the X-unconformity is correctly attributed to a glacial re-advance associated with the Gosforth Oscillation, this event terminated a deglacial period in the northern Irish Sea that is similar in duration to the 'wider Heinrich event 1 sequence' of almost 4000 years proposed by Standford et al., 2011).

The regional-scale of the Blackhall Wood Re-advance (Gosforth Oscillation) implies that it was maybe triggered by an external forcing mechanism such as the 19 cal. ka BP meltwater pulse (Table 1), an event which caused an abrupt decrease in the Atlantic meridional overturning circulation leading to widespread cooling of the NE Atlantic (Yokoyama et al., 2000; Clark, et al., 2004; Hall et al., 2006); although without direct chronological control this can only be tentatively inferred (e.g. see Livingstone et al., 2010e).

A coeval re-advance of ice in the Tyne Gap may have led to deposition of the Butterby Till Member of the Wear Till Formation, which locally caps the Tyne and Wear Glaciolacustrine Formation around Durham (Francis, 1970; McMillan et al., 2011) (Table 2 & Fig. 5d). However, there is no geomorphological evidence for the Tyne Gap ice connecting with the North Sea Lobe (e.g. Livingstone et al. 2008, 2010c), with the stratigraphic record in the

Tyne Valley supporting continuous westward retreat and sandur development (York et al., 2007). It is now thought that the North Sea Lobe flowed southwards into Yorkshire (and as far south as the north Norfolk coast) during two major oscillations, between 21.7-16.2 ka BP (Skipsea Till) and 16.2-15.5 ka BP (Withernsea Till) (Bateman et al., 2011). This broadly correlates with additional dating controls along the east coast of England, which place the deposition of the Skipsea Member along the Yorkshire Coast to the Dimlington stadial (~21 cal. ka BP) (Catt & Penny, 1966; Rose, 1985; Davies et al., 2009, in press). These temporal constraints provide further evidence for the asynchronicity of the BIIS, with the eastern sector of the BIIS not reaching it maximal limits until much later, during a period where other sectors of the ice sheet where undergoing major deglacial episodes (Table 1) (Hubbard et al., 2009; Clark et al., In press). Indeed, the southwards expansion of the North Sea Lobe seems to have coincided with (and was maybe triggered by) shutdown of major easterly flowing ice drainage pathways of the central sector of the BIIS (Davies et al., 2009a; Livingstone et al., 2010c). Thus, the southerly expansion of the North Sea Lobe can be constrained to stages III-VI (Table 1). Stages IV and VI offer the most plausible options as these relate to expansions of the ice sheet along the Irish coast. Sedimentological and geomorphological evidence in Yorkshire demonstrates that the North Sea Lobe most likely underwent repeated surging as it advanced southwards (Boston et al., 2010; Evans & Thomson, 2010) and that it dammed the Humber, leading to the development of Glacial-Lake Humber (Murton et al., 2009).

4.7. Stage V

<u>Deglaciation of the Solway Lowlands:</u>

Following the Blackhall Wood Re-advance ice continued to retreat out of the lowlands of the central sector of the BIIS (Fig. 13). Initially ice receded eastwards around the northern margin of the Lake District (Hollingworth, 1931; Livingstone et al., 2010b), and north-eastwards back into the Scottish Southern Uplands (Livingstone et al., 2010b) (Fig. 13). Local glacial-flow out of the northern Lake District became decoupled as the ice retreated eastwards, with ice-marginal or subglacial meltwater channels terminating in perched fans which delimit successive still-stand positions (Hollingworth, 1931; Eastwood, et al., 1968; Greenwood et al., 2007; Table 1); similarly ice-marginal channels at Biglands constrain the retreat of Scottish ice (Fig. 13). Downwasting at the northern boundary of the receding Lake District/Eden Valley ice along the southern Solway Lowlands resulted in a considerable flux of meltwater, as recorded by glaciofluvial and deltaic deposits at Carleton (Huddart, 1970; Livingstone et al., 2010b). Such an environment is replicated throughout the northern sector of the Solway Lowlands, with tripartite divisions clearly demonstrating the later re-advance of Scottish ice (Stage VI) over glaciofluvial/glaciolacustrine deposits (Dixon et al., 1926; Trotter, 1929; Huddart, 1970; Phillips et al., 2007; McMillan et al., in press).

As ice retreated across the Irthing-South Tyne watershed onto the reverse slope of the Tyne Gap it bifurcated into two distinct ice lobes, with Lake District ice receding SW into the Vale of Eden and Scottish ice contracting west towards Dumfries-shire (Trotter, 1929; Fig. 13). During this stage Glacial-Lake Carlisle formed against the reverse slope of the Tyne Gap (Trotter & Hollingworth, 1932; Huddart, 1970; Fig. 13; Table 1). The lake was impounded against ice in the Carlisle region and characterised by falling water levels and westerly expansion in response to further retreat and down-wasting of the two ice lobes (Fig. 13). The temporal evolution of the lake margin is constrained by a series of deltas, formed between 60 and 43 m a.s.l. (Huddart, 1970; Livingstone et al., 2008). It must be noted that Glacial-Lake Carlisle might also/alternatively have formed during stage IIIa, as ice retreated westwards away from the Tyne Gap (Huddart, 1970). There is a lack of sedimentological evidence related to the Blackhall Wood Re-advance in this region, but because the glaciofluvial and glaciolacustrine deposits are positioned stratigraphically below the Scottish Re-advance diamicton, they have been correlated with stage V.

Ice in the Solway Lowlands stagnated and downwasted in situ (Fig. 13) as described in detail in Section 4.4.3 (Scenario IIIb) (cf. Huddart, 1970, 1983). It is during this phase of ice stagnation that the extensive Brampton kame belt was formed (Fig. 5a, b). This glacial landform-sediment assemblage wraps around the NW edge of the northern Pennines and is associated with a suite of meltwater channels trending along the Pennine escarpment and into the Tyne Gap (Trotter, 1929; Huddart, 1970, 1983; Arthurton & Wadge, 1981; Greenwood et al., 2007; Livingstone et al., 2008, 2010d, f). It is thus envisaged that the Brampton kame belt, and associated meltwater channels, formed in a time-transgressive manner. Initial recession, westwards across the Tyne Gap, was characterised by subglacial meltwater breaching the Irthing-Tyne watershed (e.g. Gilsland meltwater channel and subglacial esker development) and feeding into the South Tyne Valley proglacial drainage network (Livingstone et al., 2010c, d). Significantly this implies that the kame belt and Pennine escarpment meltwater channels are the upstream (subglacial) extension of the Tyne Valley drainage network. This suggests a stable hydrological regime operating at the base of this sector of the ice sheet during the last glacial (cf. Boulton et al., 2007a, b; 2009).

The physiographic position of the Brampton kame belt in the lee of the Pennines and pinned against the Penrith sandstone ridge to the west (Fig. 5a, b) facilitated stagnation and in situ downwasting of the ice mass (Huddart et al., 1970, 1983; Livingstone et al., 2010d). The spatial and temporal evolution of the Brampton kame belt resulted in an array of landforms and sediment assemblages comprising flat-topped hills, depressions and linear, sharp topped ridges (Trotter, 1929; Huddart, 1970, 1983; Livingstone et al., 2010d). These are interpreted as ice-walled lake plains as at Whin Hill (Faugh), kettle holes, and ice-contact meltwater drainage networks or fluvial crevasse fills as at Whin Hill (How Mill), Faugh, Moss Nook and Hardbanks respectively (Huddart, 1970, 1983; Livingstone et al., 2010d). Livingstone et al., (2010d, g) inferred that this type of landform-sediment assemblage could have formed by enlargement of a complex glacier karst (sensu Clayton, 1964). The deposition of large glaciofluvial moraine complexes has typically been associated with inter-lobate (suture) zones (e.g. Warren & Ashley, 1994; Thomas & Montague, 1997; Punkari, 1997). Interestingly, a narrow suture zone is likely to have developed where ice flowing NW down the Vale of Eden was joined by ice emanating from the S-N orientated Pennine ice divide (see ice flow vectors in Fig. 12). Both the Pennine escarpment meltwater network and the Brampton kame belt are located along this narrow corridor and is therefore a plausible mechanism for their formation. Suture zones are associated with thinner ice-configurations that concentrate meltwater drainage, whilst the unzipping of interlobate ice masses can lead to ice stagnation and the net deposition of glaciofluvial sediment (Punkari, 1997).

The Brampton kame belt was fed by a large meltwater channel network trending along the edge of the Pennine escarpment (Fig. 5b) (Totter, 1929; Arthurton & Wadge, 1981; Greenwood et al., 2007; Livingstone et al., 2010d, f), and also from meltwater draining off the Penrith sandstone ridge (Arthurton & Wadge, 1981; Livingstone et al., 2010d, f). The drainage network documents the progressive deglaciation and down-wasting of ice in the Vale of Eden (stages V and VI) and comprises anastomosing subglacial channels and flights of lateral channels. The morphology of the meltwater system as a whole is best explained as time-transgressive, with channels running parallel to each other and formed at successively lower elevations (Livingstone et al., 2008). Ice surface lowering eventually led to the Brampton kame belt being cut-off from the meltwater drainage network. It is also tentatively suggested that the current River Eden was occupied by a large tunnel valley during the last glaciation (Livingstone et al., 2010d, f). The dense network of meltwater channels on the Penrith sandstone ridge (e.g. Livingstone et al., 2010f), and formation of an ice-contact glaciofluvial complex at Baronwood (cf. Huddart, 1970) provides a further example of ice being pinned against and stagnating on top of the ridge during ice surface lowering in the Vale of Eden (Huddart, 1970).

Further meltwater channels have been mapped at the southern end of the Vale of Eden (Letzer, 1978; Livingstone et al., 2008), comprising four distinct dendritic networks, each of which leads into a major trunk channel. (Letzer, 1978; Livingstone et al., 2008). These meltwater systems trend SSW-NNE, running parallel to the orientation of the drumlins in the region (stages III-V) and are interpreted to be subglacially formed due to their complex morphology and discontinuous profiles (cf. Letzer, 1978).

As the east coast became deglaciated as Pennine ice retreated westwards large volumes of meltwater became trapped between the western flank of the North Sea Lobe, which was still extensive at this time (see Bateman et al. 2011), and higher ground to the west. This resulted in numerous proglacial lakes developing in the zone between Pennine and North Sea ice, including Glacial-Lake Tees and Glacial-Lake Wear (Fig. 13) (Kendall 1902; Penny & Rawson 1969; Smith & Francis 1967; Smith 1981, 1994).

4.8. Stage VI

Scottish Re-advance and subsequent final retreat of ice out of the central sector of the BIIS

The final stage in the observed glacial history of the BIIS is the re-advance of Scottish ice onto the fringe of the west Cumbrian coast as far south as Annaside-Gutterby (ice-flow phase SF1 from Figs. 3b & 5a) (Trotter, 1922, 1923, 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1971a, b, 1991, 1994; Huddart & Tooley, 1972; Huddart et al., 1977, Huddart & Clark, 1994; Livingstone et al., 2010b; Stone et al., 2010) (Fig. 14; Table 1). A still-stand recorded by the Holme St. Cuthbert deltaic sequence, moraines at Annaside-Gutterby and St. Bees, proglacial lacustrine deposits in Wasdale and proglacial sandur deposits at Harrington and Broomhills mark the most obvious limits of the Scottish Re-advance (Table 1) (Huddart & Tooley, 1972; Huddart et al. 1977; Huddart, 1994; Merritt & Auton, 2000). A radiocarbon date of 14.7 ±0.6 ka BP from an infilled hollow at the top of the moraine sequence at St. Bees provides a maximum constraint on the re-advance (Coope & Joachim, 1980). The Jurby Formation on the Isle of Man, which was deposited between 18.5-14.3 cal. ka BP has also been tentatively correlated with the Scottish Re-advance, with deposition inferred to be associated with dynamic retreat of the Irish Sea Ice Stream (cf. Thomas et al., 2004), whilst McCabe et al. (1998) and McCabe & Clark (1998) have proposed a correlation with the Killard Point Stadial (~16.8 cal. ka BP) as part of a pan Irish Sea response to Heinrich Event 1 (Table 1). However, the poor chronological constraints in the central sector of the BIIS, coupled with the ill-defined imprint and inferred transient behaviour of the Scottish Re-advance make it difficult to determine conclusively whether the Scottish Re-advance formed part of a regional re-advance signal or a local internal re-adjustment of this part of the British-Irish Ice Sheet.

The Holme St. Cuthbert delta provides evidence for ice-dammed lake formation in the vicinity of Wigton (Fig. 10), with the foreset structures revealing a lake height between 42-49 m a.s.l. and a water depth of ~30 m (Huddart, 1970; Huddart & Tooley, 1972; Livingstone et al., 2010b). Subdued eskers and a thin, patchy upper till sheet in the Solway Lowlands have allowed the margin to be traced as far inland as Thursby, Sowerby Wood and Carleton and possibly as far east as Lanercost (Dixon et al., 1926; Trotter et al., 1929; Trotter &

Hollingworth, 1932; Huddart, 1970, 1973, 1994; Livingstone et al., 2010b). Huddart (1970) suggested that in the Carlisle plain the readvance ice reworked marginal slopes but did not override deltas at 30m a.s.l. He located an upper till as far east as Greenholme (near Corby Hill), but this did not reach as far as Trotter's (1929) readvance margin between Lanercost, Brampton and Cumwhitton. The thin geometry of the till sheet, coupled with the lack of observable strandlines and glaciolacustrine sediments associated with the ice-dammed lake suggest that the Scottish Re-advance probably represents a short-lived event (Trotter, 1929; Livingstone et al., 2010b). The composition of the Annaside-Gutterby and St. Bees moraines (complex glaciotectonised sequences of pre-existing sediments) do not necessitate repeated or prolonged subglacial transport of sediment and is therefore compatible with this model (e.g. Williams et al., 2001; Evans & Ó Cofaigh, 2003).

A lobate re-advance of Lake District ice is evident from a S-N orientated flowset of subglacial lineations (ice-flow phase LT6 in Vale of Eden from Fig.'s 3b and 5a, b) encroaching into the Solway Lowlands from the Vale of Eden (Livingstone et al., 2008, 2010a,b; Fig. 14). This re-advance is bounded at its northern end by a belt of thick diamicton interpreted as an end-moraine, while meltwater channels emanating from the former glacier margin, and coalescing with the Dalston overspill channel, are also evident at the northern edge of the Vale of Eden (Livingstone et al., 2008, 2010a,b; McMillan et al., in press). It is, however, difficult to reconcile whether this event was coeval with the Scottish Re-advance or merely a late-stage internal re-adjustment of Lake District-Howgill Fells ice (Livingstone et al., 2010b; Merritt, 2010).

As ice down-wasted and retreated into upland massifs, flow became topographically constrained, as illustrated by a series of final late-stage ice-flows which are clearly restricted to valleys of the main upland dispersal centres (ice-flow phases LT7 from Fig. 3b) (e.g. Charlesworth, 1926; Raistrick, 1926; Trotter, 1929; Hollingworth, 1931; Salt & Evans, 2004; Livingstone et al. 2008). Cosmogenic exposure ages from Wasdale in the Lake District and Glen Trool in the Southern Uplands suggest that ice persisted in these upland dispersal centres after ~14.3 cal. ka BP (McCarroll et al., 2010).

In comparison to the NW coast, the dating control in eastern England permits more direct correlation between ice-flow events and forcing mechanisms. During Heinrich Event 1 (which may have occurred during stage VI), the North Sea Lobe continued to surge southwards, damming Glacial-Lake Humber and depositing the Withernsea Till (Fig. 14) (Bateman et al., 2008, 2011; Evans & Thomson, 2010; Davies et al., in press). Pennine ice had retreated from the coast by this time, leaving ice-dammed lakes in eastern England, with the North Sea Lobe still acting as a barrier to meltwater drainage (Fig. 14).

5. CONCLUSIONS

This review demonstrates the inherent dynamism of the central sector of the BIIS during the last glaciation. The ice sheet was characterised by flow switches, initiation (and termination) of ice streams, draw-down of ice into marine ice streams, repeated ice-marginal fluctuations and the production of large volumes of meltwater, locally impounded to form ice-dammed glacial lakes. Numerical ice-sheet modelling demonstrates that ice-flow switches can occur rapidly over short time-scales (Evans et al., 2009; Hubbard et al., 2009). This depiction of a dynamic ice sheet heavily influenced by fast-flowing ice streams is in accord with recent fieldwork and modelling studies that recognise dynamic shifts in ice flow direction during the glaciation of the last BIIS (e.g. Greenwood & Clark, 2008, 2009a,b; Hubbard et al., 2009; Finlayson et al., 2010). The topographic complexity of the region provides an underlying control for the multi-phase ice flow patterns exhibited. Upland regions provided both the initial and final focal points from which ice radiated (under topographic control), while during later phases ice was streamed through topographic lows (e.g. Tyne Gap) and was influenced by changes in the dominance of upland ice-dispersal centres. Dynamic shifts in flow direction were primarily driven by the migration of ice-dispersal centres and ice divides. This is depicted in the generalised migration patterns of ice divides and ice-dispersal centres of the BIIS, with ice first expanding and then contracting back into upland massifs (e.g. Evans et al., 2009). Ice streams both within and in the immediate vicinity of the central sector of the BIIS also influenced the dynamics of the ice sheet through the draw-down (and thus thinning) of the ice (e.g. Irish Sea Ice Stream). This impacted upon the migration of ice divides initiating rapid changes in flow behaviour and direction; e.g. in the Tyne Gap. Ice streams themselves are observed to be non-permanent and highly sensitive features of the BIIS; with the Tyne Gap, for example, shown to be heavily influenced by both the Southern Uplands and Lake District ice-dispersal centres, before eventually disintegrating as the Irish Sea ice divide shifted northwards. The Irish Sea Ice Stream however, did not influence the central sector of the BIIS until a late stage of deglaciation (stage IV), probably with the ice divide in the northern part of the Irish Sea Basin instead providing the major flux of ice from this sector.

Six major phases of ice-flow activity have been recognised in this paper and these are summarised below and detailed in Table 1 and Figs. 9-14:

1. Stage I comprised the greatest recorded mass of ice in the central sector of the last BIIS and is therefore correlated with maximum ice sheet expansion (~29-23 cal. ka BP). Ice drained eastwards through prominent topographic corridors (Tyne Gap and Stainmore Gap) of the northern Pennines.
2. Stage II was characterised by a gradual reduction in ice volume resulting in the migration of ice divides back towards uplands massifs. This eventually caused the cessation of the Stainmore Gap ice-flow path and a switch to northerly ice flow down the Vale of Eden into the Solway Lowlands. This ice-flow switch may have been coincident with the shutdown of the Vale of York Ice Lobe. The easterly flowing Tyne Gap Ice Stream likely weakened during this stage due to the increased importance of the central Southern Uplands ice-dispersal centre, with ice flow shifting to a SE direction down the North Tyne Valley.
3. Stage III was associated with widespread deglaciation of the central sector of the last BIIS. The Tyne Gap ice stream decoupled from the North Sea Lobe leaving an ice-free enclave in eastern England. This allowed a large proglacial drainage network to develop in the Tyne Valley, whilst a sandur was deposited on the Durham coast against the western margin of the expanding North Sea Lobe. We offer two scenarios for ice retreat in the Solway Lowlands: IIIa – the formation of an ice-free enclave constrained by the Irish Sea Ice Stream to the west and characterised by proglacial lake development; or IIIb – stagnation and in situ downwasting of ice in the valleys of the Solway Lowlands. The fundamental difference between these two scenarios is that IIIb does not recognise the stage IV re-advance (which is based on varved glaciolacustrine sediments sandwiched between two till units), and is therefore the equivalent of stage V of the IIIa model (see Table 1).
4. Stage IV records the re-advance of ice into the Solway Lowlands (Blackhall Wood-Gosforth Oscillation) and is associated with a switch in ice direction and behaviour, with ice flowing rapidly into the Irish Sea, perhaps as a tributary of the Irish Sea Ice Stream.
5. Stage V was characterised by the deglaciation of the Solway Lowlands, with

stagnation and downwasting significant (see stage IIIb). This led to the formation of an extensive glaciofluvial complex in the lee of the Pennines (Brampton kame belt), which formed part of a larger meltwater network trending along the Pennine Escarpment and into the Tyne Gap. As the east coast of England became deglaciation as Pennine ice retreated westwards numerous proglacial lakes developed against the flank of the North Sea Lobe, including Glacial-Lake Tees and Glacial-Lake Wear.

6. The final stage (VI) was the re-advance of Scottish ice onto the fringe of the west Cumbrian coast as far south as Annaside-Gutterby, and was followed by the final retreat of ice out of the central sector of the last BIIS.

During stages IV to VI the North Sea Lobe flowed southwards into Yorkshire, during two major oscillations between 21.7-16.2 ka BP (Skipsea Till) and 16.2-15.5 ka BP (Withernsea Till) (Bateman et al. 2011). Geomorphological and sedimentological evidence demonstrates that the North Sea Lobe underwent repeated surging as it advanced southwards (Boston et al. 2010; Evans & Thomson, 2010), whilst it also dammed the Humber leading to the formation of Glacial-Lake Humber until after 16.6 ka BP.

Two major oscillations (the Blackhall Wood Re-advance and Scottish Re-advance) have been identified within the central sector of the BIIS. Both re-advances may be part of more regionally extensive Irish Sea Ice Basin re-adjustments and possibly related to the Gosforth Oscillation at ~19.5 cal. ka BP (Merritt & Auton, 2000) and Killard Point Stadial at ~16.8 cal. ka BP (McCabe et al., 2007) respectively although the absence of chronological control from the central sector of the ice sheet make these correlations tentative.

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Figures:

Fig. 1: Location map showing bounds of central sector of the last BIIS, orography, key towns and fieldsites mentioned in the text, and major rivers. Dashed boxes show the locations of Fig. 5. H-St C: Holme-St. Cuthbert; BW: Blackhall Wood; MN: Moss Nook; WH(F): Whin Hill (Faugh); HB: Hardbank; HM: How Mill; Gr: Greenholme; PSR: Penrith Sandstone Ridge; A-I-W: Appleby-In-Westmorland.

Fig. 2: NEXTMap image illustrating the complex topography & array of glacial landforms of northern England and southern Scotland (lines refer to transects in Fig. 7).

Fig. 3: Geomorphological maps: (a) Generalized ice flow directions in northern England and southern Scotland (after Taylor et al.1971): (b) Generalised ice-flow phases in the central sector of the last BIIS during the Late Devensian and Table showing relative chronology flow stacks from cross-cutting bedform relationships (from Livingstone et al. 2008). Flowsets are stacked vertically according to cross-cutting relationships (with the bottom block (e.g. LT1) being the oldest and the top block the youngest (e.g. LT6)) with horizontal blocks denoting discrete regional groupings where cross-cutting relationships can be directly observed.

Fig. 4: Distribution of glacial erratics in the central sector of the last BIIS (modified from Harmer, 1928).

Fig. 5: NEXTMap DEMs of the key flow corridors of the central sector of the last BIIS, showing the relationship between the glacial geomorphology, flow vectors (based on Fig. 3b) and key field sites: (a) Solway Lowlands; (b) Vale of Eden; (c) Tyne Gap; and (d) Stainmore Gap. The dotted arrows in Fig. 5d refer to general flow directions that were not outside of the relative chronology derived in Fig. 3b.

Fig. 6: Distribution of named surficial units of till within the Caledonia Glacigenic Group in northern England and southern Scotland (after McMillan et al., 2011)

Fig. 7: (a) Generalised stratigraphy of western Cumbria (after Merritt and Auton, 2000); (b) generalised stratigraphy of the Solway Lowlands (from BGS, 2006); (c) generalised stratigraphy of Dumfries-shire (from Stone et al., 2010); and (d) generalised stratigraphy of NE England (from Stone et al., 2010).

Fig. 8: (a) Reconstruction of the pattern of retreat of the BIIS based on the distribution of moraines, eskers, ice-dammed lakes and drumlins and their relationship to bed topography. Solid black lines record palaeo-margins and dotted lines are interpolations between them (from Clark et al. In press); and (b) Isochrones of ice retreat of the BIIS based on retreat margin positions (Fig. 5a) and the compilation of dates associated with retreat (from Clark et al. in press).

Fig. 9: Stage I: Eastwards ice flow through prominent topographic corridors of the north Pennines during maximum expansion of the BIIS. Dashed-dotted lines refer to ice divides (with the thick dotted lines indicating possible ice-saddles) and the arrows indicate ice flow vectors (dotted arrows indicate alternative ice flow scenarios). This reconstruction is based on this review of the central sector of the last BIIS, integrated with regional ice sheet reconstructions (e.g. Salt & Evans, 2001; Finlayson, 2010), new dates from eastern England (Bateman et al. 2011) and ice-sheet scale reconstructions (Hughes, 2008; Clark et al. 2011). The inset diagrams are based on reconstructed British Ice Sheet time slices from Hughes (2008). The ice divide in the Vale of Eden is inferred from lineation orientations leading into the Stainmore Gap. The red dashed vectors refer to Scenario Ib, with Scottish ice flowing into the Stainmore Gap. Although this diagram depicts one moment in time, during stage I both the Stainmore Gap and Tyne Gap were associated with shifting ice flow directions in response to

migrating ice divides and ice-dispersal centres. Glacial-Lake Pickering is inferred, based on the ice sheets configuration.

Fig. 10: Stage II: Cessation of the Stainmore Gap ice flow pathway and northwards migration of the North Irish Sea Basin ice divide. Eastward flow through the Tyne Gap was superseded by SE flow of ice down the North Tyne Valley and out of Bewcastle Fells due to the eastward migration of the Southern Upland ice divide.

Fig. 11: Stage III: Stagnation and retreat of the Tyne Gap Ice Stream. This resulted in the decoupling of the Tyne Gap Ice Stream from the North Sea ice and development of a major proglacial drainage network in the eastern Tyne Valley. An ice-free enclave developed on the east coast of northern England, with a large proglacial sandur accreted during initial advance of the North Sea Lobe. In scenario IIIa, the lowland regions of NW Cumbria also became ice-free, with ice-dammed lakes (Glacial-Lake Blackhall Wood) developing against the eastern margin of the Irish Sea Ice Stream. In contrast scenario IIIb envisages a much more complex series of depositional environments associated with downwasting, stagnating Main Glaciation ice in the Solway Lowlands (see Stage V). The ice-limits during this retreat are not very well constrained, both on the eastern and western coast, and are therefore speculative.

Fig. 12: Stage IV: Blackhall Wood Re-advance (from Scenario IIIa). Reversal of ice flow associated with drawdown of ice into the Irish Sea Ice Stream. On the eastern side of the Pennines it is envisaged that the North Sea Lobe was coevally surging southwards as far as Norfolk. This resulted in the formation of Glacial-Lake Humber. It is inconclusive whether the Tyne Gap re-connected with the North Sea Lobe during this stage and is therefore only shown to be for simplicity. Similarly, it is not known when Stainmore fully deglaciated and this may have occurred much early than illustrated here.

Fig. 13: Stage V: Deglaciation of the Solway Lowlands. This stage was initially associated with clockwise retreat of ice around the northern margin of the Lake District followed by ice stagnation in the Vale of Eden and Solway Lowlands (cross-hatching). Scottish ice retreated back towards the Southern Uplands. As ice was downwasting during this stage it is difficult to reconcile ice thicknesses. However, it is likely that nunataks become prevalent over the Lake District and Pennines during this stage and therefore that flow was constrained by topographic divides (faint blue dashed-line). Stagnation of ice in the lee of the Pennines, possibly as part of a suture zone between Vale of Eden and Pennine ice resulted in development of the Brampton kame belt. This glaciofluvial feature was associated with a large glacial hydrological system trending along the Pennine escarpment and across the Tyne Gap. Melt-water ponded against the reverse slope of the Tyne Gap as ice retreated out of the Solway Lowlands leading to the formation of Glacial-Lake Carlisle.

Fig. 14: Stage VI: Scottish Re-advance and subsequent final retreat of ice out of the central sector of the BIIS. Scottish ice dammed the drainage of meltwater out of the Solway Lowlands leading to the formation of Glacial-Lake Wigton in NW Cumbria and also potentially Glacial-Lake Carlisle against the reverse-slope of the Tyne Gap. The blue hatched line indicates a possible coeval re-advance of Lake-District-Pennine-Howgill Fell ice down the Vale of Eden. The Lake District and Pennines would probably have been characterised by valley glaciation by this time and it is therefore not shown on this figure. The North Sea Lobe was repeatedly surged during this stage with ice-dammed lakes forming against its western flank (Glacial-Lake Pickering, Glacial-Lake Wear).

Tables:

Table 1: Event Stratigraphy for the central sector of the last British-Irish Ice Sheet.

Table 2: Summary of tills in the central sector of the last BIIS.

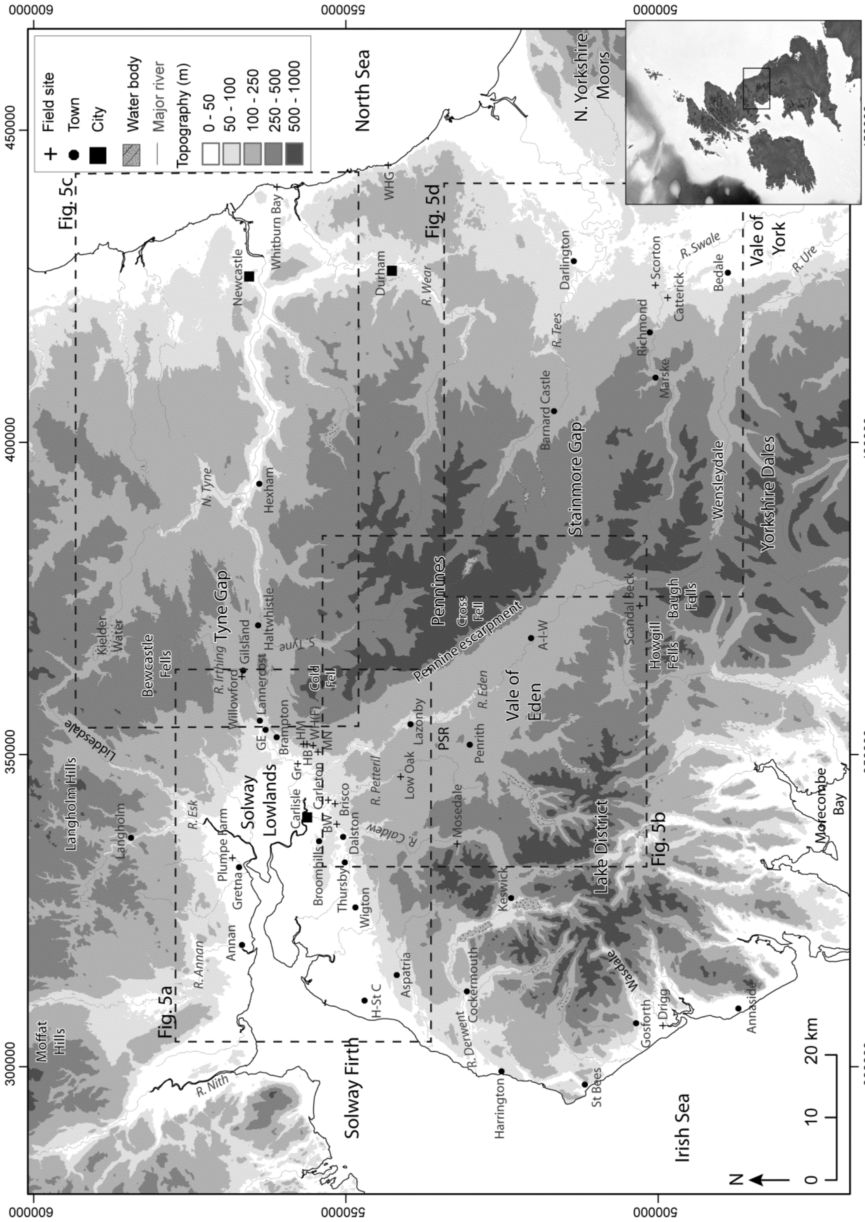

Field site
Town
City
Water body
Major river
Topography (m)
0 - 50
50 - 100
100 - 250
250 - 500
500 - 1000
North Sea
Irish Sea
Solway Firth
Morecambe Bay
Fig. 5a
Fig. 5b
Fig. 5c
Fig. 5d
Solway Lowlands
Tyne Gap
Pennines
Pennine escarpment
Cross Fell
Vale of Eden
Lake District
Stainmore Gap
Yorkshire Dales
Vale of York
N. Yorkshire Moors
Howgill Fells
Baugh Fells
Bewcastle Fells
Kielder Water
Langholm Hills
Moffat Hills
Liddesdale
Cold Fell
Wensleydale
Wasdale
R. Nith
R. Annan
R. Esk
R. Irthing
N. Tyne
S. Tyne
R. Petteril
R. Caldew
R. Eden
R. Derwent
R. Wear
R. Tees
R. Swale
R. Ure
Newcastle
Durham
Carlisle
Whitburn Bay
WHG
Darlington
Scorton
Catterick
Richmond
Marske
Bedale
Barnard Castle
Hexham
Haltwhistle
Gilsland
Lanercost
Willowford
Brampton
GE
HM
HB
WH
MN
GH
Carleton
BW
Brisco
Dalston
Broomhills
Thursby
Wigton
Plumpe Farm
Gretna
Langholm
Annan
Low Oak
Lazonby
PSR
Penrith
Mosedale
A-L-W
Scandal Beck
Keswick
Cockermouth
Aspatria
H-St C
Harrington
St Bees
Gosforth
Drigg
Annaside
20 km
N
300000
350000
400000
450000
500000
550000
600000

Fig. 1

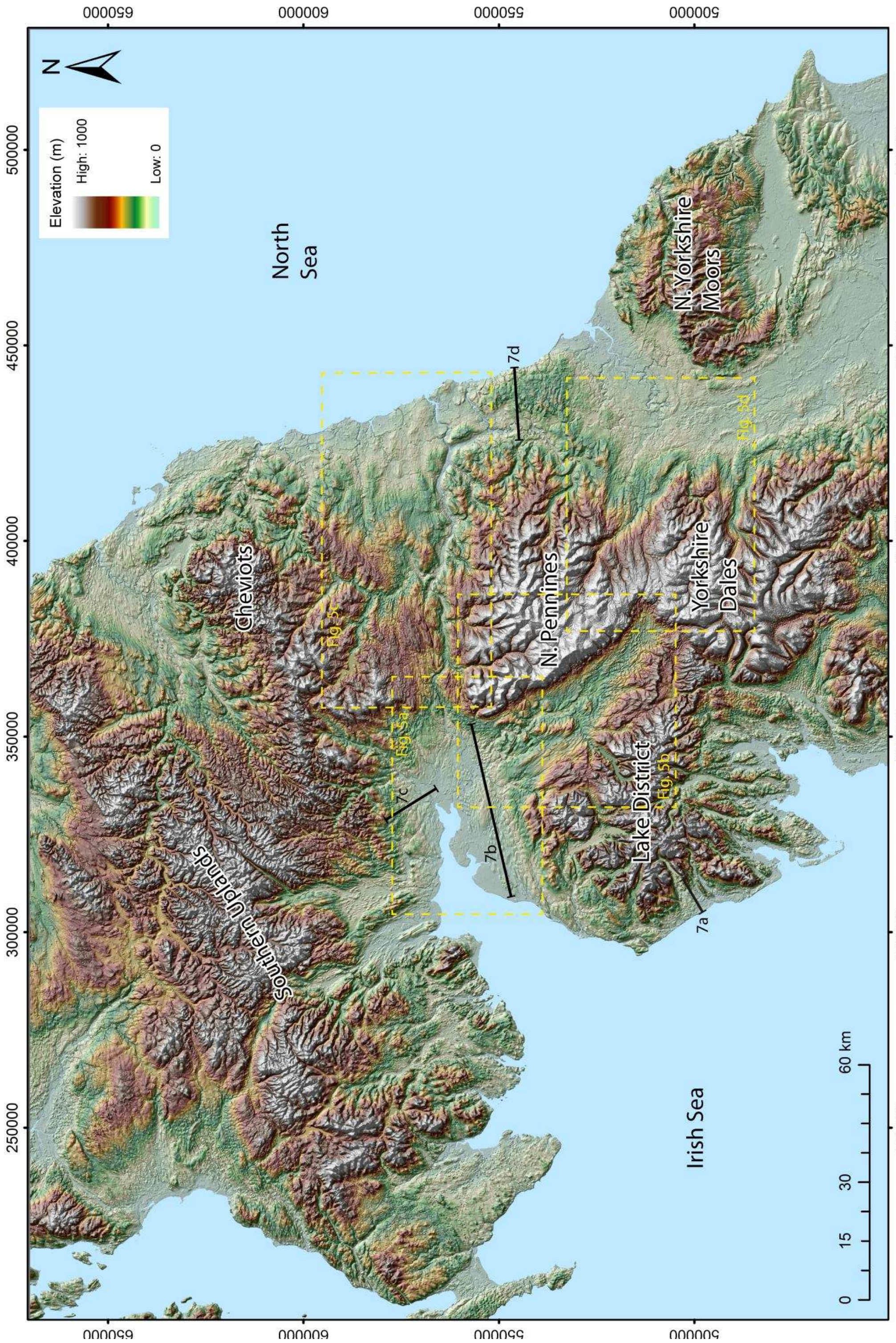

North Sea
Irish Sea
Elevation (m)
High: 1000
Low: 0
Cheviots
Southern Uplands
N. Pennines
Lake District
Yorkshire Dales
N. Yorkshire Moors
Fig. 5a
Fig. 5b
Fig. 5c
Fig. 5d
7a
7b
7c
7d
60 km
0
15
30
250000
300000
350000
400000
450000
500000
550000
600000
650000

Fig. 2

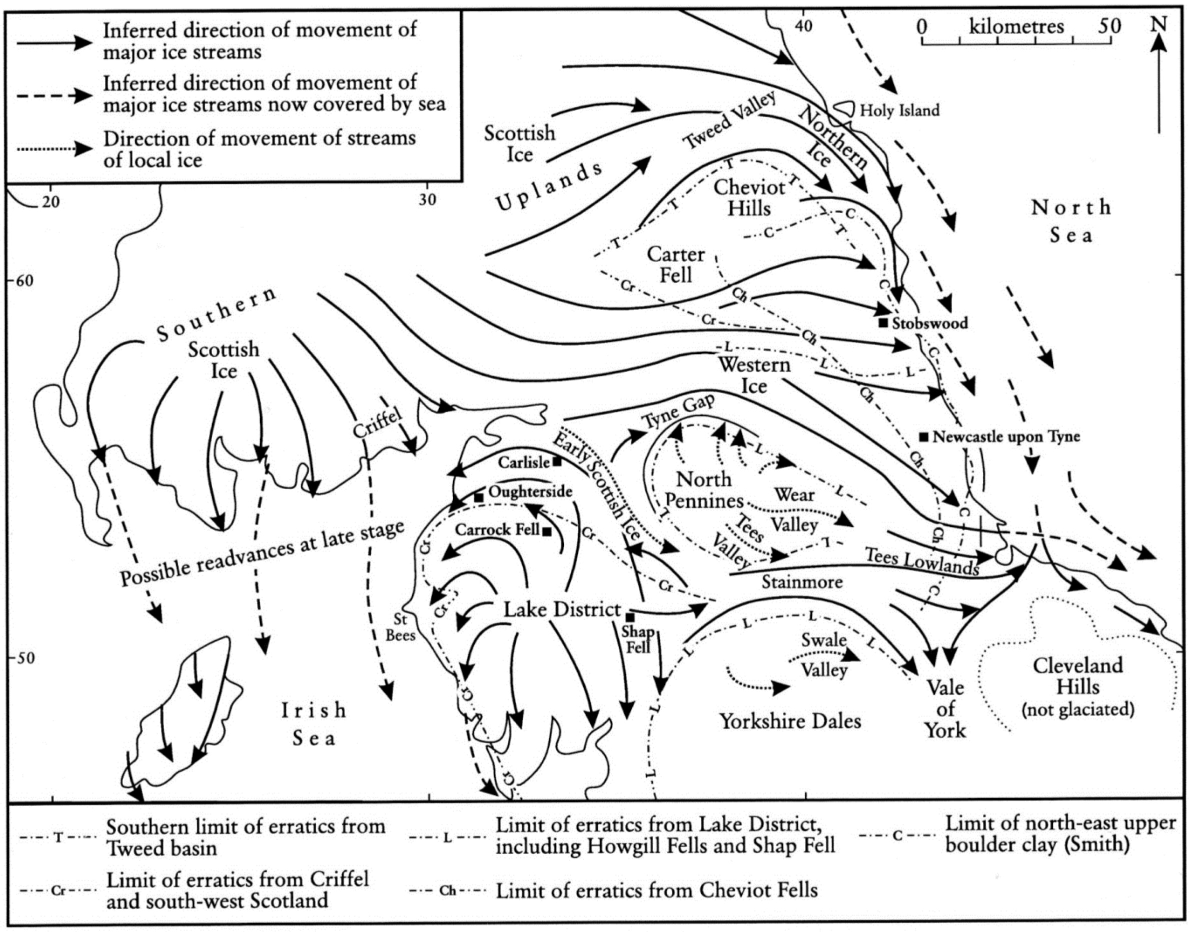

Inferred direction of movement of major ice streams
Inferred direction of movement of major ice streams now covered by sea
Direction of movement of streams of local ice
0 kilometres 50
N
40
20
30
60
50
Holy Island
Scottish Ice
Uplands
Tweed Valley
Northern Ice
Cheviot Hills
North Sea
Carter Fell
Stobswood
Southern
Scottish Ice
Western Ice
Criffel
Newcastle upon Tyne
Tyne Gap
Carlisle
Early Scottish Ice
North Pennines
Wear Valley
Oughterside
Carrock Fell
Tees Valley
Tees Lowlands
Possible readvances at late stage
Stainmore
St Bees
Lake District
Shap Fell
Swale Valley
Cleveland Hills (not glaciated)
Irish Sea
Yorkshire Dales
Vale of York
Southern limit of erratics from Tweed basin
Limit of erratics from Lake District, including Howgill Fells and Shap Fell
Limit of north-east upper boulder clay (Smith)
Limit of erratics from Criffel and south-west Scotland
Limit of erratics from Cheviot Fells

Fig. 3a

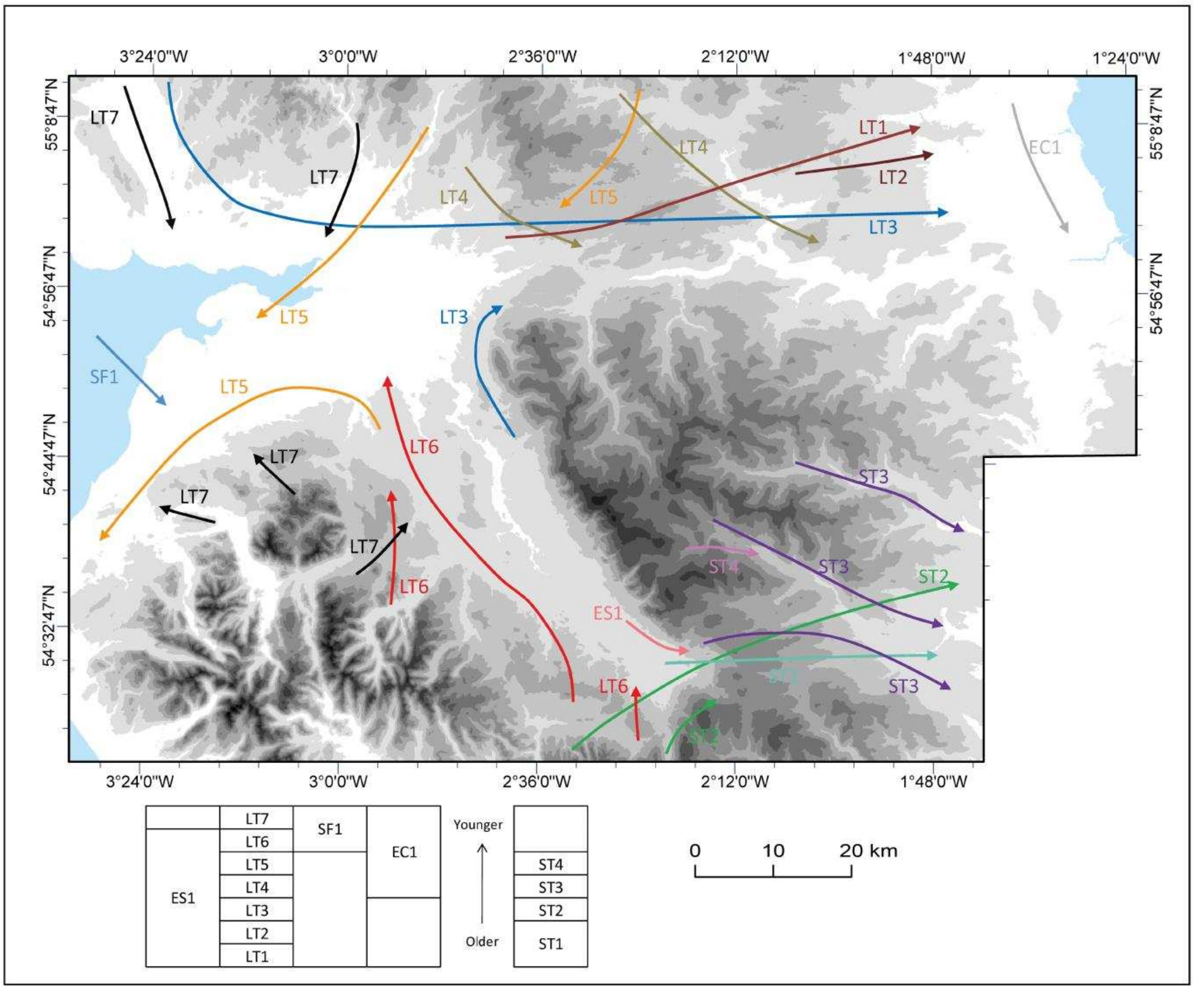
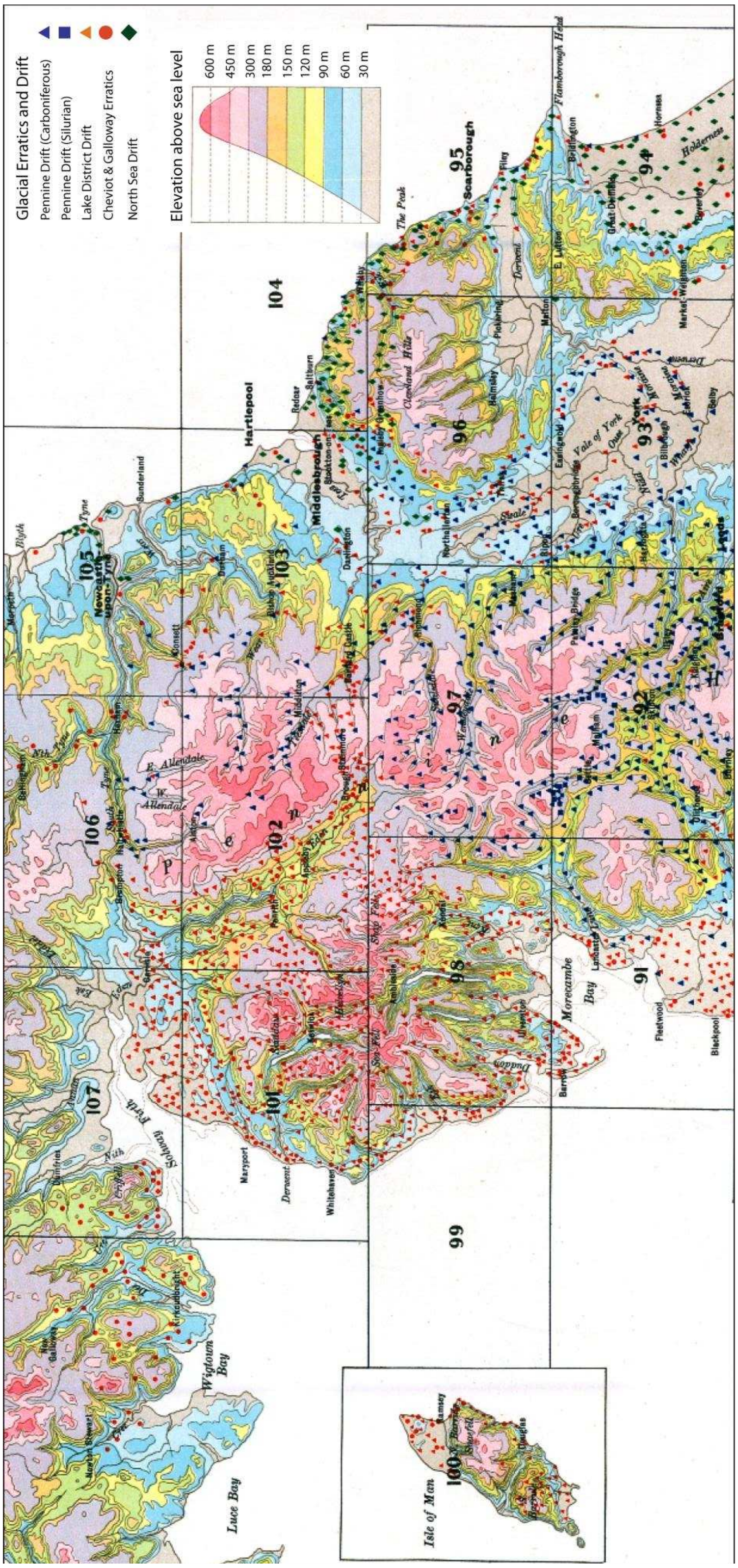
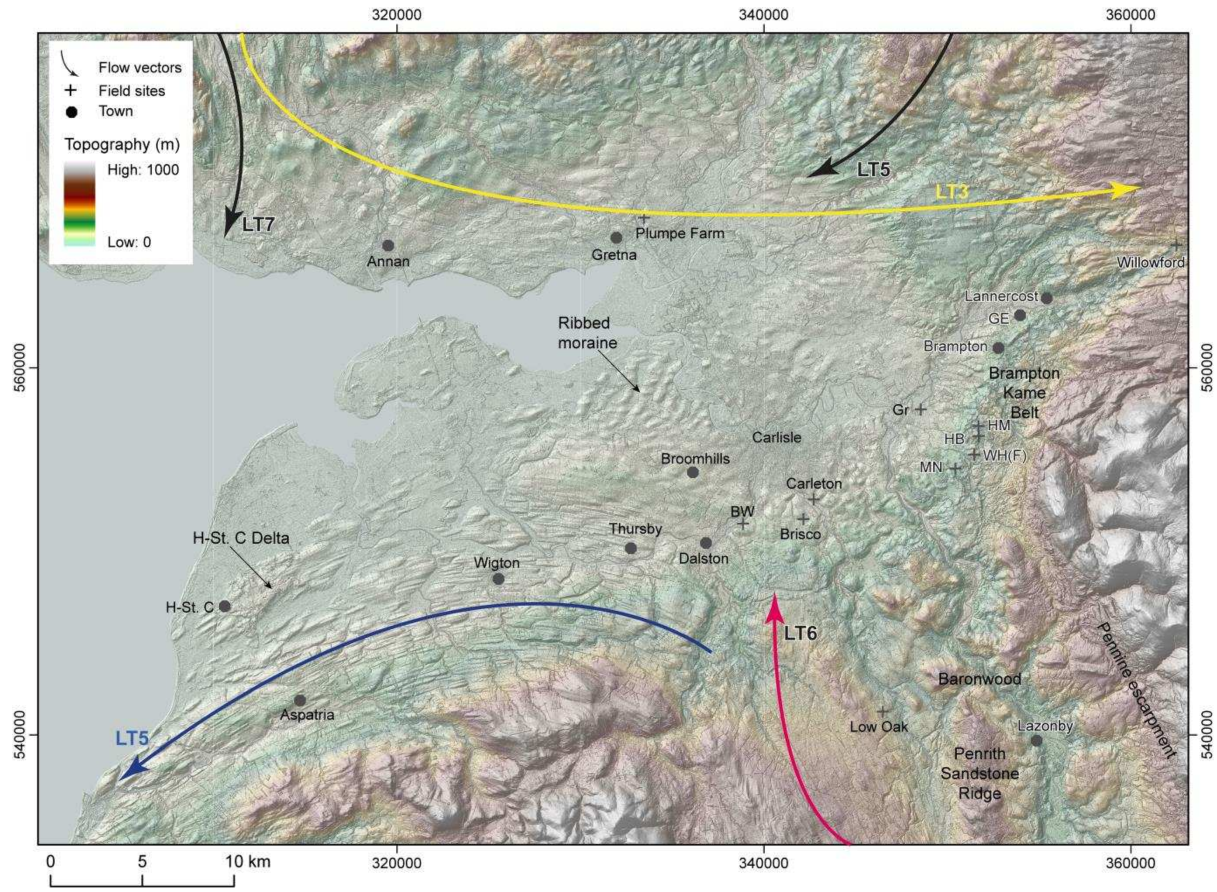

3°24'0"W
3°0'0"W
2°36'0"W
2°12'0"W
1°48'0"W
1°24'0"W
55°8'47"N
54°56'47"N
54°44'47"N
54°32'47"N
LT1
LT2
LT3
LT4
LT5
LT6
LT7
EC1
SF1
ES1
ST2
ST3
ST4
0 10 20 km
Younger
Older
ST1

Fig. 3b

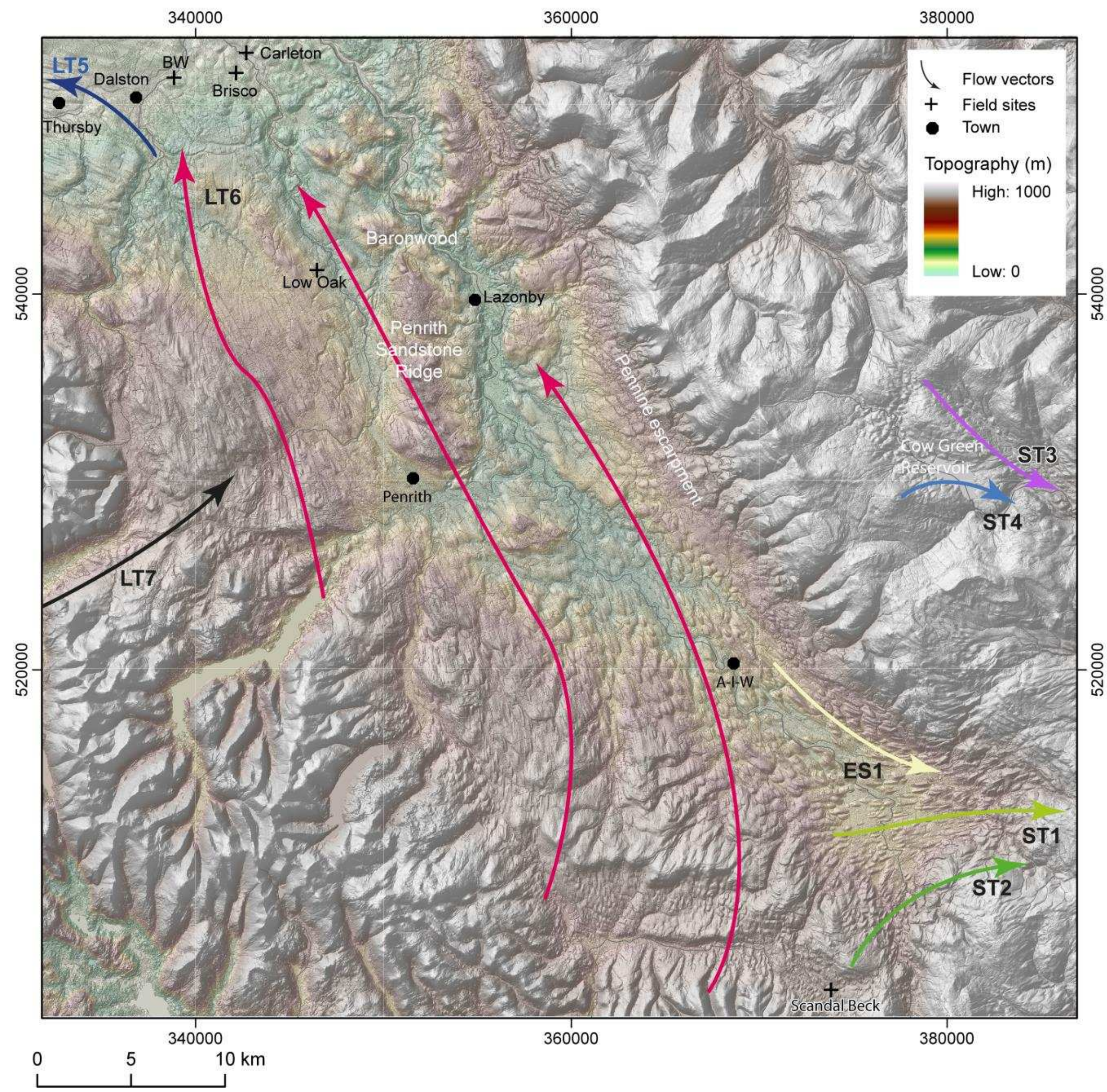


1857
1858 Fig. 4



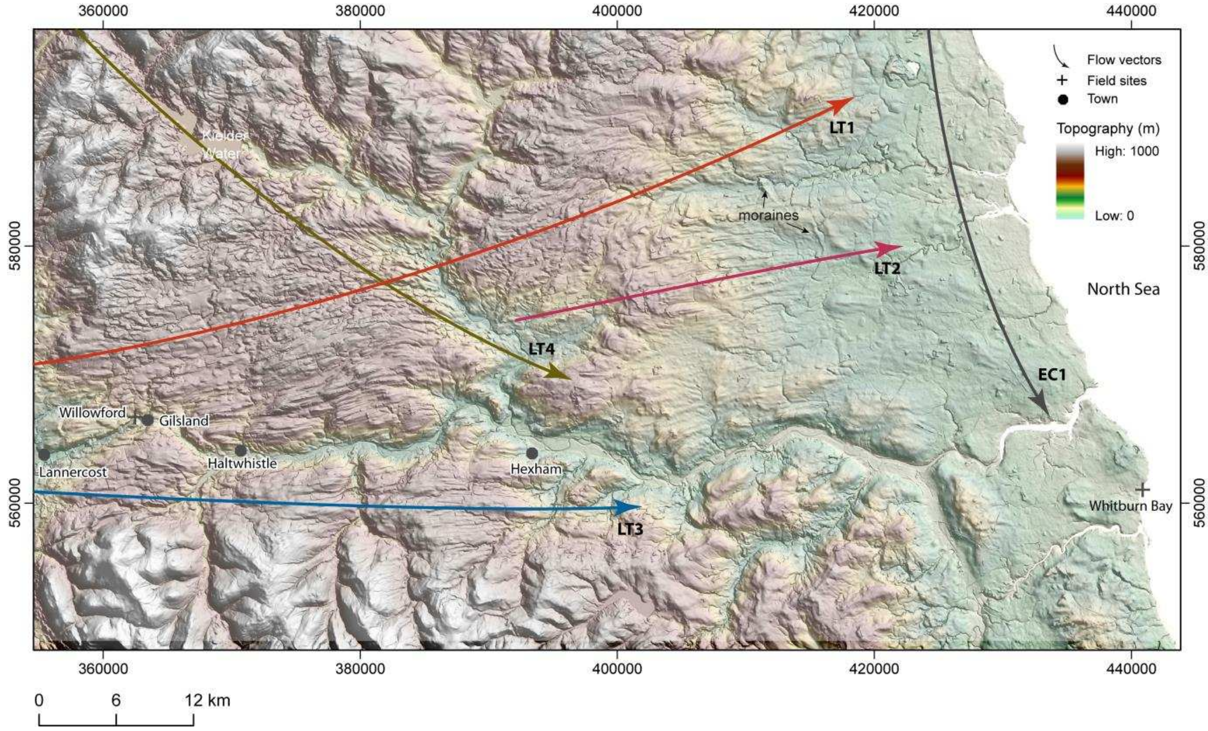
Flow vectors
Field sites
Town
Topography (m)
High: 1000
Low: 0
320000
340000
360000
560000
540000
LT7
LT5
LT3
Plumpe Farm
Annan
Gretna
Willowford
Lannercost
GE
Brampton
Brampton
Kame
Belt
Ribbed
moraine
Gr
HM
HB
WH(F)
MN
Carlisle
Broomhills
Carleton
BW
Brisco
Thursby
Dalston
H-St. C Delta
Wigton
H-St. C
LT6
Pennine escarpment
Baronwood
Aspatria
LT5
Low Oak
Lazonby
Penrith
Sandstone
Ridge
0
5
10 km

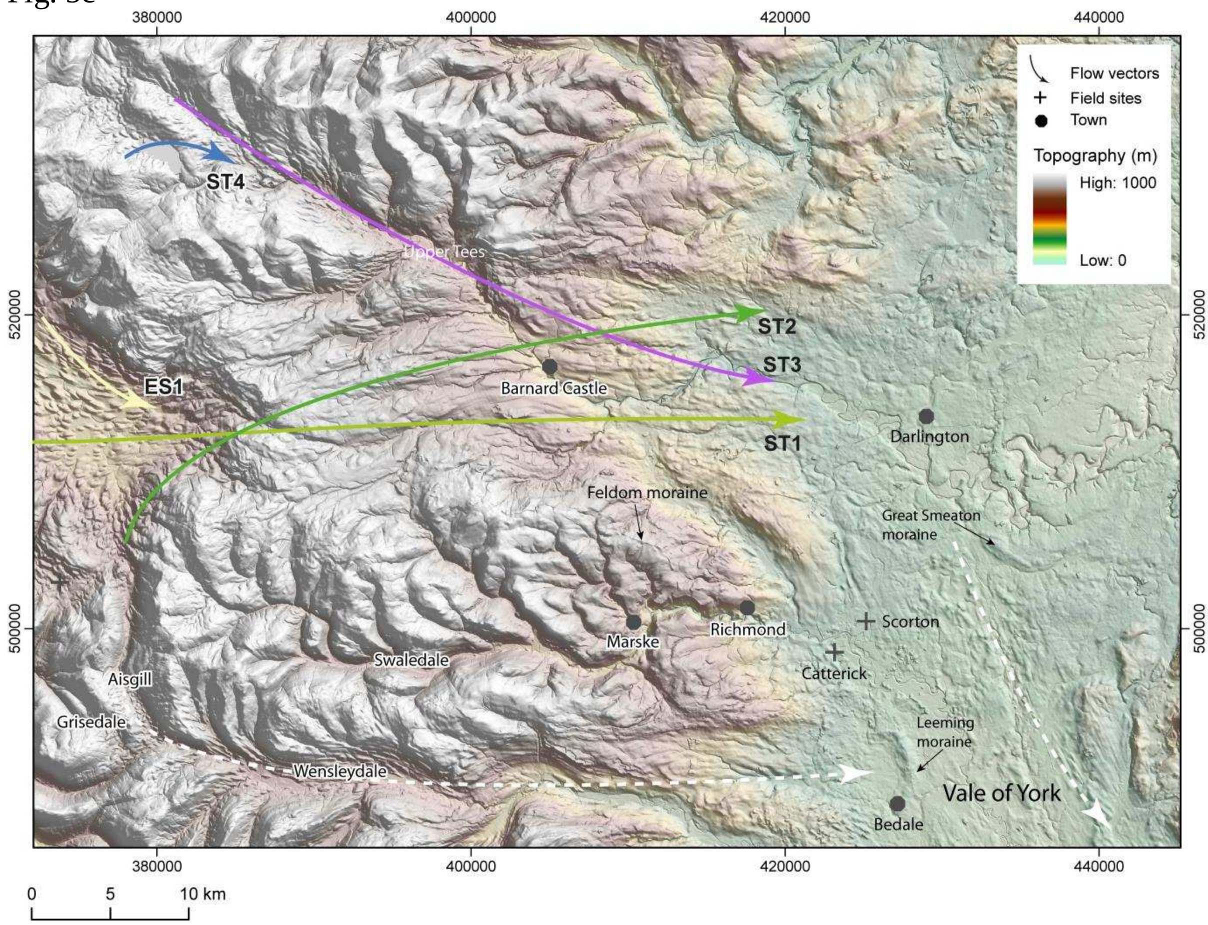
Fig. 5a



340000
360000
380000
540000
520000
LT5
Dalston
BW
Carleton
Brisco
Thursby
LT6
Baronwood
Low Oak
Lazonby
Penrith Sandstone Ridge
Pennine escarpment
Penrith
LT7
Cow Green Reservoir
ST3
ST4
A-I-W
ES1
ST1
ST2
Scandal Beck
Flow vectors
Field sites
Town
Topography (m)
High: 1000
Low: 0
0
5
10 km

Fig. 5b


Flow vectors
Field sites
Town
Topography (m)
High: 1000
Low: 0
LT1
LT2
LT3
LT4
EC1
moraines
North Sea
Willowford
Gilsland
Lannercost
Haltwhistle
Hexham
Whitburn Bay
360000
380000
400000
420000
440000
580000
560000
0
6
12 km
Fig. 5c


Flow vectors
Field sites
Town
Topography (m)
High: 1000
Low: 0
ST4
Upper Tees
ST2
ST3
ST1
ES1
Barnard Castle
Darlington
Feldom moraine
Great Smeaton moraine
Marske
Richmond
Scorton
Catterick
Swaledale
Aisgill
Grisedale
Wensleydale
Leeming moraine
Vale of York
Bedale
380000
400000
420000
440000
520000
500000
0
5
10 km
Fig. 5d

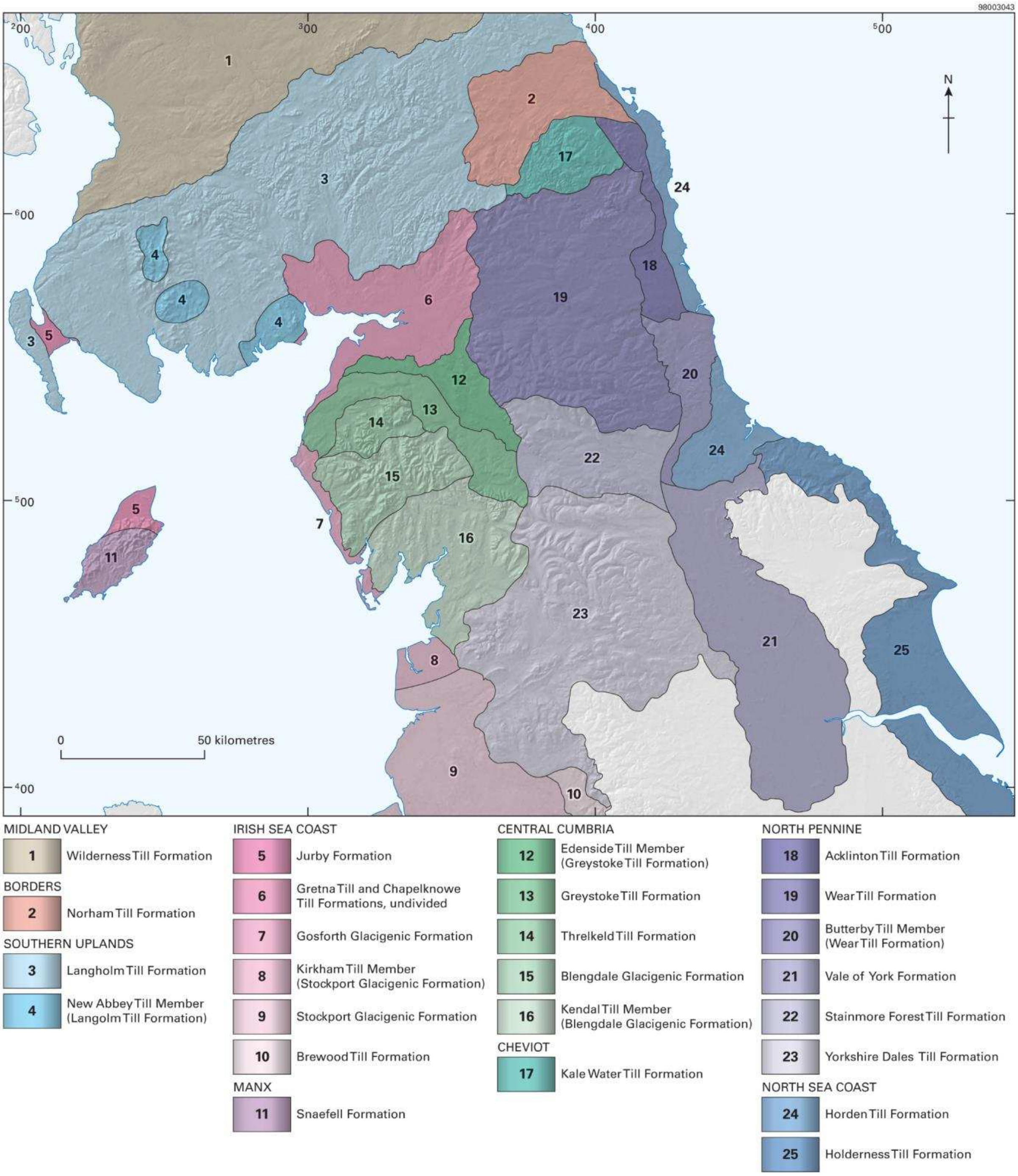

98003043
200
300
400
500
600
500
400
N
0
50 kilometres
MIDLAND VALLEY
1 Wilderness Till Formation
BORDERS
2 Norham Till Formation
SOUTHERN UPLANDS
3 Langholm Till Formation
4 New Abbey Till Member (Langolm Till Formation)
IRISH SEA COAST
5 Jurby Formation
6 Gretna Till and Chapelknowe Till Formations, undivided
7 Gosforth Glacigenic Formation
8 Kirkham Till Member (Stockport Glacigenic Formation)
9 Stockport Glacigenic Formation
10 Brewood Till Formation
MANX
11 Snaefell Formation
CENTRAL CUMBRIA
12 Edenside Till Member (Greystoke Till Formation)
13 Greystoke Till Formation
14 Threlkeld Till Formation
15 Blengdale Glacigenic Formation
16 Kendal Till Member (Blengdale Glacigenic Formation)
CHEVIOT
17 Kale Water Till Formation
NORTH PENNINE
18 Acklinton Till Formation
19 Wear Till Formation
20 Butterby Till Member (Wear Till Formation)
21 Vale of York Formation
22 Stainmore Forest Till Formation
23 Yorkshire Dales Till Formation
NORTH SEA COAST
24 Horden Till Formation
25 Holderness Till Formation

Fig. 6

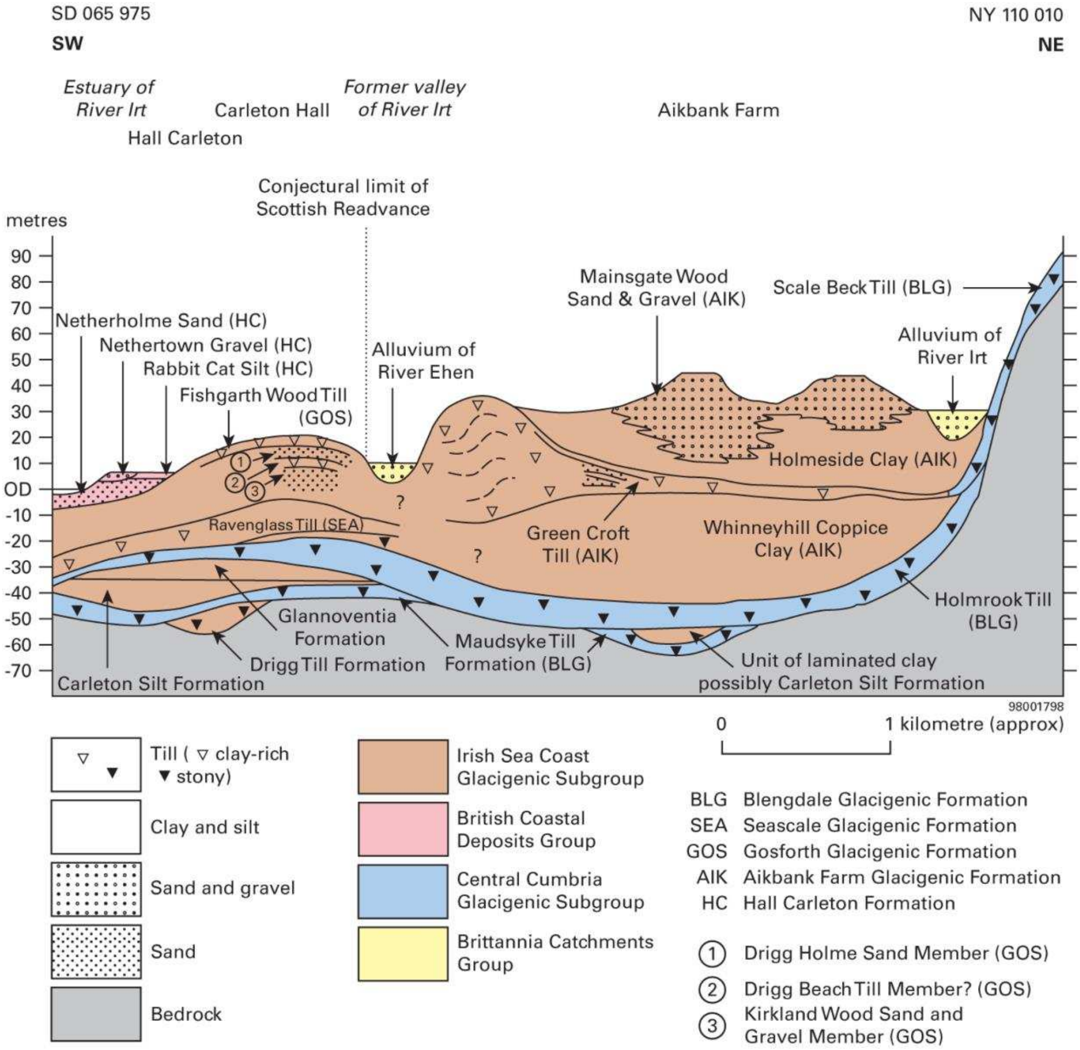

SD 065 975
SW
NY 110 010
NE
Estuary of River Irt
Carleton Hall
Hall Carleton
Former valley of River Irt
Aikbank Farm
Conjectural limit of Scottish Readvance
metres
90
80
70
60
50
40
30
20
10
OD
-10
-20
-30
-40
-50
-60
-70
Netherholme Sand (HC)
Nethertown Gravel (HC)
Rabbit Cat Silt (HC)
Fishgarth Wood Till (GOS)
Alluvium of River Ehen
Mainsgate Wood Sand & Gravel (AIK)
Scale Beck Till (BLG)
Alluvium of River Irt
Holmeside Clay (AIK)
Ravenglass Till (SEA)
Green Croft Till (AIK)
Whinneyhill Coppice Clay (AIK)
Holmrook Till (BLG)
Glannoventia Formation
Maudsyke Till Formation (BLG)
Drigg Till Formation
Carleton Silt Formation
Unit of laminated clay possibly Carleton Silt Formation
98001798
0
1 kilometre (approx)
Till (▽ clay-rich ▼ stony)
Clay and silt
Sand and gravel
Sand
Bedrock
Irish Sea Coast Glacigenic Subgroup
British Coastal Deposits Group
Central Cumbria Glacigenic Subgroup
Brittannia Catchments Group
BLG Blengdale Glacigenic Formation
SEA Seascale Glacigenic Formation
GOS Gosforth Glacigenic Formation
AIK Aikbank Farm Glacigenic Formation
HC Hall Carleton Formation
1 Drigg Holme Sand Member (GOS)
2 Drigg Beach Till Member? (GOS)
3 Kirkland Wood Sand and Gravel Member (GOS)

Fig. 7a

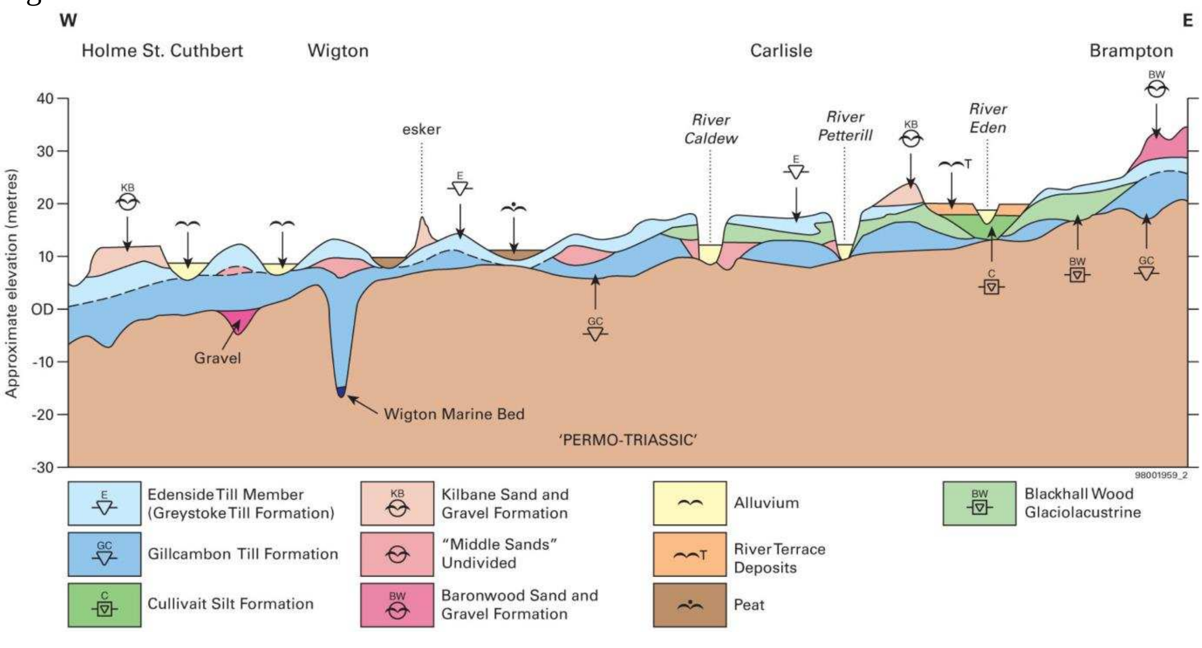

W
E
Holme St. Cuthbert
Wigton
Carlisle
Brampton
Approximate elevation (metres)
40
30
20
10
OD
-10
-20
-30
esker
River Caldew
River Petterill
River Eden
Gravel
Wigton Marine Bed
'PERMO-TRIASSIC'
98001959_2
Edenside Till Member (Greystoke Till Formation)
Gillcambon Till Formation
Cullivait Silt Formation
Kilbane Sand and Gravel Formation
"Middle Sands" Undivided
Baronwood Sand and Gravel Formation
Alluvium
River Terrace Deposits
Peat
Blackhall Wood Glaciolacustrine

Fig. 7b

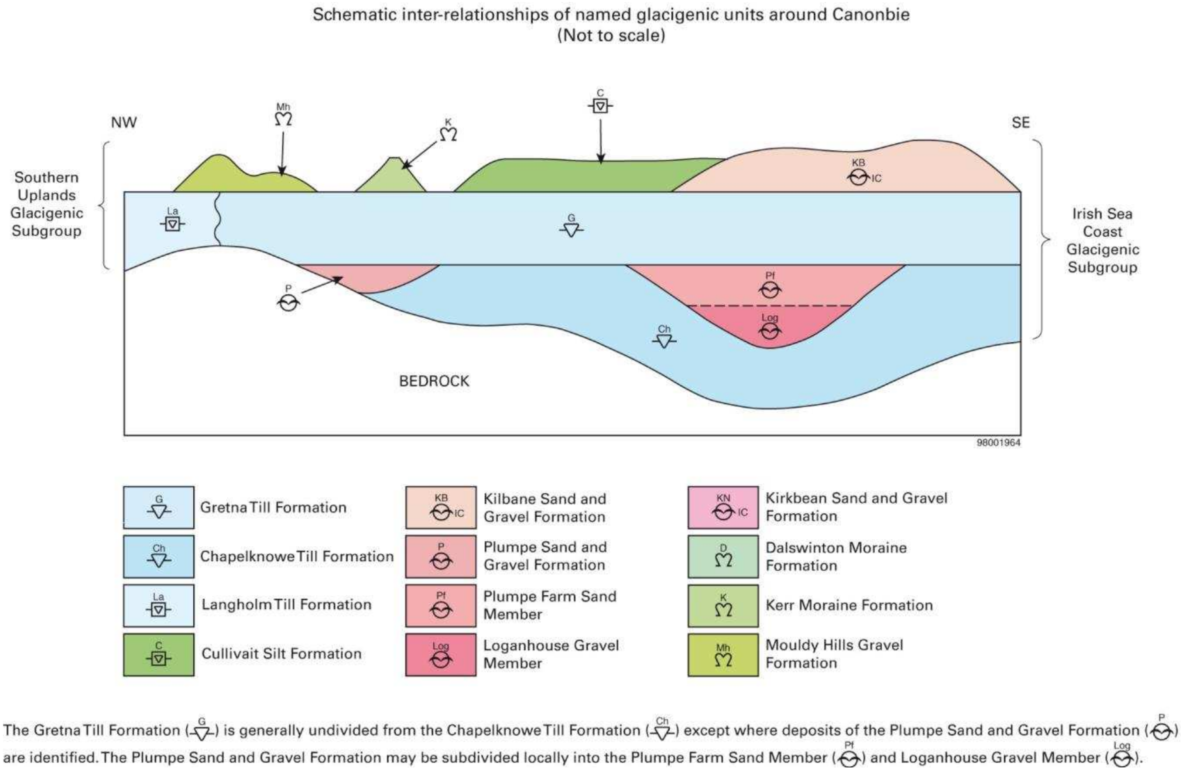

Schematic inter-relationships of named glacigenic units around Canonbie
(Not to scale)
NW
SE
Southern Uplands Glacigenic Subgroup
Irish Sea Coast Glacigenic Subgroup
BEDROCK
98001964
Gretna Till Formation
Chapelknowe Till Formation
Langholm Till Formation
Cullivait Silt Formation
Kilbane Sand and Gravel Formation
Plumpe Sand and Gravel Formation
Plumpe Farm Sand Member
Loganhouse Gravel Member
Kirkbean Sand and Gravel Formation
Dalswinton Moraine Formation
Kerr Moraine Formation
Mouldy Hills Gravel Formation
The Gretna Till Formation (G) is generally undivided from the Chapelknowe Till Formation (Ch) except where deposits of the Plumpe Sand and Gravel Formation (P) are identified. The Plumpe Sand and Gravel Formation may be subdivided locally into the Plumpe Farm Sand Member (Pf) and Loganhouse Gravel Member (Log).

Fig. 7c

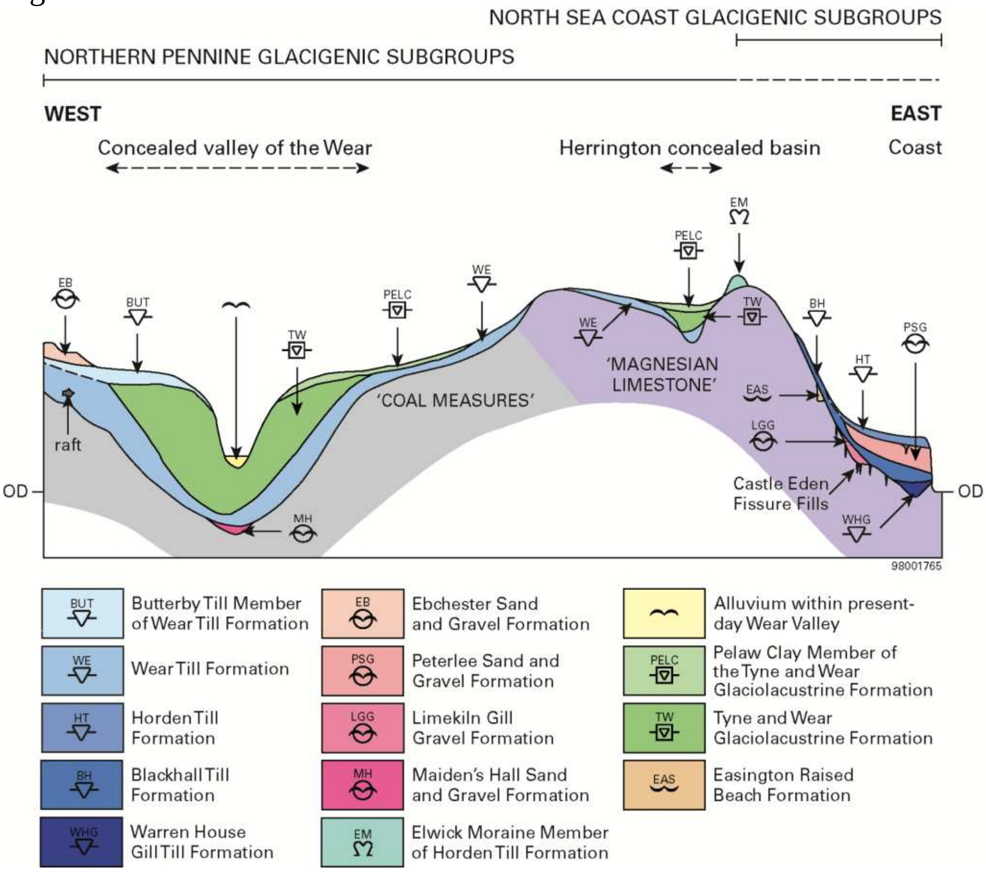

NORTH SEA COAST GLACIGENIC SUBGROUPS
NORTHERN PENNINE GLACIGENIC SUBGROUPS
WEST
EAST
Concealed valley of the Wear
Herrington concealed basin
Coast
'COAL MEASURES'
'MAGNESIAN LIMESTONE'
raft
Castle Eden Fissure Fills
OD
98001765
Butterby Till Member of Wear Till Formation
Wear Till Formation
Horden Till Formation
Blackhall Till Formation
Warren House Gill Till Formation
Ebchester Sand and Gravel Formation
Peterlee Sand and Gravel Formation
Limekiln Gill Gravel Formation
Maiden's Hall Sand and Gravel Formation
Elwick Moraine Member of Horden Till Formation
Alluvium within present-day Wear Valley
Pelaw Clay Member of the Tyne and Wear Glaciolacustrine Formation
Tyne and Wear Glaciolacustrine Formation
Easington Raised Beach Formation

Fig. 7d

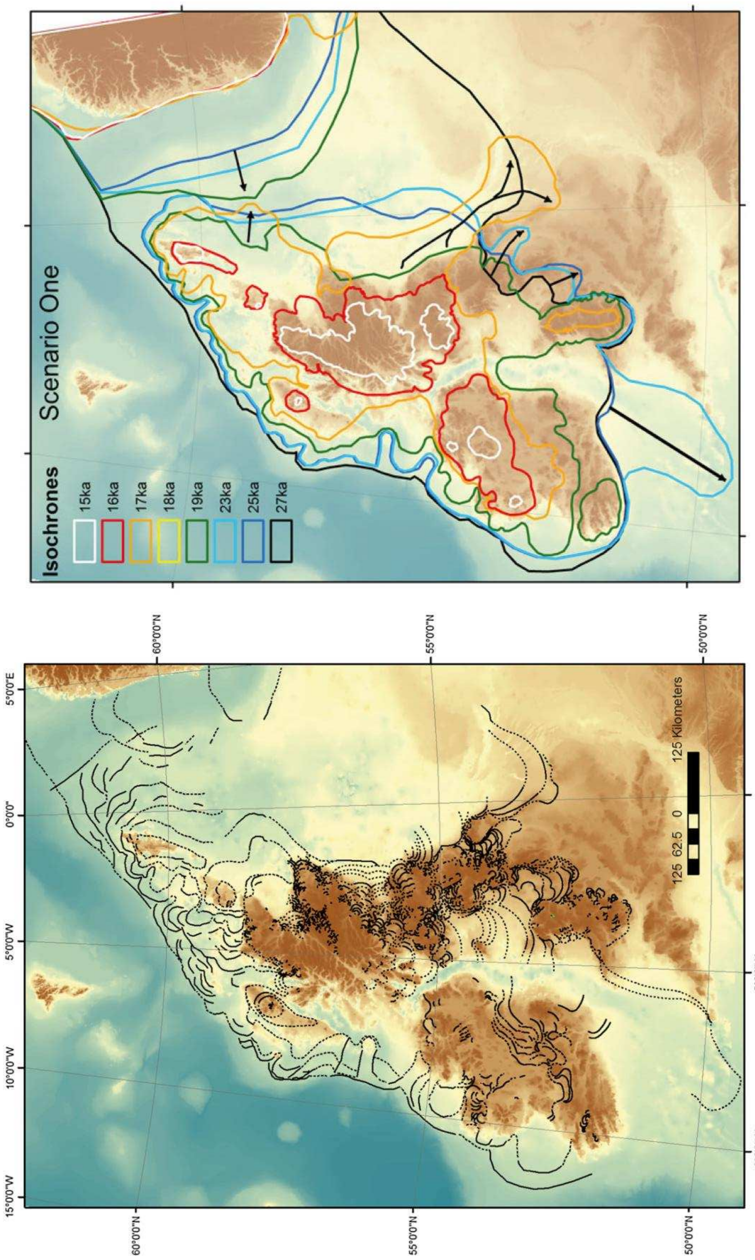
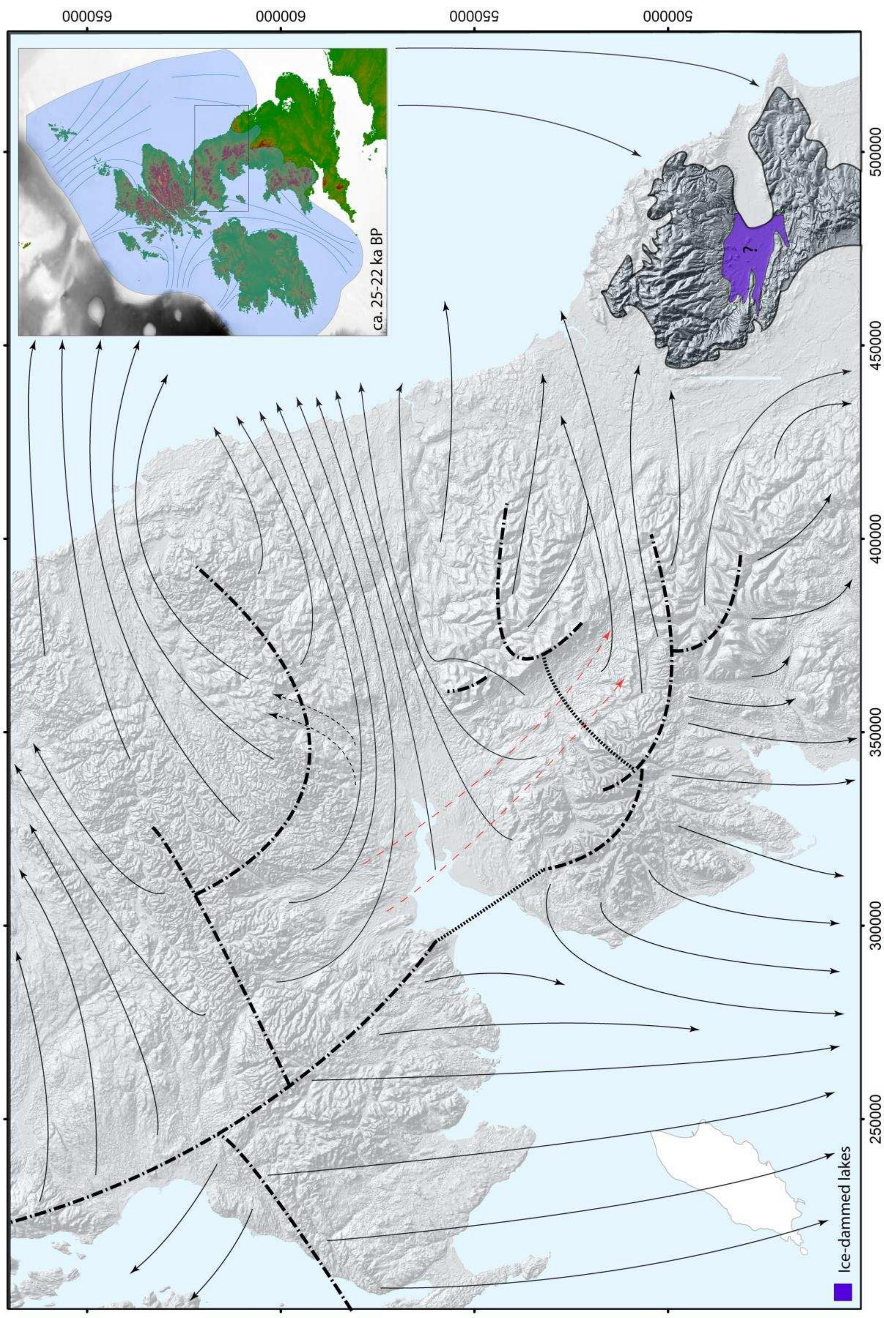
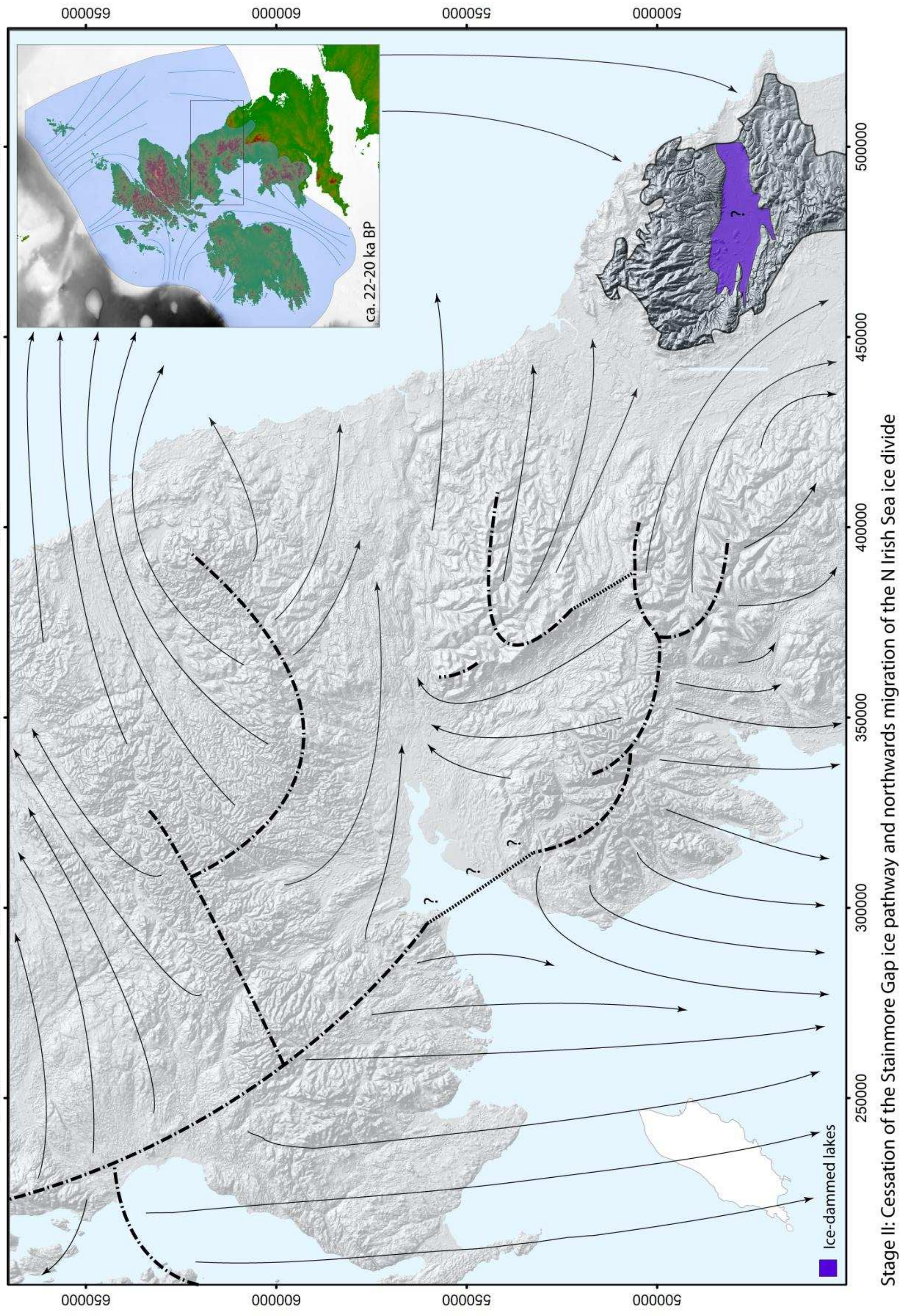

Scenario One
Isochrones
15ka
16ka
17ka
18ka
19ka
23ka
25ka
27ka
60°0'0"N
55°0'0"N
50°0'0"N
5°0'0"E
0°0'0"
5°0'0"W
10°0'0"W
15°0'0"W
125 Kilometers
125 62.5 0

Fig. 8



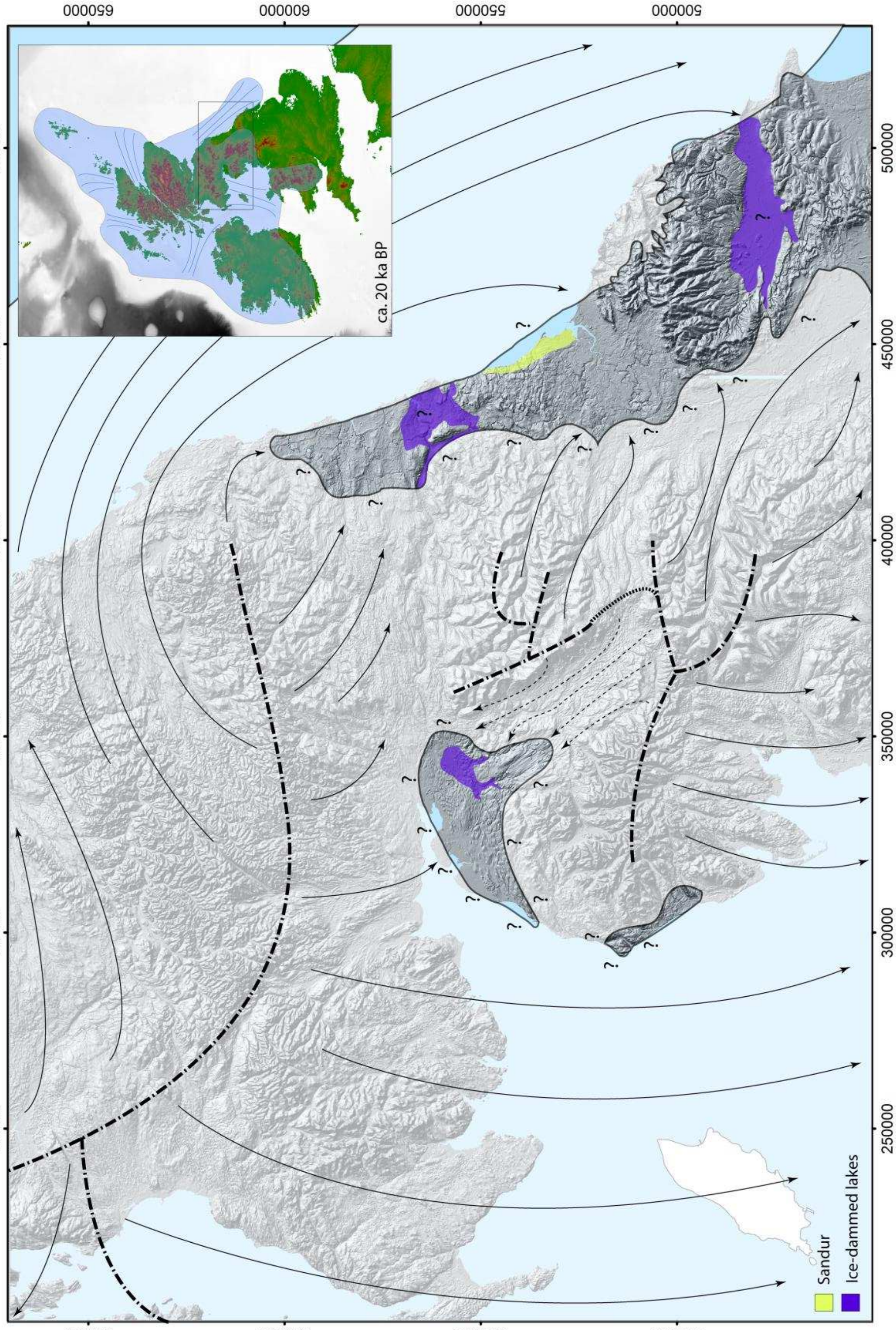
Stage I: Eastwards ice flow through prominent topographic corridors of the north Pennines

Fig. 9



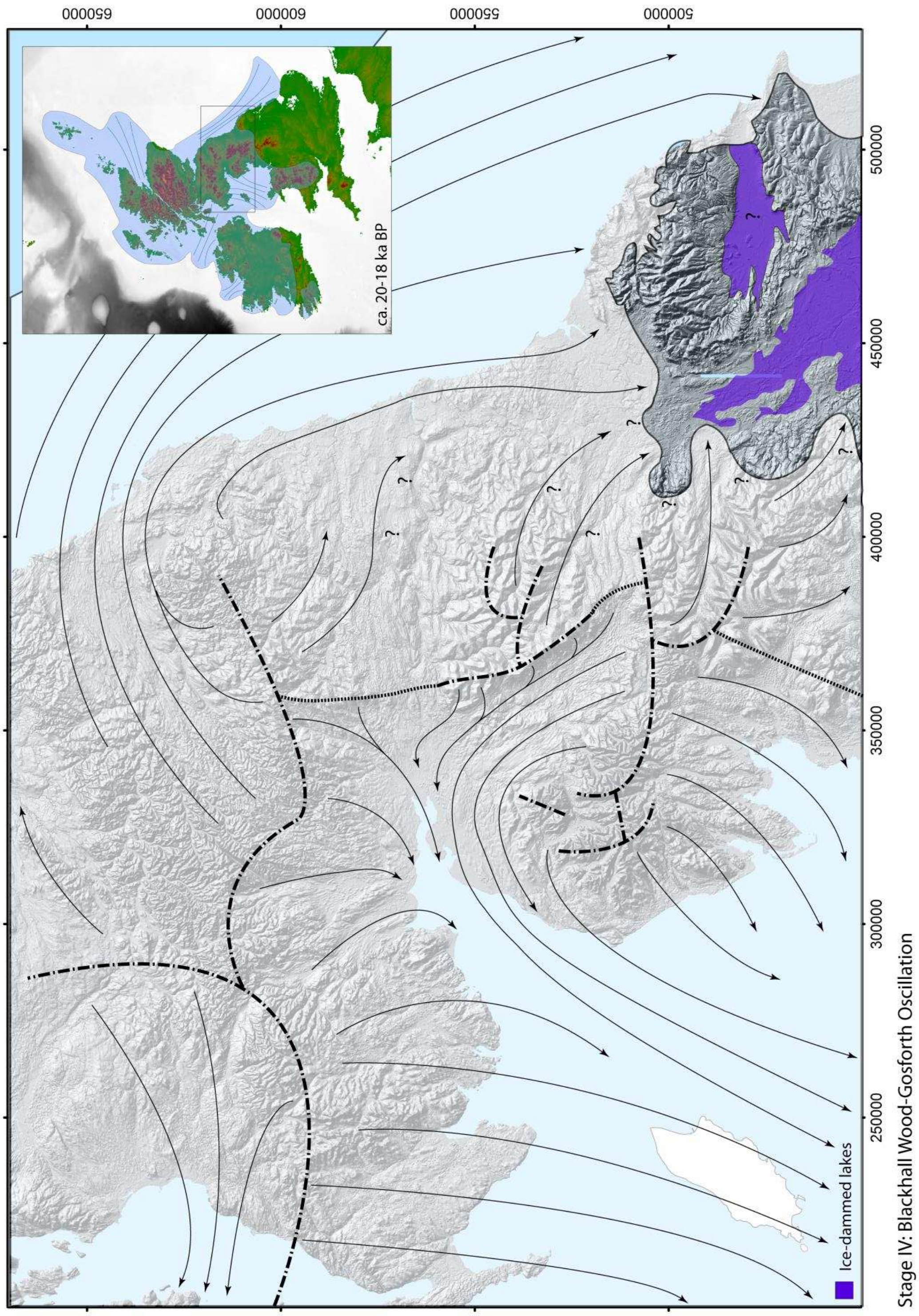
Stage II: Cessation of the Stainmore Gap ice pathway and northwards migration of the N Irish Sea ice divide
Ice-dammed lakes
ca. 22-20 ka BP
650000
600000
550000
500000
500000
450000
400000
350000
300000
250000

Fig. 10



Stage IIIa: Ice-free enclaves in NW Cumbria and NE England, and the development of ice-dammed lakes along the margin of the Irish Sea Ice Stream

Fig. 11



Stage IV: Blackhall Wood-Gosforth Oscillation
Ice-dammed lakes
ca. 20-18 ka BP
500000
550000
600000
650000
250000
300000
350000
400000
450000
500000

Fig. 12

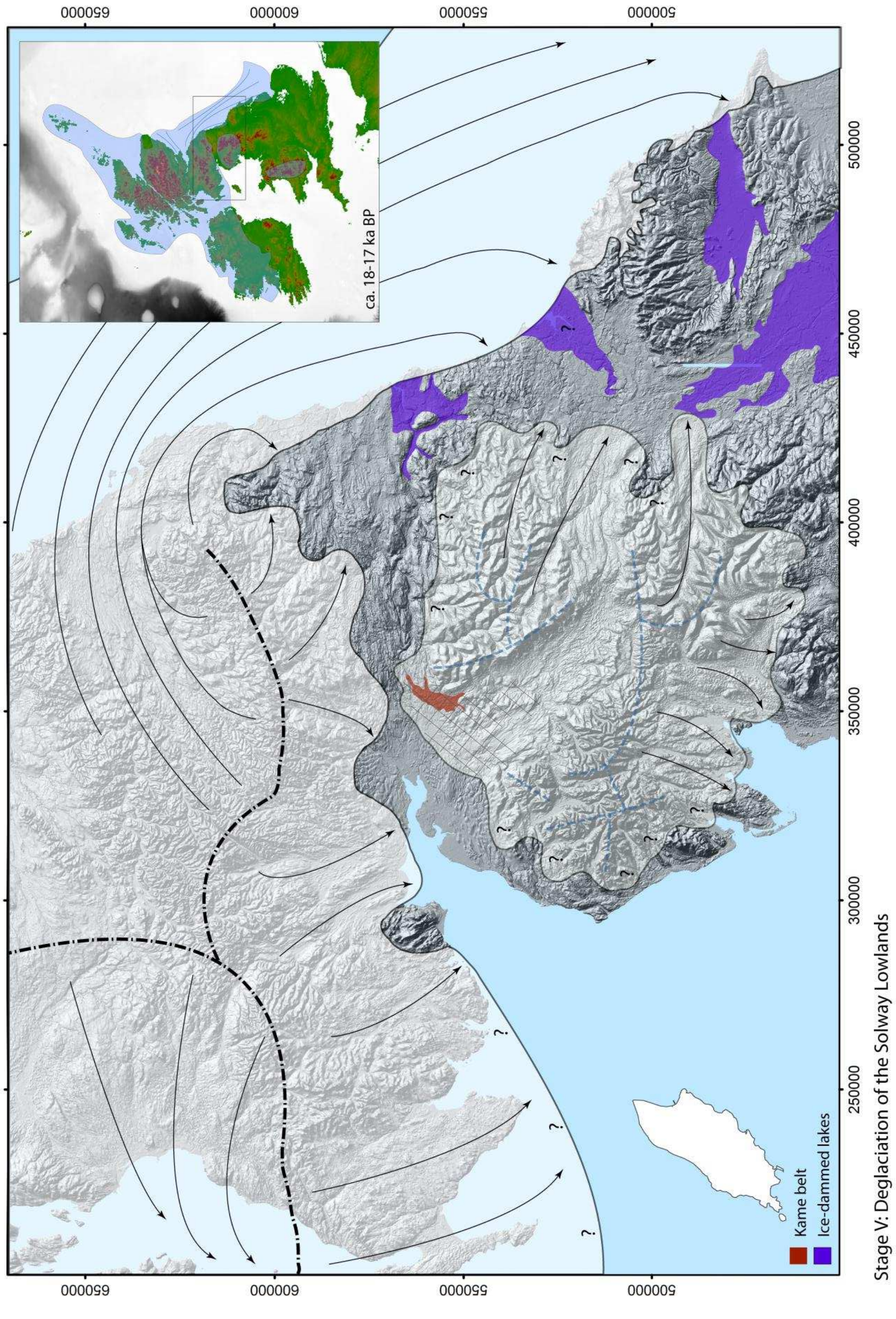

ca. 18-17 ka BP
Kame belt
Ice-dammed lakes
Stage V: Deglaciation of the Solway Lowlands
650000
600000
550000
500000
250000
300000
350000
400000
450000
500000

Fig. 13

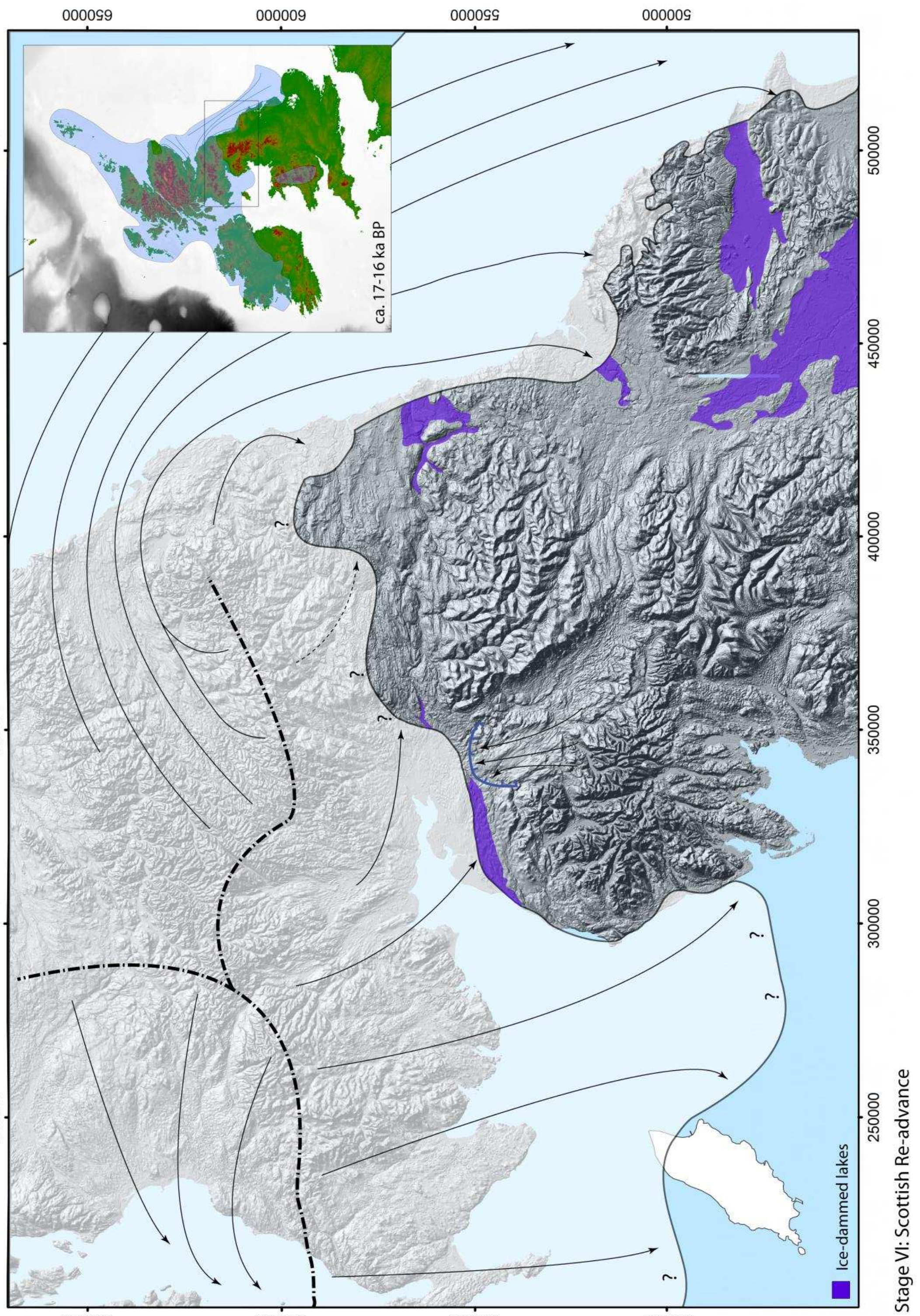

Stage VI: Scottish Re-advance
Ice-dammed lakes
ca. 17-16 ka BP
650000
600000
550000
500000
500000
450000
400000
350000
300000
250000

Fig. 14

<table>
<tr><th rowspan="2">Stage</th><th rowspan="2">Event</th><th colspan="4">Lithostratigraphic Formation</th><th rowspan="2">Key Geomorphic Features & Events</th><th rowspan="2">Flow-phases</th><th rowspan="2" colspan="2">Regional Correlation</th><th rowspan="2" colspan="2">Date (cal. ka BP)</th></tr>
<tr><th>Cumbria</th><th>Dumfries-shire</th><th>Tyne Gap</th><th>Durham Coast</th></tr>
<tr><td>V1</td><td>Scottish Re-advance</td><td colspan="2">Plumpe Bridge Till M [Gretna Till F]; Kilblane S & G F; Cullivait Silt F</td><td rowspan="2">Ebchester S & G F</td><td rowspan="3">Teesside Clay F Horden Till F</td><td>Holme St. Cuthbert delta; Glacial-Lake Wigton; Glacial Lake Carlisle (?); Thursby and Sowerby Wood eskers; Gretna sands and gravels</td><td>SF1; LT6; (LT7)</td><td colspan="2">St Bees & Annaside-Gutterby moraines; Jurby moraines on the IOM (?), Killard Point Stadial (?); North Sea Lobe surging southwards & depositing the Withernsea Till; Glacial-Lake Humber.</td><td>~16.8</td><td rowspan="6">Main Late Devensian</td></tr>
<tr><td>V (IIIb)</td><td>Deglaciation</td><td colspan="2">Plumpe S & GF; Great Easby Clay F</td><td>Glacial-Lake Carlisle; Wiza Beck mwc; Brampton kame belt; Pennine escarpment mwc</td><td></td><td colspan="2">Glacial-Lake Pickering & Glacial-Lake Humber, & the formation of Glacial-Lake Tees & Glacial-Lake Wear; Orrisdale moraines on the IOM (?)</td><td></td></tr>
<tr><td>IV</td><td>Blackhall Wood Re-advance</td><td>Greystoke Till F</td><td rowspan="2">Gretna Till F and Chapelknowe Till F undivided</td><td>Butterby Till F</td><td>Ice-flow tributary of the Irish Sea Ice Stream</td><td>LT5; EC1</td><td colspan="2">Gosforth Oscillation; Clogher Head Stadial (?); SW ice-flow over the Isle of Man; North Sea Lobe advances and deposits the Horden Till (& Skipsea Till?); deglaciation of the Vale of York.</td><td>~21 to 19.5</td></tr>
<tr><td>IIIa</td><td>Partial deglaciation</td><td>Blackhall Wood Clay F</td><td>Tyne & Wear Glaciolacustrine F</td><td>Peterlee S&G F</td><td>Glacial-Lake Blackhall Wood</td><td>(LT4)</td><td rowspan="2">Rapid retreat of Irish Sea Ice Stream from Celtic Basin. Initial development of Glacial Lake Wear (?)</td><td>Ice free enclaves in eastern England and the formation of a large drainage network in the Tyne Valley and sandur at Peterlee</td><td></td></tr>
<tr><td>II</td><td rowspan="2">Main Glaciation</td><td rowspan="2">Gillcambon Till F*</td><td rowspan="2">Chapelknowe Till F</td><td>Acklinton Till F</td><td rowspan="2">Blackhall Till F</td><td>SE flow down the N. Tyne Valley</td><td>LT4; ST3-4; ice flow down the Vale of Eden; (EC1)</td><td rowspan="2">Escrick moraine, Vale of York (Skipsea Till?); Glacial Lake Humber (?); Glacial-Lake Pickering (?).</td><td rowspan="2">~23-29</td></tr>
<tr><td>I</td><td>Wear Till F</td><td>Eastward ice-flow through the Tyne Gap (as a topographic ice stream) and Stainmore Gap.</td><td>LT1-3; ST1-2; ES1</td><td>Maximum extent of Irish Sea Ice Stream and Irish Ice Sheet; Shelf-edge glaciation; confluent BIIS and FIS in the North Sea Basin.</td></tr>
<tr><td></td><td>Early Devensian interstadials</td><td>Scandal Beck Peat B; Mosedale Beck Peat B</td><td></td><td>Maiden’s Hall S & G F</td><td></td><td></td><td></td><td></td><td></td><td colspan="2">MIS 5a-d</td></tr>
<tr><td></td><td>Ipswichian Interglacial</td><td>Wigton Marine Bed; Troutbeck Palaeosol B</td><td></td><td colspan="2"></td><td></td><td></td><td></td><td></td><td colspan="2">Ipswichian MIS 5e</td></tr>
<tr><td></td><td>Pre-Devensian Advance</td><td>Thornsgill Till F (Troutbeck)</td><td></td><td colspan="2">Warren House Till F</td><td></td><td></td><td></td><td></td><td colspan="2">Pre-Ipswichian</td></tr>
</table>

Table 1

Region	Formation	Member	Facies/Sediments	Chrono-stratigraphy	Regional correlatives	Reference
C & W Cumbria	Gosforth Glacigenic F	Fishgarth Wood Till M	Thin, red tills with some Scottish erratics capping sequences in West Cumbria.	MIS2	Gretna Till F	Nirex, 1997; Merritt & Auton, 2000.
		How Man Till M				
		Drigg Beach Till M	Thin, red, clayey till with some Scottish erratics.		Gretna or Greystoke Till F	
		St Bees Till M	Grey, clayey 'Irish Sea' till with shell fragments			
	Aikbank Farm Glacigenic F	Greencroft Till M	Thin, grey LD till within glaciolacustrine sequence.	MIS2	Greystoke Till F	Merritt & Auton, 2000.
	Seascale Glacigenic F	Ravenglass Till M	Red till with Cumbrian Coalfield and some Scottish erratics.	MIS2	Greystoke Till F	Nirex, 1997; Merritt & Auton, 2000.
		Lowca Till M				
Solway Lowlands & VofE/Dumfries-shire	Gretna Till F	Plumpe Bridge Till M	Red, thin till sheet, which caps the sequence in Dumfries-shire & Langholm and pinches out in the Solway Lowlands. Dominated by greywacke & granite erratics of the Southern Uplands.	MIS2		McMillan et al., 2011. In press; Stone et al. 2010.
Solway Lowlands & VofE	Greystoke Till F	Edenside Till M	Red-brown till, generally 5-20 m thick that forms the drumlinoid bedforms of the Vale of Eden. Mixed provenance of local, LD & Scottish erratics.	MIS2		Eastwood et al., 1968; Huddart, 1971; McMillan et al., 2011.
	Gillcambon Till F		Lower red-brown to grey-brown, compact, & in parts fissile, till with a mixed provenance of LD, Scottish & local erratics. Outcrops in the Vale of Eden.	MIS2	Wear Till F, Tyne Gap.	Goodchild, 1875; Eastwood et al., 1968; Huddart 1971.
	Thornsgill Till F		Deeply weathered till with LD erratics.	Mid-Pleistocene		Boardman, 2002.
Dumfries-shire-Langholm	Chapelknowe Till F		Lower, red till dominated by Southern Upland erratics (including Criffel & Dalbeattie granites). Outcrops…	MIS2	Gillcambon Till F	McMillan et al., 2011; McMillan et al., 2011, In Press
Tyne Gap	Butterby Till F		Thin grey till with Tyne Gap and Pennine erratics around Durham.	MIS2	Horden Till F in part	Francis, 1970; Stone et al., 2010.
	Acklinton Till F		Upper grey till with Northumberland and Cheviot erratic.	MIS2	Horden Till F in part	McMillan et al., 2011
	Wear Till F		Red to grey-brown till with LD & Scottish erratics, but increasingly dominated by local Carboniferous lithologies towards the east. The till outcrops throughout the Tyne Gap & valley & ranges in thickness from veneers to ~90 m in in-filled valleys.	MIS2	Blackhall Till F, Durham coast.	Francis, 1970; Hughes, 1998; Livingstone et al. 2010c; Stone et al., 2010; Yorke et al., submitted.
Pennines & Stainmore Gap	Yorkshire Dales Till F		Massive, compact till, with a dominance of Pennine Carboniferous erratics, extensive throughout western Pennines.	MIS2		Avelin & Hughes, 1888; Dakyns et al., 1890, 1891; McMillan et al., 2011.

	Stainmore Forest Till F		Compact till with local Carboniferous & LD erratics. Extensive outcrops in the Stainmore Gap.	MIS2	Vale of York Till F	Burgess & Holliday, 1979; Mills & Hull, 1976; McMillan et al., 2011.
Durham Coast	Horden Till F		Overlies & interdigitates with Peterlee S&G F. Brown sandy till with Cheviot and near-coast North Sea erratics. Extensive outcrops on Durham coast.	MIS2	Skipsea Member (Yorkshire), which is dated to 21 cal. ka BP using radiocarbon on mosses. Bolders Bank Formation (North Sea).	Francis, 1970; Lunn, 1995; Davies et al., 2009a; Davies et al., submitted; Stone et al., 2010.
	Blackhall Till F		Thick, grey-brown till, Permian & Carboniferous lithologies, extensive outcrops on Durham coast.	MIS2	Wear Till, Tyne Gap. Bolders Bank Formation (North Sea).	Davies et al., 2009a; Francis, 1970, 1972; Davies et al., submitted.
	Warren House F	Ash Gill M	Glaciomarine to subglacial diamicton, sparse North Sea & Scandinavian erratics.	Mid-Pleistocene		Davies et al., in press; Davies et al., submitted.

Table 2