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Glaciolacustrine sedimentation in the Solway Lowlands (Cumbria, UK): evidence for a major glacial oscillation during Late Devensian deglaciation.

STEPHEN J. LIVINGSTONE, COLM Ó COFAIGH, DAVID J.A. EVANS and ADRIAN PALMER

This paper is a sedimentological investigation of Late Devensian glacial deposits from the Solway Lowlands, NE England, in the central sector of the last British-Irish Ice Sheet. In this region laminated glaciolacustrine sediments occur, sandwiched between diamicton lithofacies interpreted as subglacial tills. At one location these laminated sediments have been interpreted as varves, and they indicate the former presence of a proglacial lake. Correlation of these varves with other laminated sediment indicate that that the glacial lake was at least 140 km^2 in area and probably much larger. Extensive beds of sand, silt and gravel throughout the Solway Basin associated with the lake demonstrate ice free conditions over a large area. Based on the number of varves present, the lake was in existence for at least 261 years. The stratigraphic sequence of varves bracketed by tills implies a major glacial oscillation in this region prior to the Scottish re-advance (16.8 cal ka BP). This oscillation is tentatively correlated with the Gosforth oscillation at *ca*. 19.5 cal ka BP. Subsequent overriding of these glaciolacustrine sediments during a westwards moving readvance demonstrates rapid ice loss and then gain within the Solway Lowlands from ice dispersal centres in the Lake District, Pennines and Southern Uplands. It is speculated that the influence of this and other lakes along the north-eastern edge of the Irish Sea Basin would have had an impact on ice sheet dynamics.

Glacial lakes; varves; re-advance; Irish Sea Basin; ice sheet dynamics

Stephen J. Livingstone[1], Colm Ó Cofaigh[1], David J.A. Evans[1] and Adrian Palmer[2]

s.j.livingstone@dur.ac.uk; **colm.o'cofaigh@durham.ac.uk**; **d.j.a.evans@durham.ac.uk**; **a.palmer@rhul.ac.uk**

[1] *Department of Geography, Durham University, South Road, Durham, DH1 3LE*

[2] *Department of Geography, Royal Holloway, University of London, Egham, Surrey, TW20 0EX*

Introduction

Much recent evidence suggests that the Late Devensian (Marine Isotope Stage 2) British-Irish Ice Sheet (BIIS) was characterised by multiple ice dispersal centres drained by ice streams and fast-flowing outlet glaciers (e.g. Merritt *et al*. 1995; Boulton & Hagdorn 2006; Bradwell *et al*. 2007; Ó Cofaigh & Evans 2007; Roberts *et al*. 2007; Hubbard *et al*. 2009; van Landeghem *et al*. 2009). Perturbations to the ice-sheet system from both external and internal forcings are therefore envisaged to have been highly sensitive, with enhanced ice flow conditions rapidly transmitting changes far into the ice-sheet interior leading to relatively rapid, non-linear and complex re-configurations of the BIIS (cf. Bamber *et al*. 2007; Evans *et al*. 2009; Hubbard *et al*. 2009). This is indicated by the geomorphic and sedimentological evidence, which advocates a dynamic ice sheet characterised by multiple ice flow switches (e.g. Livingstone *et al*. 2008; Greenwood & Clark 2008; Evans *et al*. 2009) and an oscillatory margin (Eyles & McCabe 1989; McCabe 1996; Evans & Ó Cofaigh 2003; Thomas *et al*. 2004; Thomas & Chiverrell 2007; McCabe *et al*. 2005, 2007). This sensitivity has, in some instances, been correlated to abrupt millennial-scale climate changes occurring in the North Atlantic Ocean (e.g. McCabe & Clark 1998; Scourse *et al*. 2000), with recent modelling studies suggesting a dynamic BIIS modulated and phase lagged by external climatic forcings (Hubbard, *et al*. 2009).

The Irish-Celtic Sea was thought to contain an extensive Irish Sea Ice Stream that advanced to the Scilly Isles at some point during the Main Late Devensian glaciation (Scourse & Furze 2001; Hiemstra *et al*. 2006). The Irish Sea Ice Stream is envisaged to have been highly responsive during deglaciation with evidence for multiple oscillations of its margin (e.g. Evans & Ó Cofaigh 2003; Thomas *et al*. 2004; Thomas & Chiverrell 2007). Readvances have been identified in northeast Ireland between 18.3 – 17.0 cal ka BP and after 17.0 cal ka BP (McCabe & Clark 1998; McCabe *et al*. 2005, 2007). This latter readvance has been correlated to Heinrich Event 1 as part of an extensive Irish Sea glacial-response to North Atlantic climate change (McCabe & Clark 1998). Less chronologically well constrained re-advances include the Gosforth oscillation at *ca*. 19.5 cal ka BP

(Merritt & Auton 2000) and the Scottish re-advance at *ca*. 16.8 cal ka BP (Trotter 1929, Merritt & Auton 2000) along the Cumbrian coast.

The Solway Lowlands, in the north-east sector of the Irish Sea Basin, (Fig. 1) are situated within a palimpsest glacial landscape characterised by multiple phases of flow and containing evidence of a Scottish re-advance (cf. Livingstone *et al*. 2008). This complexity is partially explained by its location, situated between major ice dispersal centres of the Pennines, Lake District and Southern Uplands, with ice coalescing and subsequently draining through outlets such as the Tyne and Stainmore gaps (Trotter 1929; Hollingworth 1931; Livingstone *et al*. 2008). The Solway Lowlands also acted as a tributary to the Irish Sea Ice Stream at some point during the Main Late Devensian (Roberts *et al*. 2007; Livingstone *et al*. 2008). Stratigraphic and geomorphological evidence suggests that Scottish ice re-advanced into the Solway Lowlands following partial deglaciation, although there is some debate as to the extent of this readvance (Trotter 1929; Trotter & Hollingworth 1932; Huddart 1970, 1971, 1991, 1994).

This geomorphic complexity manifests itself within the stratigraphic record (see Table 1). The oldest glacigenic deposits associated with the Main Late Devensian Glaciation in the Vale of Eden, Solway Lowlands and Dumfries-shire are the Gillcambon and Chapelknowe till formations (Stone *et al*. in press). These lower-most till units have been assigned to the ‘Early Scottish’ advance of ice up the Vale of Eden and across the Stainmore Gap (cf. Huddart & Glasser 2002; Stone *et al*. in press). An upper till in the Vale of Eden (Greystoke till formation: Stone *et al*. in press), separated from the Gillcambon till formation by sporadic sand, silt and clay deposits (‘Middle sands’), is assigned to the ‘Main Glaciation’ (Dimlington stadial) (Hollingworth 1931). Goodchild (1875, 1887) and Huddart (1970) attributed all the deposits to the melting (subglacially) of a single, stagnant ice mass, while Hollingworth (1931) surmised that the ‘Middle’ sands and laminated clays were formed proglacially, thus delimiting a period of partial deglaciation, followed by re-advance. Resolving the uncertainty associated with this glacial re-organisation has major implications for the glacial history in Cumbria in terms of its ice flow phases, ice-sheet behaviour, glacial lake development and correlations between regional ice dynamics and the rest of the BIIS.

In the northern Solway Lowlands and Dumfries-shire a tripartite division of sediments is recognised, with the Chapelknowe and Greystoke till formations overlain by a series of sands and clays (Plumpe Farm sand and gravel formation & Great Easby clay formation) and then capped by a thin (< 5 m) upper red till (Gretna till formation) (Stone *et al.* in press). The Great Easby clay formation was deposited during formation of Glacial Lake Carlisle during deglaciation of the region; while the Gretna till formation corresponds to a late-stage re-advance of Scottish ice into the Solway Lowlands (e.g. Trotter & Hollingworth 1932). Within the Solway Lowlands the Great Easby clay formation has been extended south of Carlisle (Trotter 1929; Huddart 1970), with the tripartite divisions exposed here thought to correspond to a period of deglaciation followed by re-advance of the Scottish ice. However, the Gretna till formation underlies drumlins associated with the arcuate flow of ice around the northern margin of the Lake District and out into the Irish Sea Basin (Fig. 2).

This study presents new evidence that intervening clay and silt deposits exposed in the northern sector of the Vale of Eden, previously assigned to the main period of deglaciation (Great Easby clay formation) are in fact chronostratigraphic with the earlier phase of glacial re-organisation (Middle sands) and demonstrates a major glacial oscillation separated by a period of glaciolacustrine activity. This suggests that the central sector of the BIIS was not just complex in terms of its ice flow phasing (cf. Livingstone *et al.* 2008), but also that it experienced major ice marginal oscillations leading to the formation of large proglacial lakes.

Study area and glacial geomorphology

The Solway Lowlands (Fig. 1) form a low lying (< 100 m Ordnance Datum) basin bounded by the Solway Firth and Irish Sea to the west and upland terrain to the north, south and east. Two mountain passes, the Vale of Eden and Tyne Gap which run SE-NW and W-E respectively, converge on the Solway Lowlands (Fig. 1). The underlying bedrock geology is Permo-Triassic sandstone and mudstone (Hollingworth 1931).

The area is predominantly composed of Late Devensian deposits that show evidence for extensive glacial activity (Fig. 2), as recorded by a dense coverage of subglacial lineations, meltwater channels, ribbed moraine, hummocky terrain, eskers and glaciofluvial accumulations, primarily below 400 m O. D. (Livingstone *et al*. 2008). The glacial geomorphology has been used to reconstruct a relative chronology of ice flow phases based on cross-cutting relationships (Livingstone *et al*. 2008, Fig. 2). Erratic trains suggest that the earliest phase of flow was southwards from Scotland, up the Vale of Eden and across the Stainmore Gap (‘Early Scottish advance’ of Hollingworth 1931). This phase is not recorded within the geomorphic record, suggesting that the evidence for it had been destroyed by subsequent ice flows. The next phase to influence the Solway Lowlands was characterised by convergent flow, sourced from Lake District and Southern Upland ice dispersal centres, moving eastwards into and through the Tyne Gap (Livingstone *et al*. 2008, Fig. 2). A major flow switch occurred when ice drained westwards into the Irish Sea Ice Basin, probably as a trunk zone feeding into the Irish Sea Ice Stream (Roberts *et al*. 2007). Superimposed on top of this flow-set is a deltaic complex located at Holme St Cuthbert (Huddart, 1970; Huddart and Glasser 2002) demarcating the re-advance of ice across the Solway Firth from Scotland (cf. Salt & Evans 2004). The final ice flow phases are typified by topographically constrained flow lobes extending into the lowland region (Livingstone *et al*. 2008, Fig. 2), which probably occurred prior to, or concurrently with, the Scottish re-advance.

Methods

Field exposures at Blackhall Wood, Maryport and Swarthy Hill (Fig. 1) enabled detailed stratigraphic logging and sedimentological analysis to be carried out. These field sites are located in lineations that are associated with the movement of ice into the Irish Sea Ice Basin (Fig. 2). Texture, sedimentary structure, colour (Munsell colour chart), bed geometry, contacts and inclusions were all measured and logged, from which lithofacies were identified. Lithofacies codes are based upon those of Evans & Benn (2004). Scaled section sketches were drawn at the larger exposures so that the lateral extent of the lithofacies could be assessed. Clast macrofabric analysis of the a-axis orientation and dip, and investigations of clast lithology supplement the geomorphological mapping of ice flow pathways. Diamicton samples for thin-section analysis were collected to provide detailed micromorphological information. Borehole logs (British Geological Survey) provided a less detailed but wider coverage, allowing regional stratigraphic correlations. At Blackhall Wood, overlapping monoliths were collected from a 1.3 m sequence of laminated clay and silt, and a continuous set of thin sections was produced for the sequence. Standard techniques were used for the impregnation and preparation of thin sections of unconsolidated sediments at the Centre for Micromorphology, University of London (Palmer *et al*. 2008a).

Sedimentology and stratigraphy

Blackhall Wood:

Blackhall Wood (NY 386 515) is situated 5 km south of Carlisle on the banks of the River Caldew (Fig. 1). The section is exposed in the north-western edge of a SE-NW trending drumlin (40 m O. D.) that forms part of an arcuate flowset sweeping around the northern edge of the Lake District (Fig. 2). Late Devensian sediments rest on thinly bedded Permo-Triassic sandstones which dip steeply towards the east. The sediment lithofacies and lithofacies associations are depicted diagrammatically in Fig. 3 and described below.

Lithofacies 1 and 2: lowermost diamictons

Description

The basal lithofacies consists of a brownish-black (7.5 YR 3/2) matrix-supported diamicton (LF1) up to 2.0 m thick (Fig. 3, 4a). The diamicton contains a moderate distribution of predominantly sub-angular to sub-rounded clasts many of which are striated. The principle lithologies (Table 2) are Silurian greywacke (31%), Carboniferous limestone (15%), Permo-Triassic sandstone (9%), sandstone (13%) and dolerite (8%), with Southern Upland granite (2.0%) and lava’s of the

Borrowdale Volcanic Group (0.8%) also identifiable. Horizontal (sometimes wavy), millimetre thick, sand partings occur throughout the lithofacies (Fig. 3). A thin section taken near the top of the lithofacies and orientated S-N exhibits lineations, turbate structures and an interconnected network of sub-horizontal and sub-vertical fissures (Fig. 5a). This fissility is also noted macroscopically (Fig. 4a, b). Two fabrics (Fig. 6) indicate a WSW – ENE orientation with S_1 eigenvalues of 0.61 and 0.55.

Overlying this lithofacies is another matrix-supported diamicton up to 2.0 m thick (LF2) characterised by a dull reddish-brown colour (5.0 YR 4/3) and massive structure (Fig. 3, 4b). The boundary between the two lithofacies is indistinct (Fig. 4b) with a slight change in colour, decrease in fissility and an increase in clast content observed (Fig. 3, 4b). A thin section taken 20 cm from the top of the LF indicates a reduction in fissures relative to LF1, although with lineations and turbate structures still present (Fig. 5b). Clasts in LF2 have fewer striae and are rounder in shape than LF1. The erratic content (Table 1) is very similar to that of LF1 with 30% greywacke, 16% Carboniferous limestone, 5% sandstone, 15% dolerite, 6% Permo-Triassic sandstone, 3% lavas from the Borrowdale Volcanic Group and 3% Southern Upland granite. However, clast macro-fabrics (Fig. 6) indicate a change in flow direction, from WSW-ENE in LF1 to SE-NW (S_1 eigenvalues of 0.63 and 0.57) in LF2.

LF1 and LF2: Interpretation

Lithofacies 1 and 2 are interpreted as subglacial traction tills (e.g. Evans *et al*. 2006). Evidence for this includes: (a) stones that are commonly striated and have a far-travelled component to their provenance; (b) fabric orientations which have predominantly high S_1 values (Benn 1995; Evans 2000) and girdle fabrics indicating transport in a deforming medium (Hicock & Fuller 1995); and (c) micro-scale structures of both ductile (skelsepic fabric, turbates) and brittle (lineations, grain stacking, fissility) deformation, typical of a high-stress polyphase till (Van der Meer 1993; Menzies 2000; Menzies *et al*. 2006).

The fissility exhibited by LF1, in combination with the frequency of shear planes (lineations) suggests that subglacial shear was a major process acting on the till (e.g. Hart & Boulton, 1991) possibly as a result of dewatering (Hicock & Fuller 1995; Evans, *et al*. 2006).The horizontal sand stringers could have formed in a low energy subglacial environment as a result of ice-bed decoupling and the development of a thin water film (cf. Piotrowski & Tulaczyk 1999; Piotrowski *et al*. 2001, 2002, 2004, 2006), or by glaciotectonisation of pre-existing stratified sediments (Hart & Boulton 1991; Hart & Roberts 1994; Hart 1995; Benn & Evans 1996; Evans 2000).

The similar provenances exhibited by the tills points to a common source region. However, the mix of different erratics from the Lake District, Southern Uplands and Vale of Eden make any interpretation of ice flow direction based purely on provenance hard to justify. This range of lithologies from different source regions can be reconciled by the incorporation and cannibalisation of previously deposited sediments. The similar provenances exhibited between the lowermost tills and LF5 (see below) coupled with the macrofabric data suggests that the sediments were deposited by ice flowing down the Vale of Eden. Initially ice flowed north-easterly, converging on the Tyne Gap (LF1) before switching to a north-westerly trajectory, flowing towards the Irish Sea Ice Basin (LF2). The Scottish erratics are likely to have been derived from the ‘Early Scottish advance’ of ice up the Vale of Eden (Hollingworth 1931).

Lithofacies Association 3 (LFA3): Heterogeneous diamicton

Description

LF2 is overlain by a series of interbedded (< 20 cm thick) medium-to-coarse sand and diamicton beds up to 1.1 m thick (Fig. 3, 4c). The diamicton is similar to that of the underlying sediments (LF2) but with fewer clasts (Fig. 4c). The sand, which becomes increasingly dominant upwards in the LFA, is medium-to-coarse textured with some fine gravel (Fig. 3). Loaded contacts and convolute bedding are indicative of soft-sediment deformation (Fig. 4c). A clast macro-fabric taken from the diamicton displays a SW-NE orientation although with a low S_1 eigenvalue of 0.49 (Fig. 6). A thin section taken across the boundary between the diamicton and sand lithofacies reveals a complex sequence characterised by multiple domains often organised into bands, grains with adherent sediment matrixes, soft sediment clasts, load structures, convolute bedding, both planar and vugh voids and rare examples of lineations and turbate structures (Fig. 5c).

The interbedded sand and diamicton grades up into massive coarse-to-medium sand and fine gravel. This lithofacies is up to 0.8 m thick and is characterised by fining upwards, silt and clay balls up to 30 cm in diameter, isolated large pebbles and a pebble layer composed of rounded-to-sub-rounded clasts (Fig. 3, 4d). The isolated pebbles are up to 14 cm in length and tend to be more prevalent towards the top of the lithofacies, while the pebble-gravel layer, which lies 5 cm from the top of the sequence, is composed of a single layer of clasts < 10 cm in length with their long axes arranged horizontally (Fig. 3). LFA3 is, with the exception of the pebble-gravel layer, laterally discontinuous (Fig. 3).

LFA3: Interpretation

The interbedded diamicton and sand beds are interpreted as the product of intermittent debris flow activity. This interpretation is based on the presence of load structures, intercalated sand and diamicton, convolute bedding, multiple domains, gradational contacts and turbate structures seen both macro- and microscopically. These are indicative of deformation associated with high water contents and subject to low-intensity shear or vertical collapse; typically associated with debris flows on the foreland of glaciers (e.g. Lawson 1979, 1981, 1982; Eyles 1987; Lachniet *et al*. 2001; McCarroll & Rijsdijk 2003; Menzies & Zaniewski 2003). Further evidence consists of (a) bands of microscopically observed darker diamicton which form ‘tile’ structures (see Fig. 5c) associated with the pulsed movement of fine grained silts and clays via pore water, which are seen as diagnostic of debris flows (cf. Menzies & Zaniewski 2003); (b) the geometry of the LFA, which is typical of observed debris flow deposits (cf. Lawson 1981, 1982); whilst (c) the upwards relationship through coarse-grained debris flows and then glaciolacustrine sedimentation (see LF4) is indicative of a progressively ice distal environment. The absence of high-strain shear structures such as thrust and detached folds, up-sequence increases in the strain rate or a clear principle strain axis coupled with the low S_1 eigenvalues distinguish LFA3 from glacially overridden sediment (e.g. McCarroll & Rijsdijk 2003; Phillips *et al*. 2008).

The debris flow deposits grade up into medium-to-coarse sand and some fine gravel. These relatively thin deposits are massive, which would suggest an outwash event (Donnelly & Harris 1989) close to the ice margin, or deposition by sediment gravity flows (Eyles *et al*. 1987). Soft sediment balls composed of clay and silt within the sand and gravel are likely to be 'rip-ups' from shallow pools on the glacier forefield (Knight 1999), while isolated pebbles, up to 14 cm in length, were probably cannibalised from pre-existing sub-, supra- or proglacial deposits.

Lithofacies 4: Laminated silts and clays

Description

A sharp boundary marks the upward transition from debris flow deposits and outwash to finely laminated sediments (Fig. 3) comprising sub-centimetre scale dull reddish-brown (5.0 YR 5/3) clay and reddish-grey (2.5 YR 4/1) silt. Sand laminae are discernible in places, especially towards the top of the unit. The upper contact of these laminated sediments with the overlying diamicton (see below) is undulatory, and as a result, the thickness of the laminated unit ranges from a few centimetres up to 1.3 m. Small drop-stones of fine gravel are infrequently found throughout the laminated sequence.

Micromorphological description of the laminated sediments follows the procedure outlined by Palmer *et al*. (2008a), where it is possible to resolve different lamination sets, which are defined by the presence of alternations between silt/sand (coarse component) and clay (fine component) laminations. Three microfacies groups are identified within the laminated section. Evidence of deformation is also described.

Microfacies 1 (0 – 7.52 cm & 21.89 – 35.18 cm from base of lamination unit) comprises 66 couplets. These consist of alternations between a fine component of a single lamination of predominantly clay and a coarse component of predominantly silt or very fine sand laminations (Fig. 7a, b). The clay has sharp upper and lower contacts, is horizontally aligned and grades from very fine silt to clay. Under cross-polarised light, the clay shows high birefringence fabric. The coarse component has multiple, normally-graded silt laminations with sharp, upper and lower contacts (Figure 7a). A maximum of seven graded laminations are identified in the coarse component of the couplet, in this microfacies group. This is sometimes overlain by a single massive, sand or silt lamination, which has a sharp contact to the overlying fine component (Figure 7b). The coarse component is thicker than the fine component. Couplets of the fine and coarse component have an average thickness of 3.1 mm, although there is a wide range (0.30 – 9.74 mm), and considerable variation (standard deviation 2.3 mm) throughout the group.

Microfacies 2 (7.52 – 21.89 cm) comprises 76 couplets and is similar to microfacies 1, but with three key differences. The first is the coarse component is generally composed of a single layer of massive, medium silt, with occasional multiple, very fine, lamination present (Fig. 7c). The second is that the thickness of the coarse component is equal to the fine component. As a result, the third difference is that this microfacies group has thinner couplets than group 1 (1.72 mm on average) with only three sets exceeding 3.5 mm.

Microfacies 3 (35.18 – 101.94 cm) comprises 119 couplets of silt and clay, similar to microfacies 1 except that the laminations are much thicker (5.61 mm on average) and the silt and clay layers are of equal thickness (Fig. 7d). Up to 15 laminations have been measured within the coarse component and large amounts of coarse material have also been observed (Fig. 7e).

Evidence of deformation is apparent towards the top of the lithofacies (101.94 – 127.41 cm) and also through microfacies 2 (11.40 – 12.69 cm). The first area is in microfacies group 2, which is composed of wavy, occasionally truncated, thin clay and silt laminae, with some sub-vertical birefringent fabric development. The second zone of deformed laminations is toward the top of the sequence and is characterised by normal faults, soft-sediment intra-clasts, fabric development within silt bands, sheared boundaries and folding around drop-grains. Further up the sequence, the laminations become heavily contorted and folded (Fig 7f), with more coarse material and vertical injections of sediment. The very top of the sequence is relatively undisturbed and is composed of thin horizontally laminated clay and silt.

LF4: Interpretation

Microfacies 1-3 exhibit a number of distinctive properties which allow them to be interpreted as clastic varves. These include alternations between a coarse grained component (silt and very fine sand), which has multiple graded laminae, and a fine grained component (clay lamina). There is a sharp contact between both the coarse component and the fine component, and between the fine and coarse component. The coarse component records multiple input events as underflows to the lake basin during the summer period, with a sharp contact suggesting a break in sedimentation, possibly related to the autumn overturn (Delaney 2007). The fine component of the clay laminae which succeeds the coarse component forms during the winter through the settling of suspended material in the water column, when the lake water freezes, which inhibits the generation of wind-driven currents in the lake basin that normally causes the re-suspension of clay during the summer (Smith 1978; Smith & Ashley 1985; Ringberg & Erlstrom 1999; Delaney 2008, Palmer 2008b). The sharp contact between the fine winter lamina and the succeeding coarse summer laminae indicates the renewed transport of sediment to the lake basin during the spring/summer melt season (Ashley 1975). The varve thickness characteristics throughout the sequence are indicative of sedimentation in distal zones of the lake basin (Ashley 1985; Smith & Ashley 1985; Ringberg & Erlstrom 1999). The lack of coarse-grained dropstones could also indicate ice-distal conditions, or alternatively shallow water. There are subtle differences in the microstructures observed within the summer layers which are highlighted below.

Microfacies 1 is characterised by micro-laminated silt layers, with up to 7 laminations of graded sediment. These micro-laminae relate to episodic density current underflows which could result from the release of meltwater from different sources (Palmer *et al*. 2008b), increases in meltwater discharge (Ringberg & Erlström 1999), surge currents from slumping sediment (Smith & Ashley 1985) or rainfall events (cf. Delaney 2007). The coarse material deposited at the top of the silt component are potentially attributed to wind generated bottom currents produced during storms in the late melt season (Ringberg & Erlström 1999), precipitation events during the Autumn (Smith 1978) or an increase in aeolian deposition (Lamoureux & Gilbert 2004). The sharp contact at the base of the silt layer marks the sudden onset of underflow sedimentation during the spring melt season, while graded structures in the summer layer (coarse component) indicate complete sedimentation from each discrete sediment pulse that enters the lake basin.

Microfacies 2 has fewer laminations of sediment within the summer layer suggesting that only extreme underflow events entering the lake have sufficient velocity to reach this point in the basin, and therefore, the varve thickness is decreased (Ringberg & Erlström 1999). This could be the product of reduced glacier extent within the catchment (Leonard 1986) and the similar thickness of the coarse and fine component may relate to increased dispersion of sediment from the input region or a reduced melt season.

Microfacies 3 is interpreted in a similar way to microfacies 1. However, a greater average thickness of the couplets, number of laminae within the coarse component and presence of coarse grained material suggests that the microfacies was relatively more ice-proximal, although still in a distal part of the basin.

Deformation exhibited within the sequence comprises faults, folds, water escape structures, loading, soft-sediment intra-clasts and sheared boundaries thus indicating that the varves were subject to post-depositional deformation. The prevalence of these structures at the top of the sequence coupled with the upwards increase in strain signature indicates that the deformation may be related to the emplacement of the capping diamicton (see below).

A total number of 261 years are recorded in the sequence and the thickness varies through time (Fig. 8). Initially the thickness decreases marked by a transition from microfacies 1 (24 years) to microfacies 2. Deposition of microfacies 2 occurred over a period of 76 years before varve thickness increases through microfacies 1 (41 years) and then the microfacies 3 (119 years) at the top of the sequence, where the thickest varves are observed (Fig. 8). These varve thickness changes may reflect variations in the position of the ice margin such that ice retreats allow the formation of glaciolacustrine varve sediments. Subsequently ice extent increases within the catchment until eventually the lake sediments are overridden indicated by the deformation of the sediments at the top of the sequence (see below). The varves subjected to deformation were not included within the time sequence due to the implicit difficulties of producing an accurate count when the laminae have been subject to shortening, compression and possible repetition due to folding and shearing. This implies that the varve count probably under-estimates the duration of the lakes existence.

Lithofacies 5: Uppermost diamicton

Description

The uppermost lithofacies consists of a massive, dull reddish-brown (5 YR 4/4) matrix-supported diamicton (Fig. 4e) up to 20 m thick that forms the bulk of the drumlin (Fig. 3). The diamicton is clast-rich with sub-rounded to rounded morphologies and rare striae. Its erratic content (Table 2) is similar to LF1 and LF2 with 40% greywacke, 15% sandstone 13%, Carboniferous limestone, 7% dolerite, 5% Permo-Triassic sandstone, 4% slate, 2% Southern Upland granites and 1% lava's from the Borrowdale Volcanic Group. The boundary between the diamicton lithofacies and the underlying laminated clay and silt is erosional and undulatory (Fig. 3). A thin section taken across this boundary displays deformed clay and silt, with normal faulting and folding (Fig. 5d). Rip-up clasts of clay have been incorporated into the diamicton unit and a darker diamicton domain towards the boundary edge shows evidence of strong banding and skelsepic plasmic fabrics (Fig. 5d). The diamicton within the thin section is dominated by dipping lineations, crushed grains, pressure shadows and occasional turbate structures. A clast fabric taken near the base of the lithofacies (Fig. 6) indicates a SE-NW orientation with a strong S_1 eigenvalue of 0.62.

LF5: Interpretation

LF5 is interpreted as a subglacial traction till (*sensu* Evans *et al*. 2006) based on the following evidence: (a) the large thickness (20 m) and massive nature of the diamicton which has been moulded into a SE-NW orientated drumlin, indicating glacial overriding (Menzies 1979; Boulton 1987; Hart 1997) and probably some localised folding and stacking (Evans *et al*. 2006); (b) the fabric orientation is consistent with the drumlin's form, while the high S_1 eigenvalue indicate strong streamlining of clasts in the direction of ice motion (Benn 1995; Evans 2000); (c) clasts contain a far travelled provenance; (d) the diamicton has a sharp, erosional boundary with the underlying, partially deformed varved clay and silt lithofacies (cf. Evans *et al*. 2006); (e) clay and soft sediment clasts have been incorporated into the till matrix, indicating cannibalisation of the underlying varved sediments and incorporation into the diamict (e.g. Hicock & Dreimanis 1989; Benn & Evans 1996; Hooyer & Iverson 2000); and (f) the thin section contains lineations, turbate structures, skelsepic and banded plasmic fabrics, crushed grains and pressure shadows typical of a polyphase till that underwent brittle and ductile deformation (Van der Meer 1993; Menzies 2000; Hiemstra & Rijsdik 2003; Menzies *et al*. 2006). Indeed the propensity of lineations and low porosity relative to other microstructures suggests that brittle shear was the dominant process near the base of the diamicton (Menzies 2000; Menzies *et al*. 2006).

The sub-rounded to rounded clast morphologies coupled with the diverse clast provenance and rare striae suggests that the till has incorporated glaciofluvial/glaciolacustrine sediment and diamicton of previously deposited sediment (Hicock & Dreimanis 1989; Hicock & Fuller 1995; Evans 2000). The orientation of the drumlin indicates that ice flow was northerly moving down the Vale of Eden before swinging west into the Irish Sea Ice Basin (Livingstone *et al.* 2008).

<u>Maryport</u>

Maryport (NY 038 378) is located on the Cumbrian coast (Fig. 1), with exposures cut into the southwest flank of a large, NE-SW orientated drumlin (56 m O. D.). The drumlin forms part of a flowset which sweeps round the northern edge of the Lake District (Fig. 2). The underlying bedrock is Permo-Triassic sandstone. The sedimentary lithofacies are depicted diagrammatically in Fig. 9 and described below.

Description

The exposure displays one lithofacies, which is up to *ca*. 20 m thick and comprises a matrix-supported, clast-rich reddish (5 YR 3/3) diamicton (Fig. 9, 10a). Clasts have sub-angular to sub-rounded morphologies and rare striae. The principle lithologies (Table 2) are Silurian greywacke (24%), sandstone (13%), siltstone (10%), Permo-Triassic sandstone (9%), dolerite (7%), lava's from the Borrowdale Volcanic Group (6%), and quartzitic sandstone (6%), with 2% Southern Upland granite and 1% Lake District granite also observed. Two clast fabrics taken in the upper exposures (Fig. 9) indicate a NE-SW orientation with S_1 eigenvalues of 0.77 and 0.60. A thin section from the diamicton exhibits lineations, turbate structures, skelsepic plasmic fabrics and adherent sediment matrixes.

The diamicton contains a series of interbedded and folded sand and fine-gravel beds (Fig. 9, 10a & b), sand stringers and pebble gravel clusters (Fig. 10c). Two large sand open folds up to 1.0 m thick exposed about half way up the drumlin (Fig. 9, 10a) show evidence of primary stratification and are attenuated-out towards the S/SW, with the hinge line orientated towards the NE. This stratification has been partially deformed, with centimetre-scale folds, faults and clay stringers. Boundaries with the diamicton are sheared and mixed. In the uppermost exposures the diamicton is inter-bedded with slightly folded/faulted sand stringers, sand lenses, fine-pebble gravel beds (up to 50 cm) and partially stratified sand beds (up to 50 cm) (Fig. 9, 10b). These upper diamictons are very clast rich with sub-rounded to rounded morphologies. A vertically orientated, clast-supported pillar of pebble-gravel *ca*. 10 cm wide by *ca*. 40 cm high was also observed (Fig. 10c)

Interpretation

The lithofacies at Maryport are consistent with an origin as a glaciotectonite (*sensu* Evans *et al*., 2006) derived from the glaciotectonisation of underlying glaciofluvial/glaciolacustrine sediments (e.g. Broster 1991; Hart & Roberts 1994; Benn & Evans 1996; Evans 2000). The surface of the diamicton has been moulded into a drumlin, with the high S_1 eigenvalues of the diamicton suggesting subglacial streamlining of a deformable substrate (Benn 1995; Evans 2000; Evans *et al*. 2006). Sand and gravel beds, which have been preserved as competent masses, have been rafted upwards into the diamicton and folded and attenuated in the direction of ice flow (cf. Hart & Boulton 1991; Hart & Roberts 1994) whilst rounded to sub-rounded clasts and adherent sediment matrixes suggest the cannibalisation of underlying sediment to form at least part of the till matrix (e.g. Hicock & Dreimanis 1989; Hicock & Fuller 1995, Evans 2000). Thus the clast provenance is likely to have been partially derived from the underlying glaciofluvial/glaciolacustrine sediments. The sand lithofacies show some preservation of primary stratification. The presence of thinly bedded (< 50 cm), massive, clast-rich diamicton, with sand lenses and thin beds of fine-to-coarse sand, in association with the gravel units, supports a debris flow interpretation (e.g. Eyles 1987). The vertical pillar of clast-supported, clast rich pebble-to-cobble gravel is likely to be a clastic dyke derived from a gravel-rich unit (cf. Van der Meer *et al*. in press). This, coupled with the streamlining of the diamicton and the sand, and the attenuated sand structures all indicate that there were high pore water pressures (e.g. Broster 1991) and that simple shear was a key mechanism in the creation of internal structure (e.g. McCarroll & Rijsdijk 2003).

Swarthy Hill

Exposures at Swarthy Hill (NY 069 403) are located in the southwest flank of a NE-SW orientated drumlin (31 m O. D.). The drumlin is part of an arcuate flowset which sweeps round the northern edge of the Lake District (Fig. 1 and 2). The underlying bedrock geology is composed of Permo-Triassic sandstone. The sedimentary lithofacies are depicted diagrammatically in Fig. 11 and described below.

Description

The exposures at Swarthy Hill are dominated by fine sand beds up to 1.5 m thick (Fig. 10d, 11). These are chiefly massive in structure with some fine-to-cobble gravel (maximum diameter 20 cm), occasional silt and clay, and ripple structures (orientated towards the SE/ESE). Where present, pebble-to-cobble gravel is dispersed throughout the sand. The sand has been heavily deformed (Fig. 10d), with clay and silt stringers, clay pods, diamicton inclusions, fold structures, reverse and normal faults (with centimetre-scale offsets), flame structures, convolute bedding, coarse sand injections and cross-cutting laminae composed of fine sand, clay and silt.

Overlying the sand lithofacies are beds of up to 1.0 m thick matrix-supported, massive fine-to-cobble gravel, comprising sub-rounded to rounded clasts up to 30 cm diameter (Fig. 11). Thin (< 5 cm) diamicton and fine sand units are inter-bedded with the gravel (Fig. 11). The gravel lithofacies has a loaded lower boundary.

Interbedded within the deformed sand is a thin (up to 0.5 m thick) bed of matrix supported, dull reddish-brown (5 YR 4/4) diamicton (Fig. 10e) that dips towards the southeast. The diamicton has a sandy composition with rounded to sub-rounded clasts up to 30 cm in diameter, indistinct sheared boundaries and a NW-SE fabric (S_1 eigenvalue of 0.48) (Fig. 11). The provenance of the diamicton is very similar both to the subglacial till identified at Maryport and the sand and gravel exposed at Swarthy Hill (Table 2). This includes 32% sandstone, 15% Silurian greywacke, 10% lava's from the Borrowdale Volcanic Group, 6% siltstone 4% dolerite, 2% Scottish granite and 2% Lake District granite. Thin units of diamicton (< 10 cm thick) are also interbedded within the gravel lithofacies, and as stringers and pods within the sand (Fig. 11).

Sheets of sub-vertically and sub-horizontally aligned fine-grained sand, with sand, silt and clay laminae up to 5 cm wide are observed throughout the exposure (Fig. 10f, 11). These structures have consistent widths, although with tapered-out, dendritic ends (Fig. 11) and occasional 'burst-out' (Rijsdijk *et al*. 1999, Van der Meer *et al*. in press) structures (Fig. 10f). The laminae run parallel to the walls of these structures and often cross-cut each other. These structures divert around clasts, finger-up into diamicton lithofacies and run vertically into the gravel beds.

Interpretation

The stratified sand and gravel lithofacies are thought to have been initially deposited as proglacial glaciofluvial and glaciolacustrine sediments in an ice proximal location, with dropstones indicating ice-rafting into a standing water body. Infrequent ripples suggest a SE/ESE flowing palaeocurrent, while the sheared interbedded diamicton is probably a debris flow (Lawson 1981, 1982). Partial destruction of the primary structure by extensive deformation and dewatering clearly demonstrates subsequent overridding and glaciotectonisation of the pre-existing sediments into a drumlinoid form (e.g. Krüger & Thomsen 1984; Boulton 1987; Boyce & Eyles 1991; Menzies & Brand 2007). Although the sediments are composed of stratified sand, gravel and diamicton, a potential genesis as a lee side cavity fill is ruled out due to: (a) the lack of distinctive dipping stratification (cf. Hillefors 1973; Dardis *et al.* 1984; Dardis 1985; Levson & Rutter 1989a,b); (b) the deposits show extensive evidence of deformation which is absent in the cores of cavity-fills (Dardis *et al.* 1984; Dardis 1985); (c) diamicton debris flows present within the sequence are interpreted as subaerial rather than the subaqueous flows diagnostic of cavity-fills, due to the lack of internal stratification/lamination (Evenson *et al*. 1977), the thin geometry, sheared boundaries and positioning between deformed sand lithofacies (Lawson 1981, 1982); and (d) the ripple structures are orientated at right angles to the orientation of the drumlin, which is at odds with the cavity-fill model of water flow into the cavity, and indeed suggests a mode of deposition independent of drumlin formation.

During glaciotectonisation the sediment is inferred to have been subject to fluctuating pore water pressures. This is consistent with: (a) centimetre-scale coarse-sediment injections (Fig. 9d), interpreted as water escape structures caused by pervasive de-watering; and (b) the bundles of sub-horizontally and sub-vertically orientated fine grained laminae, with occasional burst-out structures, which are interpreted as clastic dykes (e.g. Rijsdijk *et al*. 1999; Van der Meer *et al*. 1999, in press; Kjaer *et al*. 2006). Clastic dykes are thought to form by hydro-fracturing as a result of changing water pressures (Van der Meer *et al*. in press). The clastic dykes cut across multiple lithofacies thus indicating that dewatering must have occurred post-depositionally.

Borehole stratigraphy

Borehole data gathered from the motorway M6 between Penrith and Carlisle (Fig. 1) augments observations from this study and can be compared and correlated with previous research (Goodchild 1875; Dixon *et al*. 1926; Huddart 1970; Arthurton & Wadge 1981) in order to better define the regional stratigraphy. Figure 12 presents the locations and thicknesses of the laminated sediments and Figure 13 displays the general stratigraphy of the Vale of Eden. The thickness of the upper diamicton is also recorded (Fig. 12c). Evidence for the tripartite sequence is commonplace in the lower reaches of the Vale of Eden (especially in the Petteril and Caldew Valleys) below *ca*. 70 m (Fig. 12, 13). Two firm to stiff reddish-brown diamictons have been recorded separated by up to 15 m of sand, gravel and laminated clay/silt lithofacies (Fig. 13). Above 70 m O. D., interbedded sand with some laminated clay/silt is still evident at Calthwaite, Langwathby and Low Dyke (Fig. 12b, 13). Some evidence of a tripartite sequence is also observed to the west of Blackhall Wood, with laminated clays observed in boreholes at Thursby (Fig. 12) and middle sands exposed along the River Wampool (Dixon *et al*. 1926). However, borehole evidence seems to suggest that most drumlins comprising the westerly orientated flowset tend to be diamicton cored (with a capping unit of sand and gravel). The heights of the laminated clay-silt lithofacies within the stratigraphic sections generally range between 46-54 m O. D, although they are also infrequently recorded higher up the Vale of Eden (e.g. 133 m O. D. at Low Dyke) (Fig. 12b, 13). Thicknesses of the laminated clay and silt range widely from < 0.5 – 5.2 m, with the thickest exposures exposed in the Petteril Valley below the 100 m contour (Fig. 12b). There seems to be a general thinning west and east of the Petteril Valley, although high values are persistent northwards towards Carlisle (Fig. 12b). The thickness of the upper diamicton is variable, ranging from less than 5 m to over 20 m (Fig. 12c). The general up-down trend in thickness is probably a function of drumlins vs intervening hollows (Fig. 12c), although it is interesting to note a couple of particularly thick sequences of the upper diamicton at Calthwaite and Foul Bridge (Fig. 12c, 13). The large thickness of diamicton at Foulbridge is of particular importance as it corresponds with a faint ridge which can be traced to the west and east along the edge of a lobate flowset extending northwards into the fringe of the Solway Lowlands (Fig. 2).

Stratigraphic evidence derived from prior investigations and borehole logs can be correlated to the field section at Blackhall Wood. The uppermost dull reddish-brown till at Blackhall Wood is ubiquitous, being observed throughout the study area (Dixon *et al*. 1926; Trotter 1929; Hollingworth 1931), making up the drumlinoid features that cover the Vale of Eden. The middle sand, silt and clay lithofacies, which are often found sandwiched between diamicton units, exhibit similar heights and characteristics, with a propensity for laminated sequences in close proximity to Blackhall Wood. Thus, the middle sand, gravel and laminated clay-silt lithofacies can be correlated with some confidence. The laterally discontinuous, lower reddish-brown diamicton probably correlates to the lowermost Blackhall Wood tills (LF1 and LF2). Dixon *et al*. (1926) highlighted the lithological similarities between the lower and upper tills, which both have a reddish-brown colour. In the past, the lower diamicton has been correlated with an early (Late Devensian) Scottish advance across Stainmore (e.g. Dixon *et al*. 1926; Hollingworth 1931). However, evidence produced in this study would suggest that, whilst the erratic support for a Scottish-sourced ice flow is compelling, both the lower and upper tills, in the Solway Lowlands, were deposited from ice moving down the Vale of Eden.

Discussion

The sedimentary and stratigraphic data obtained during this study indicates that initial deglaciation of the Solway Lowlands resulted in an ice-free enclave with ice situated in the Solway Firth promoting the rapid development of 'Blackhall Wood' Lake. Varve records suggest that this lake persisted for at least 261 years; however, the truncated upper boundary and deformed varves mean that it probably existed for longer. Re-advance of ice over the glaciofluvial and glaciolacustrine sediments produced the characteristic tripartite sequence, with the Irish Sea Ice Stream drawing Scottish and Lake District ice westwards into the Irish Sea Basin and leaving behind what is interpreted to be a fast-flow signature constrained by hummocky terrain and ribbed moraine along the northern margin of the 'flowset' (Fig. 2, 14) (cf. Ross et al. 2009). A subglacial interpretation for the tripartite sequence (cf. Huddart 1970; Goodchild 1875, 1887) is rejected here based on the presence of clastic varves and their association with both debris flow and outwash deposits at the Blackhall Wood site.

Incorporation of Scottish erratics within the lowermost till units at Blackhall Wood supports previous research that proposed an initial southwards flow of Scottish ice up the Vale of Eden and across the

Stainmore Gap (e.g. Goodchild 1875; Hollingworth 1931). However, the fabric and provenance data can be reconciled with geomorphic evidence (cf. Livingstone *et al*. 2008) that suggests that these tills were deposited during ice flow convergence on the Solway Lowlands from the Southern Uplands and the Lake District, before moving eastwards through the Tyne Gap (Fig. 14a). At some point, probably as ice was downwasting, ice flow switched direction and instead flowed westwards towards the Irish Sea Ice Basin (LF 2: Blackhall Wood).

The debris flow deposits at Blackhall Wood record an increasingly ice distal environment as the Solway Lowlands became deglaciated. Fieldsites at Swarthy Hill and Maryport, which show evidence of glaciotectonised proglacial glaciofluvial and glaciolacustrine sediment, indicate that ice retreated west of the Cumbrian coast. Therefore, a stretch of at least 42 km along the Solway Lowlands became deglaciated either concurrently or time-transgressively, in relation to changes in the ice sheet's configuration (Fig. 14b). Boreholes along the M6 show evidence for extensive deglaciation within the Vale of Eden (Fig. 14b). However, the debris flows composed of diamicton and outwash at Blackhall Wood indicate that ice was in close proximity. Indeed the Solway Lowlands probably formed an ice-free enclave, constrained to the west by the still active Irish Sea Ice Stream (e.g. Merritt & Auton 2000). Tripartite sequences north of Carlisle and along the Scottish border have been correlated with the latter Scottish readvance (cf. Stone *et al*. in press) with the corollary being that ice probably maintained a permanent presence within this region during the earlier retreat (Fig. 14).

Formation of a glacial lake in the Solway Lowlands must have occurred when drainage westwards into the Irish Sea Ice Basin became impeded (Fig. 14b). The most logical reason for this involves the re-advance of ice eastwards into the Solway Lowlands, cutting off drainage pathways to the west and north, and causing the natural topographic basin to rapidly fill with meltwater. The sharp contact between the clastic varves and underlying outwash implies that this switch to glaciolacustrine conditions occurred rapidly. Indeed the concept of a re-advance is supported at Blackhall Wood where the varves thicken up-sequence, typical of an increasingly ice-proximal depositional environment (cf. Smith & Ashley 1985). The slightly thicker varves observed at the very base of the sequence probably relate to initially high sediment fluxes derived from unconsolidated proglacial sediment (cf. Palmer *et al*. 2008b).

The spatial extent of the lake is hard to estimate due to the likely time-transgressive mode of formation associated with changes in the ice-sheet configuration directly impacting upon the evolution of the ice-contact lake. This is also compounded by both the re-advance of ice westwards into the Irish Sea Basin which would likely have eroded much of the evidence and the subsequent Scottish re-advance, which produced a separate tripartite division in the northern-most region of the Solway Lowlands (cf. Stone *et al*. in press). However, the height of the laminated clay-silt lithofacies are generally between 46 - 54 m O. D. indicating that the lake surface must have been at least 54 m O. D. This is deep enough to flood the entire Solway Basin (an extent > 1,350 km^2), and enables rough estimates demarcating the topographic boundaries of the lake (Fig. 14). The northern and western limits are currently unknown, with the only constraint being the position of the ice barrier which would have retarded drainage and caused the lake to form. This could have been anywhere from the current Cumbrian coastline to the edge of the Lias plateau (Fig. 14b). Borehole evidence of laminated lithofacies indicate that the lake was probably dammed up against the Vale of Eden slope with the lake extent at least 140 km^2 in size (Fig. 14). Formation of the lake is therefore regarded as a final stage of ice-free conditions within the Solway Basin. Whereas deposits associated with the ‘main lake’ are consistently observed between 46 – 54 m O. D., disparate laminated lithofacies observed at a range of higher elevation, up to 133 m O. D. may imply more localised ponding.

The glacial oscillation within the Solway Lowlands is hard to constrain chronologically. Superimposed glacial landforms, including eskers at Thursby and the Holme St Cuthbert deltaic complex (Huddart 1970; Livingstone *et al*. 2008) coupled with the stratigraphic position of the tripartite sequence within Vale of Eden drumlins relating to late-stage flow into the Irish Sea Ice Basin indicates that the oscillation must have occurred prior to the Scottish re-advance (*ca*. 16.8 cal ka BP). The Gosforth oscillation observed within the Sellafield District, Cumbria at *ca*. 19.5 cal ka BP (Merritt & Auton 2000), prior to the Scottish re-advance offers a possible regional correlation with the Blackhall Wood re-advance. The evidence in this study is consistent with the reconstruction of Merritt

& Auton (2000) at Sellafield, with valley ice retreating back into the Lake District while the retreating Irish Sea Ice Stream remained along the coast, thus leading to the formation of proglacial lakes. The subsequent Gosforth re-advance then coalesced with the Irish Sea Ice Stream, leading to extensive drumlinisation (Merritt & Auton 2000). All these features are noted in the Solway Lowlands.

Given that the Blackhall Wood oscillation occurred during a relatively early stage of deglaciation, prior to the Scottish re-advance, it is likely that the Irish Sea Ice Stream was still quite extensive, stretching down the west Cumbrian coast (Merritt & Auton 2000) and reaching as far south as the Isle of Man, at the very minimum (cf. Thomas *et al*. 2004).

The data from this study indicate that the subsequent flow of ice, sourced in the Vale of Eden/Lake District was in fact a re-advance (Fig. 14c). Ice moved down the Vale of Eden, before swinging round the northern edge of the Lake District and coalescing with the Irish Sea Ice Stream (cf. Livingstone *et al*. 2008). Structural characteristics exhibited within sections at both Swarthy Hill and Maryport suggest drumlinisation of saturated sediment. Glaciofluvial and glaciolacustrine sediments were glaciotectonised and rafted, with evidence for subsequent de-watering. The northern margin of this ice flow imprint is constrained by a transition into less elongate drumlins and then hummocky terrain and ribbed moraine (Fig. 2) (e.g. Ross *et al*. 2009). This seems to depict the lateral margin of the flow set where inferred ice velocities were significantly reduced (Livingstone *et al*. 2008; Ross *et al.* 2009).

As the 'Blackhall Wood' lake developed within an ice free enclave, constrained by the Irish Sea Ice Stream to the west, it is interesting to speculate on the likely impact of large scale drainage of this, and other lakes in the central sector of the BIIS. The northern and western borders of the 'Blackhall Wood' lake are as yet, unconstrained, and therefore, could possibly have stretched at least 1,350 km^2, while further ponding dammed by the Irish Sea Ice Stream has been proposed by Merritt and Auton (2000) along the west Cumbrian coast, during the Gosforth oscillation. Furthermore, lake development seems to have been a regular occurrence throughout the Solway Lowlands, with the Holme St. Cuthbert delta (Huddart, 1970, 1991, 1994), and glaciolacustrine deposits east of Carlisle (Dixon *et al*. 1926; Trotter 1929) recording the development of large ice-dammed lakes during laterstage episodes of retreat and re-advance. Another significant lake is thought to have existed in the Machars, SW Scotland (Salt 2001). There, ice-contact sub-aqueous cones at 50 m O. D., stretching from the Isle of Whithorn to Arbrack (Charlesworth 1926; Salt 2001), have been interpreted to mark the location of a lake dammed up by Scottish ice to the NE and Lake District ice coming out of the Solway Firth (Salt 2001). The influence of widespread lake development on ice dynamics in the Irish Sea Ice Basin has not previously been considered, yet offers a potential mechanism for triggering fast ice flow both ice-marginally and regionally (Stokes & Clark, 2004).

Conclusions

Evidence presented in this study indicates that a major glacial oscillation, leading to ice free conditions and proglacial lake formation, occurred within the Solway Lowlands prior to the well documented Scottish readvance at *ca.* 16.8 ka BP. This ‘Blackhall Wood oscillation’ is tentatively correlated with the Gosforth oscillation (Merritt & Auton 2000) and records the first partial retreat of the Main Late Devensian glaciation within the central sector of the BIIS.

The interpretation of the lamination silt and clay lithofacies within the ‘tripartite sequence’ as varves has allowed a subglacial hypothesis to be discarded in favour of proglacial lake formation. Micromorphological analysis of the varved sediments has enabled a detailed model of the lake’s development to be formalised, and the derivation of a high resolution ‘floating point’ chronology.

The ‘Blackhall Wood oscillation’ demonstrates the non-linear interplay between the Irish Sea Ice Basin and the Solway Lowlands. Despite the scale of the ‘Blackhall Wood oscillation’, which saw one of the core regions of the BIIS become ice free, ice continued to operate within the Irish Sea Basin, probably as an ice stream. This period of ice-free conditions was then followed by a significant re-advance of ice capable of flowing out into the Irish Sea Basin, depositing thick diamicton sequences and moulding elongate landforms. This indicates that the oscillation was not a small-scale fluctuation, rather a period when ice in a core region of the BIIS and in close proximity to key upland dispersal centres rapidly lost and then gained mass, whilst other regions remained relatively stable.

This study suggests that the Irish Sea Ice Stream was still active at a late stage of the Main Late Devensian glaciation, at a time following the 'Blackhall Wood oscillation'. The Solway region itself does not seem to have acted as a major tributary to the Irish Sea Ice Stream barring both the initial contraction and subsequent re-advance of ice associated with the 'Blackhall wood oscillation'. Prior to this oscillation the geomorphic evidence indicates an easterly ice trajectory through both the Stainmore and Tyne gaps from the Vale of Eden and Solway Lowlands (Livingstone *et al*. 2008).

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Figure captions:

Figure 1: Topographic map showing key locations and fieldsites.

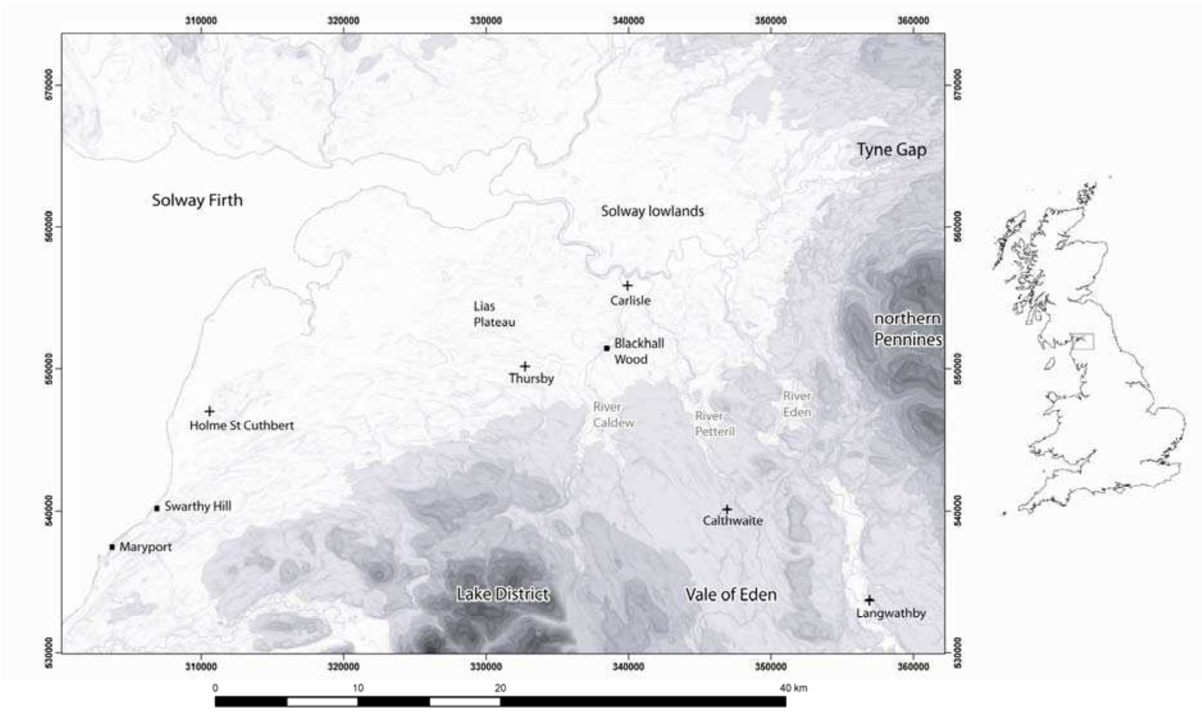


Figure 2: Glacial geomorphology of the Solway Lowlands. Re-produced from Livingstone *et al*., (2008).

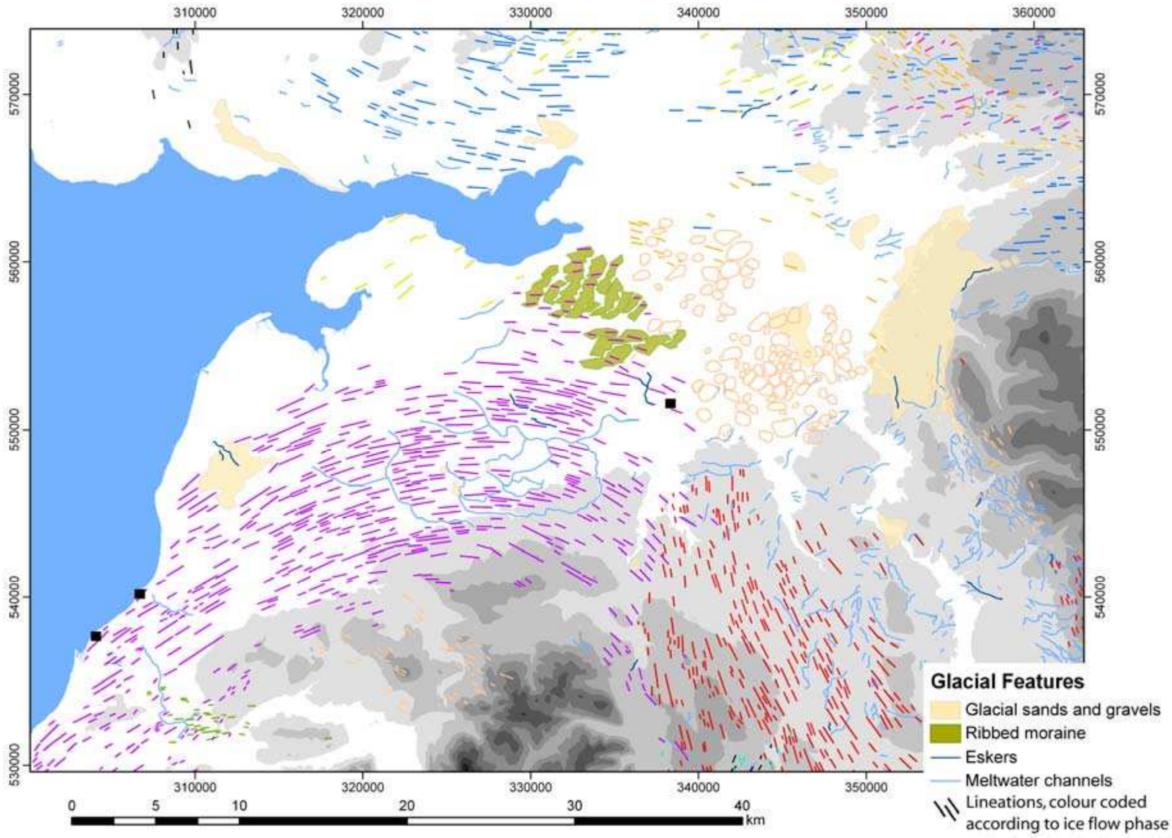


Figure 3: Blackhall Wood section logs from a ca. 20 m long cliff exposure. Thin section numbers refer to Figure 5.

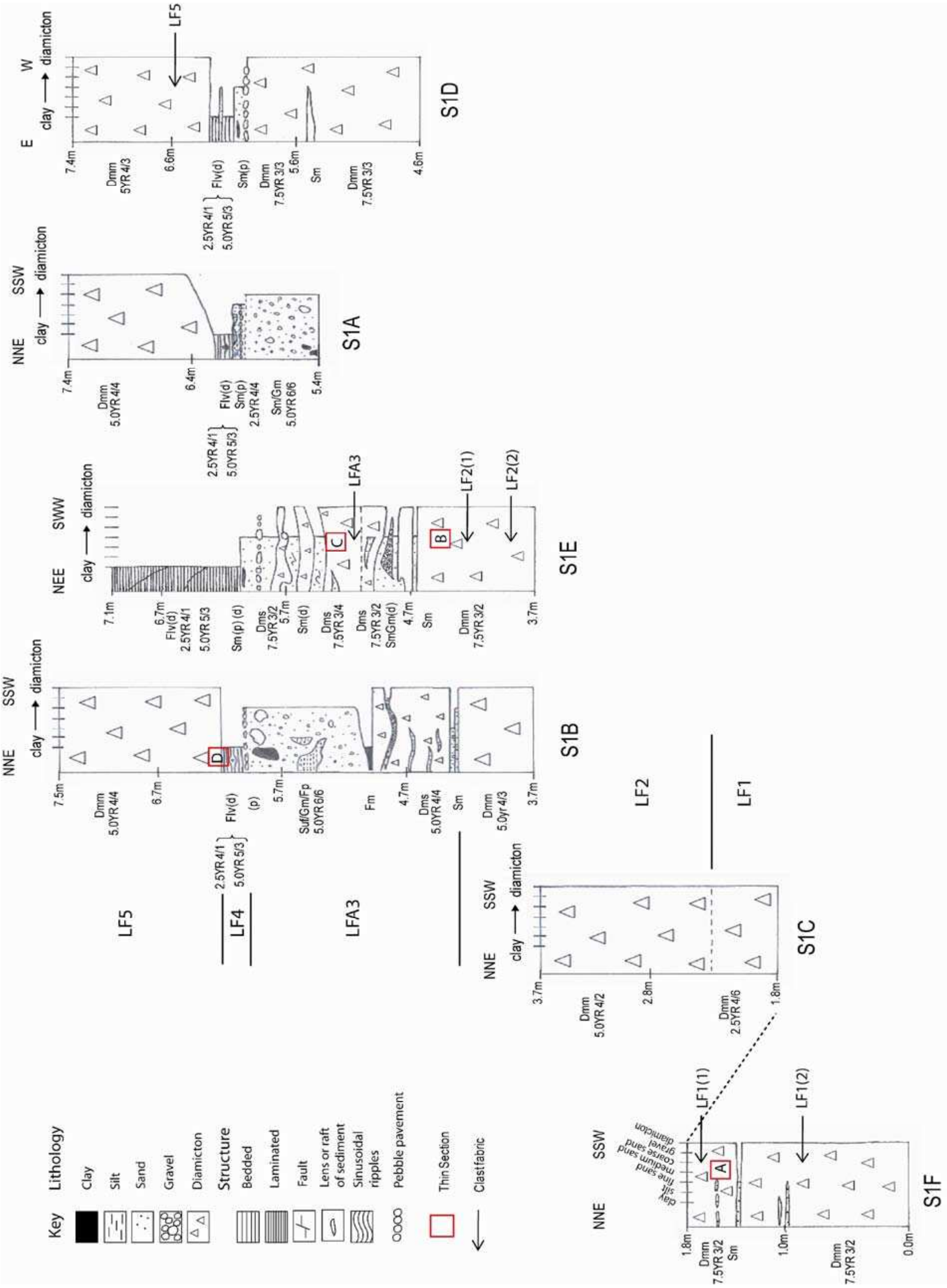


Figure 4: Pictures of Blackhall Wood lithofacies: (a) LF1 characterised by sand partings and a high degree of fissility; (b) indistinct boundary between LF1 and LF2; (c) interbedded diamicton and sand (LFA3); (d) sand lithofacies with clay/silt balls, drop stones and a pebble pavement; and (e) laminated silt and clay lithofacies (LF4) capped by diamicton (LF5). Red box indicates position of thin section taken from LFA3.

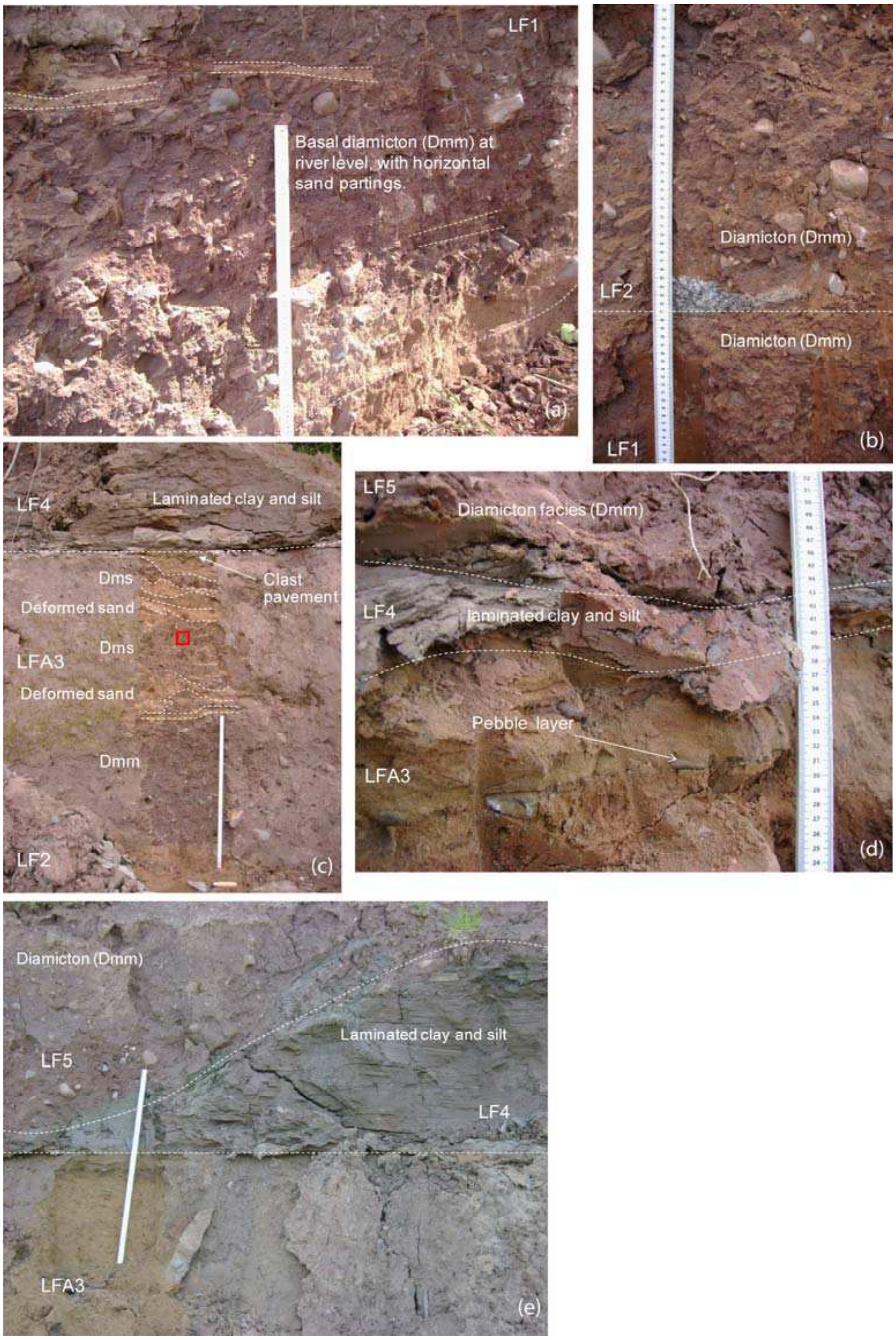


Figure 5: Thin sections from Blackhall Wood. All paired-images are scans of the original thin sections, while the scale bar is located at the bottom of each original image. In all images dotted lines mark the boundary between domains, while arrows indicate turbate structures, voids are delineated by shading and lines demarcate lineations/grain stacking. In image (d) in the laminated clay and silt lithofacies the lines instead indicate the position of faults: (a) LF1: lineations, grain stacking, turbate structures, adherent sediment matrixes, strong skelsepic plasmic fabric and an interconnected network of sub-horizontal and sub-vertical fissures; (b) LF2: lineations and grain stacking with some small turbate structures and a weak skelsepic plasmic fabric; (c) LFA3: interfingering sand, and diamicton of which there a number of domains. Planar and vugh voids, soft sediment pods and adherent sediment matrixes are also evident and (d) LF4 and 5: the laminated clay and silt is heavily deformed, with a series of normal faults and contortions. The boundary with LF5 is uneven, indistinct and graded with evidence of loading strictures. LF5 is characterised by two domains, with evidence of soft sediment rip up from the underlying lithofacies, soft sediment pods containing LF4, lineations, grain crushing and small turbate structures. LF5 exhibits a strong skelsepic fabric with evidence of some banding.

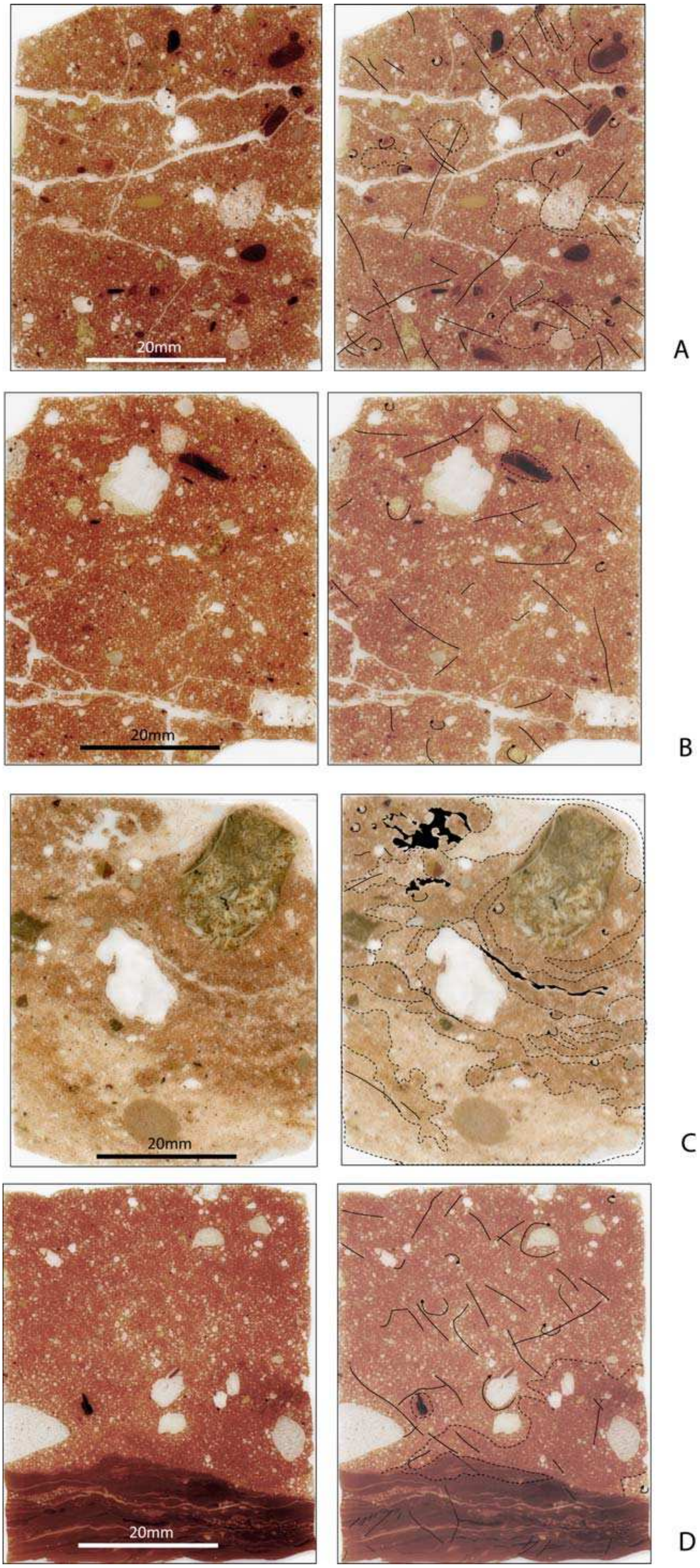

20mm
A
20mm
B
20mm
C
20mm
D

Figure 6: Clast fabrics for Blackhall Wood

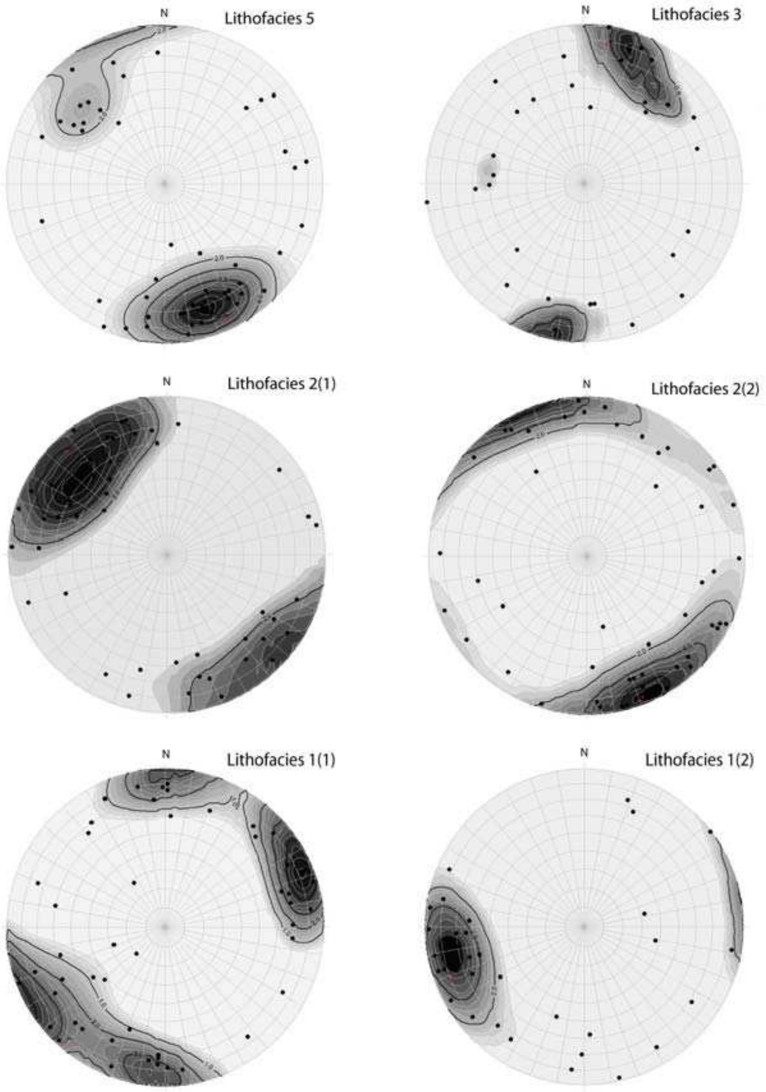


Figure 7: Microphotographs of typical laminated sediment from Blackhall Wood. All images have a scale bar in one of their corners and are under cross-polarized light. Black bars delimit one varve. (a) microfacies group 1: note the sharp contacts between the top of the fine lamina and beginning of the coarse component, which is characterised by multiple, normally graded silt laminations; (b) microfacies 1: note the coarser laminae of very fine sand and silt towards the top of the coarse component, which is normally graded; (c) microfacies 2: very thin regular laminae with sharp boundaries; (d) microfacies 3: thick laminae which are characterised by multiple normally graded laminations within the coarse component; (e) microfacies 3: characterised by coarser very fine sand

and silt towards the top of the coarse component; and (f) deformed laminae: heavily contorted and folded laminae, with large amounts of coarse material.

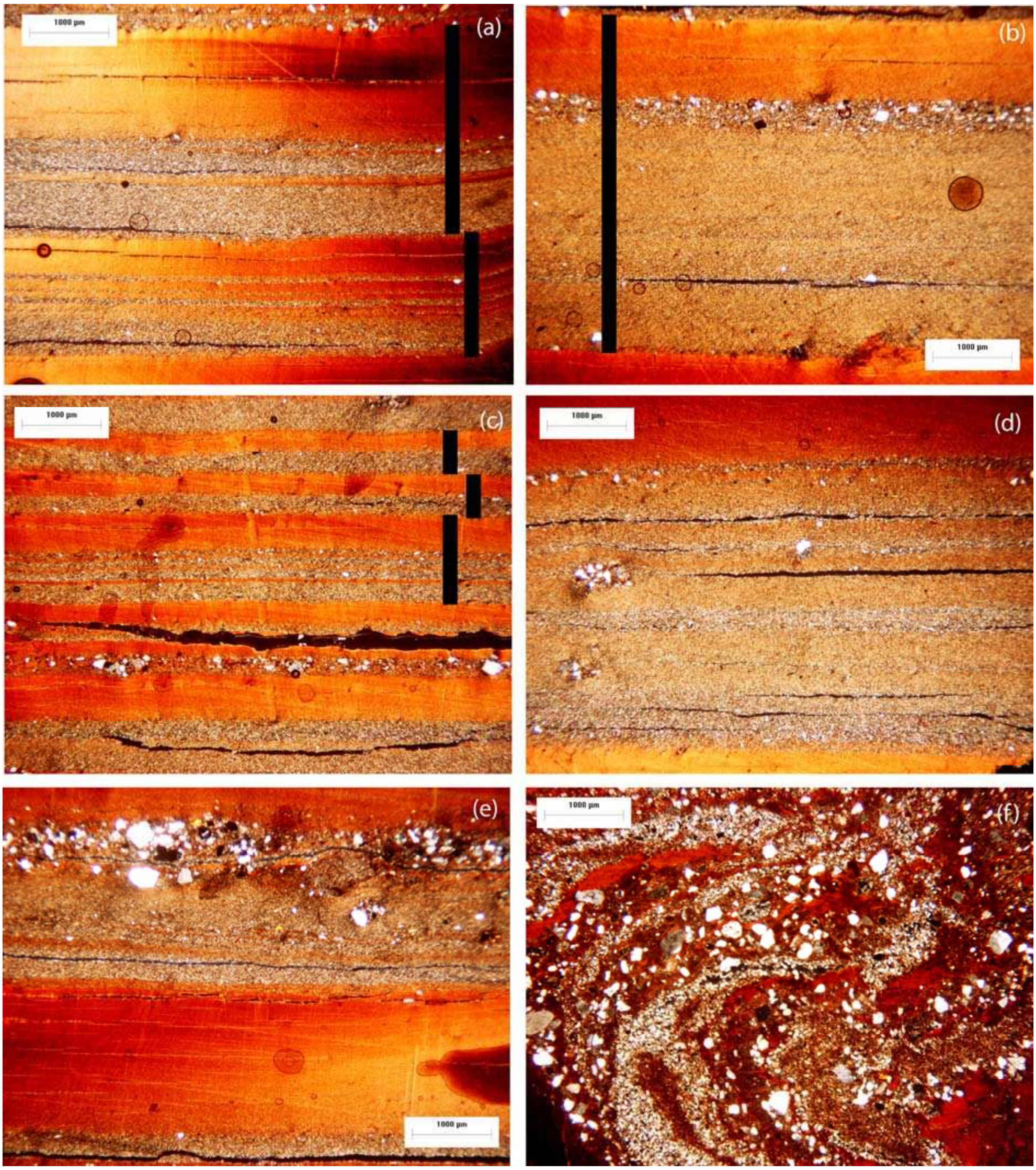


Figure 8: varve thickness and count

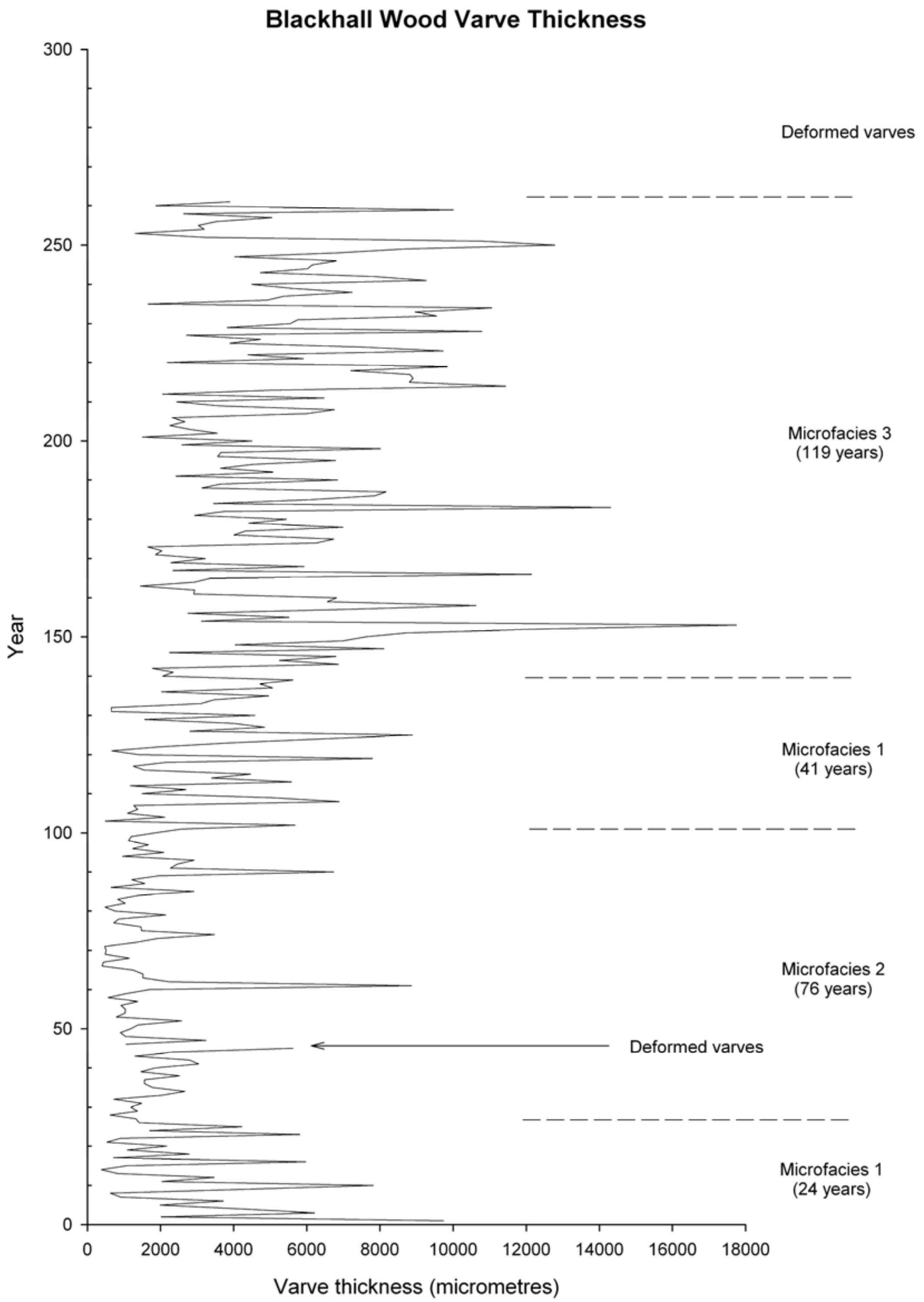


Figure 9: Maryport section logs and fabrics

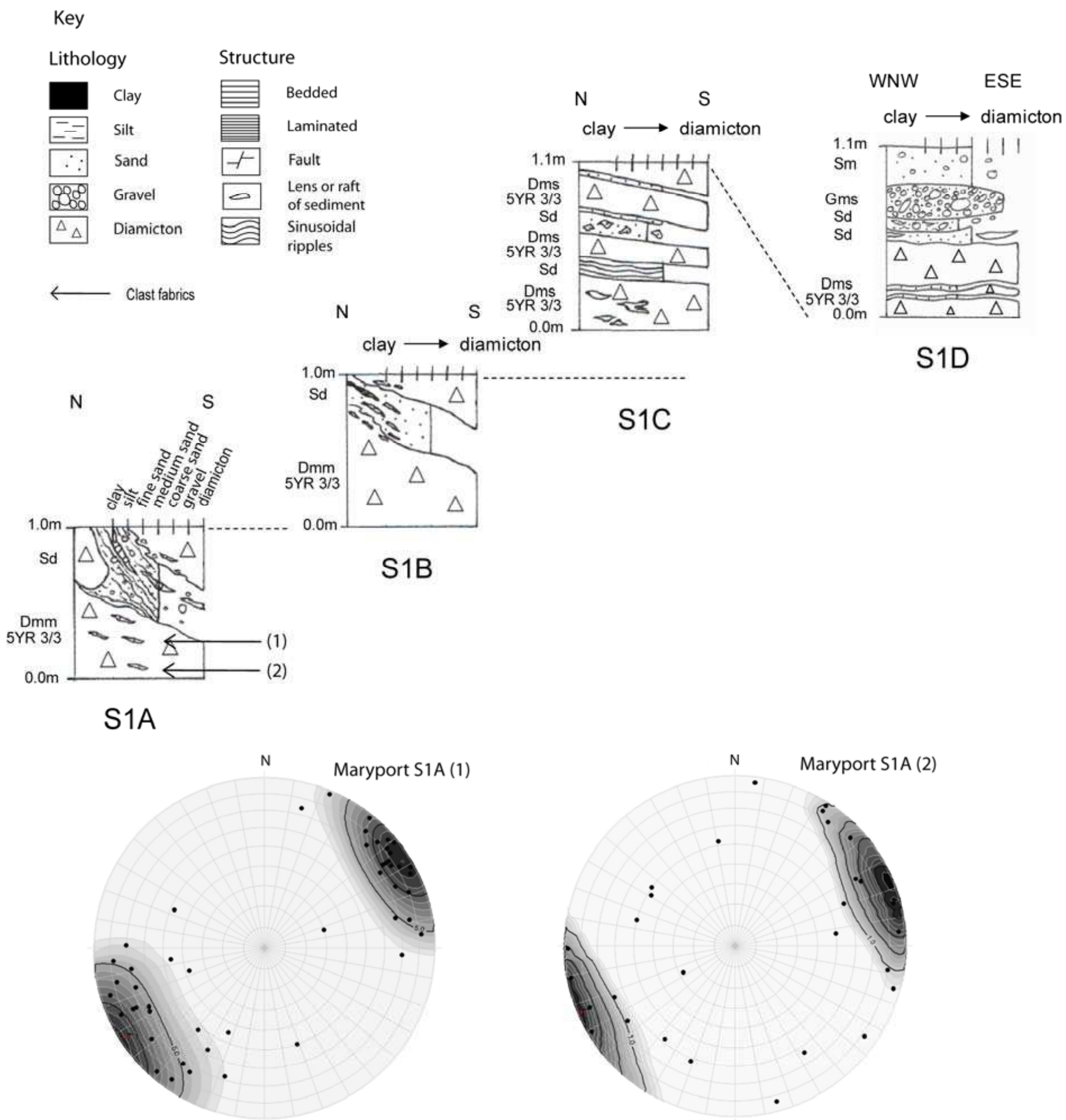


Figure 10: Photographs from Maryport and Swarthy Hill: (a) Maryport: diamicton and sand folds; (b) Maryport: folded and interbedded sand; (c) Maryport: pebble-cobble gravel cluster; (d) Swarthy Hill: fine sand beds with clastic dykes leading into gravel beds; (e) Swarthy Hill: diamicton lithofacies interbedded with sand; and (f) Swarthy Hill: clastic dykes.


Dmm
Deformed sand
Dmm
Deformed sand
Dmm
S1A+B
(a)
Interbedded deformed sand and diamicton
S1C
(b)
(c)
Dmm (s)
Indistinct, sheared boundary
Coarse sand lens
Flame structure
Dewatering structures
S1B
(d)
Deformed sands with dewatering structures
Dmm (s)
Deformed sands with dewatering structures
S1B
(e)
(f)

Figure 11: Swarthy Hill section logs, sketch of clastic dykes and fabrics

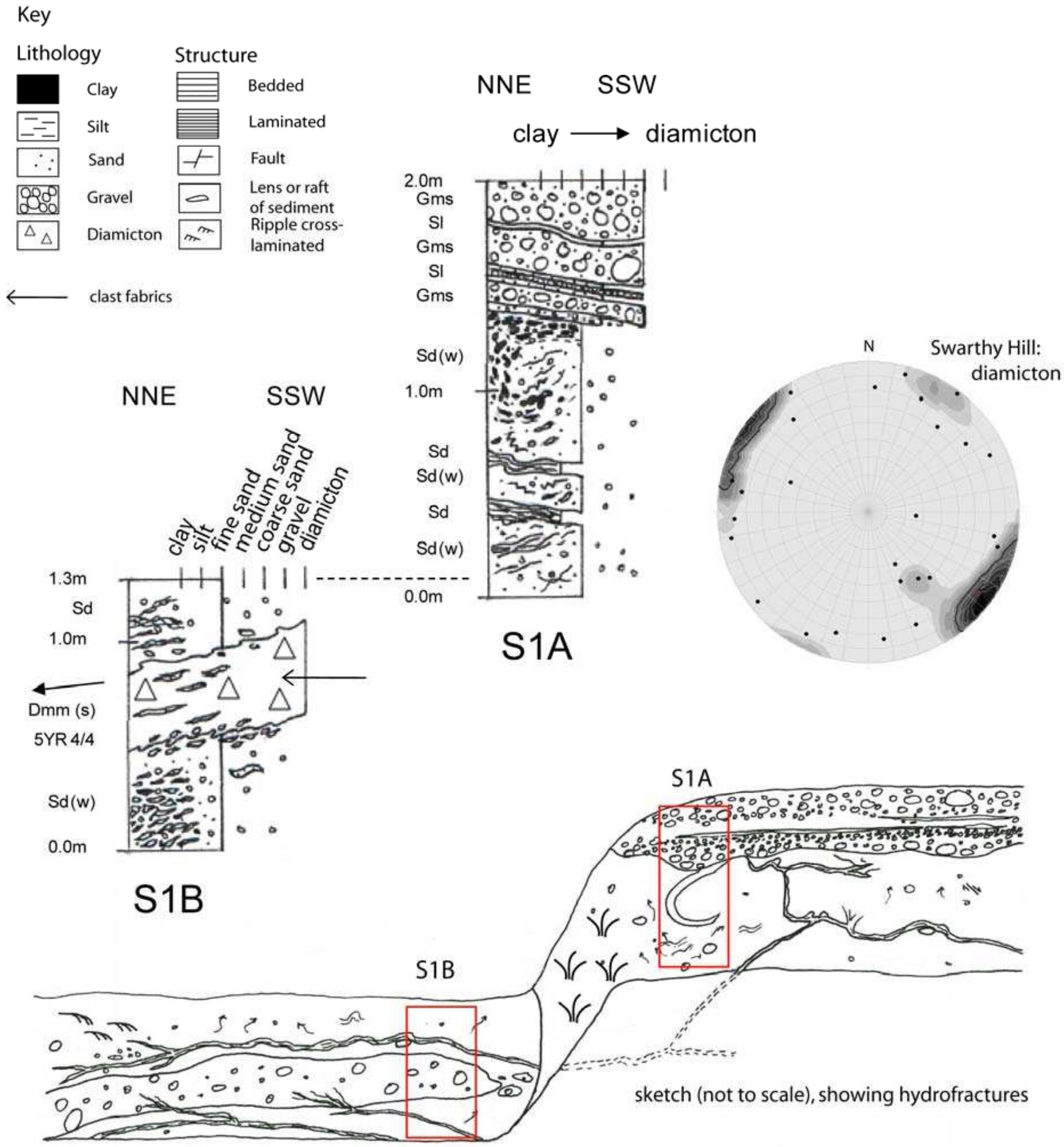


Figure 12: (a) Field sites, borehole locations and stratigraphic sections from prior research (Goodchild, 1875; Dixon *et al*., 1926; Huddart, 1970; Arthurton & Wadge, 1981); (b) contour map of the thickness of laminated clay and silt within the Vale of Eden (data from Fig. 11 and 13), mapped against topography; (c) contour map of the thickness of the upper diamicton (data from Fig. 11 and 13) mapped against the NEXTMap DEM.

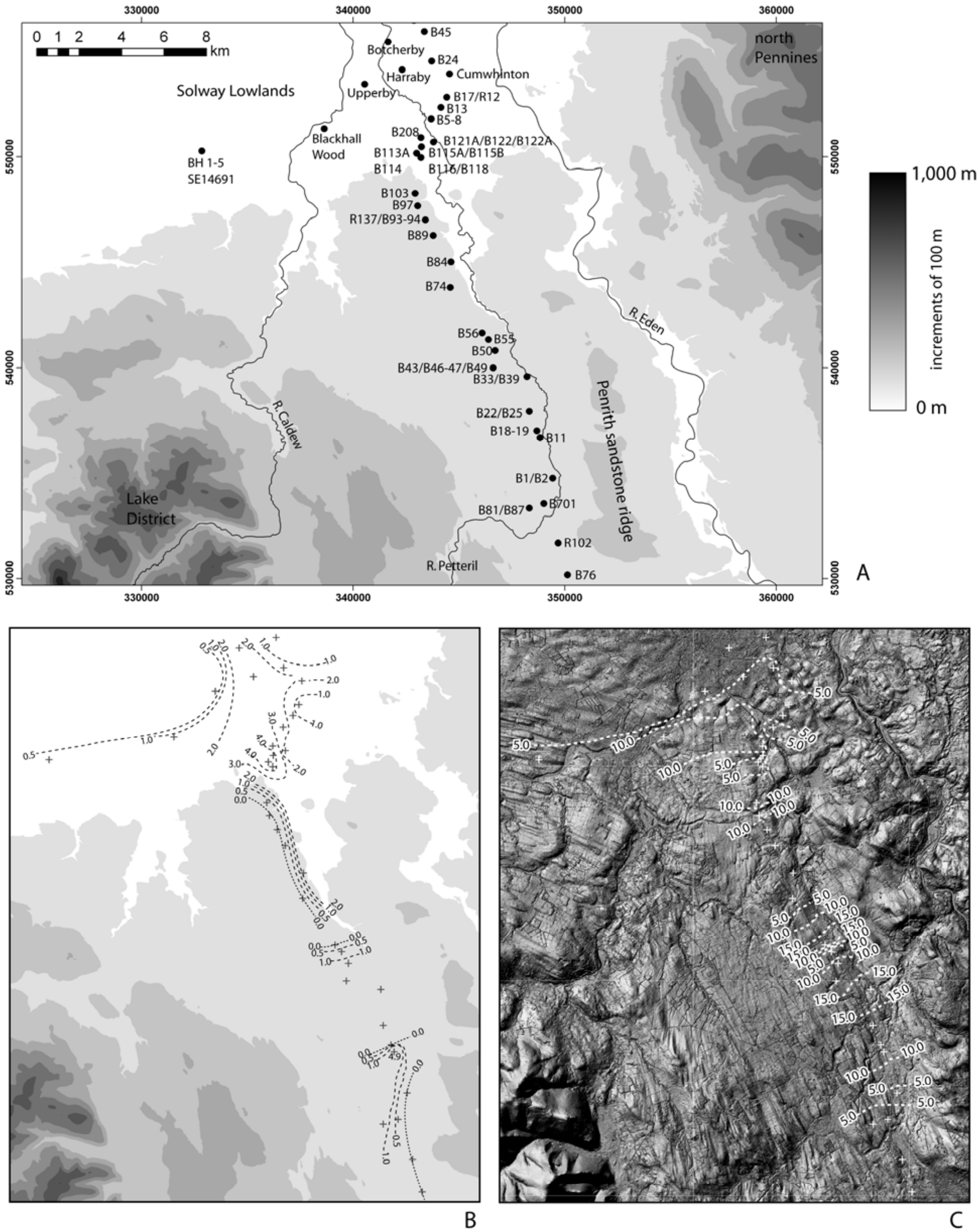


Figure 13: Stratigraphic logs from key sites through the Vale of Eden. See Fig. 12 for detailed location map of boreholes.

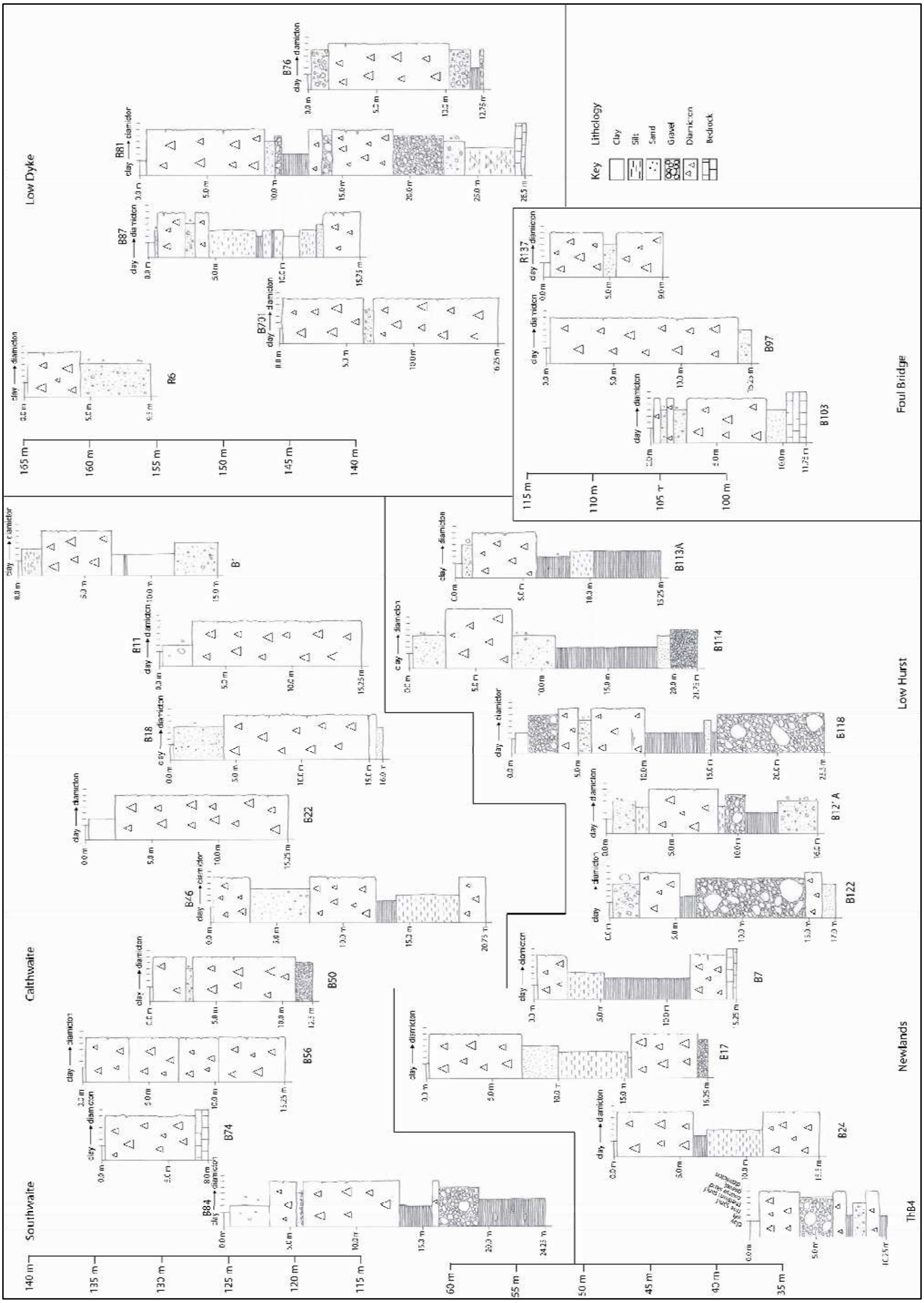
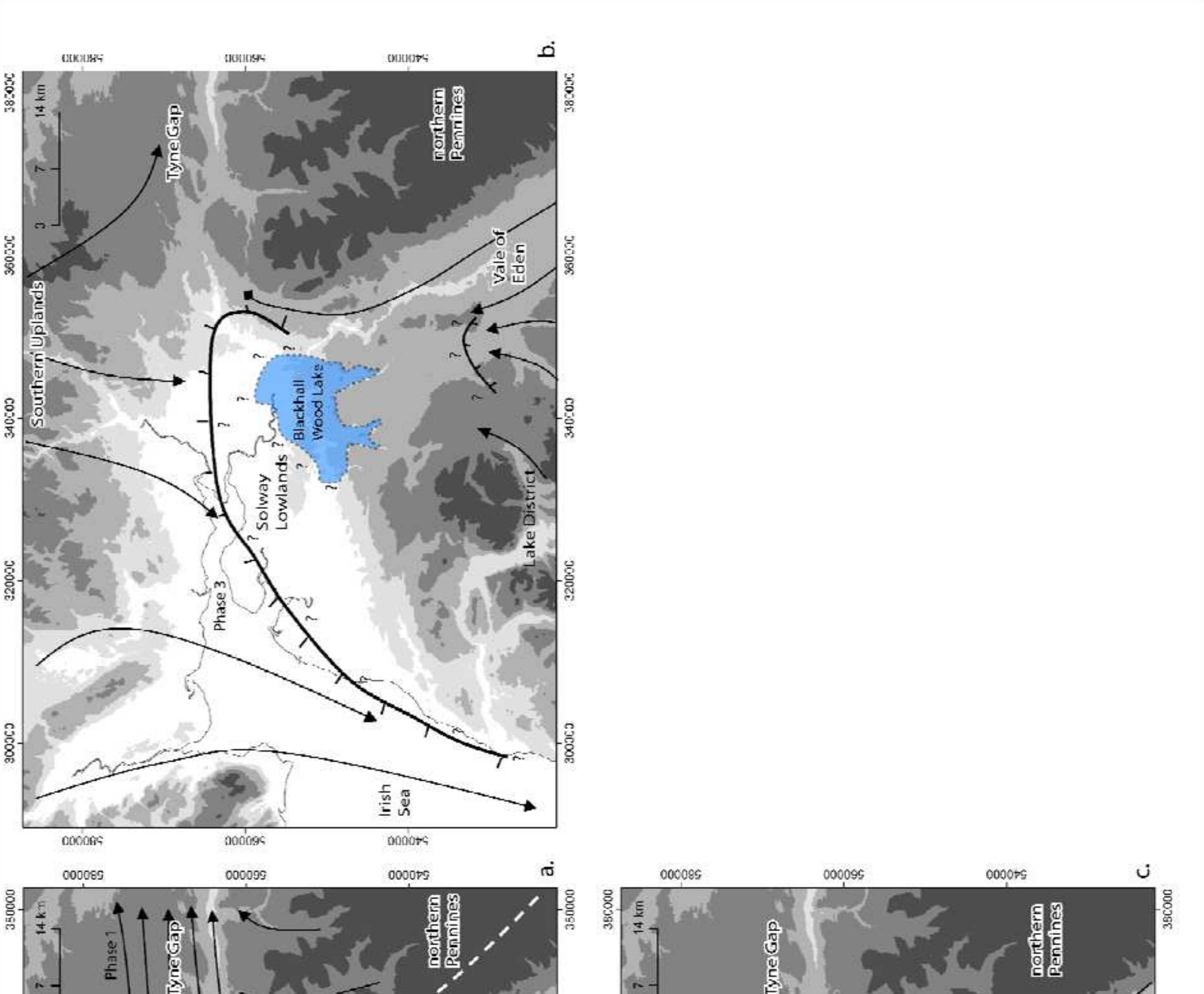

Low Dyke
B76
B81
B87
B701
R6
Key
Lithology
Clay
Silt
Sand
Gravel
Diamicton
Bedrock
Foul Bridge
R137
B97
B103
Low Hurst
B113A
B114
B118
B122
Calthwaite
B11
B18
B22
B46
B50
B56
Newlands
B7
B17
B24
Southwaite
B74
B84
TrB4
165 m
160 m
155 m
150 m
145 m
140 m
115 m
110 m
105 m
100 m
135 m
130 m
125 m
120 m
60 m
55 m
50 m
45 m
40 m
35 m

Figure 14: Glacial history of the 'Blackhall Wood' re-advance in the Solway Lowlands: (a) Prior to the 'Blackhall Wood' event, ice from Scotland and the Vale of Eden converged on the Solway Lowlands before moving eastwards through the Tyne Gap (phase 1). Phase 2 records a switch in direction, with ice instead moving westwards out into the Irish Sea; (b) ice flow phase 3 documents the deglaciation of the Solway Lowlands, leading to an ice free enclave, with ice still active in the Irish Sea Basin. Proglacial Lake 'Blackhall Wood' developed during this stage (although the northern, western and eastern extents are unconstrained). The lake is produced purely on the basis of the contour map in Fig. 12b, and is likely to have been significantly larger. The ice free enclave has been traced to the western edge of Cumbria and to the border of Scotland. However, again this is a minimum extent; and (c) Ice from both the Lake District and Scotland re-advanced out into the Celtic Basin as a tributary of the Irish Sea Ice Stream. (Topographic map (100 m intervals) overlaid onto a NEXTMap DEM, with lighting from a bird's eye view).


a.
Southern Uplands
Tyne Gap
Phase 1
northern Pennines
Vale of Eden
Solway Lowlands
Phase 2
Lake District
Irish Sea
14 km
b.
Southern Uplands
Tyne Gap
northern Pennines
Vale of Eden
Blackhall Wood Lake
Solway Lowlands
Phase 3
Lake District
Irish Sea
14 km
c.
Southern Uplands
Tyne Gap
northern Pennines
Vale of Eden
Solway Lowlands
Lake District
Irish Sea
14 km

Tables:

Table 1: Simple Event Stratigraphy for the field area (based on Trotter, 1929; Hollingworth, 1931; Huddart, 1970). Lithostratigraphic formations as used by the British Geological Society (Stone *et al*. in press). Regional correlations from: [1] McCabe *et al*. 1998; [2] Merritt & Auton (2000).

<table>
<tr><th rowspan="2">Event</th><th colspan="2">BGS Lithostratigraphic Formations</th><th rowspan="2">Regional Correlation</th><th rowspan="2">Date (cal ka BP)</th></tr>
<tr><th>Cumbria</th><th>Dumfries-shire</th></tr>
<tr><td>Scottish re-advance</td><td colspan="2">Gretna Till Formation</td><td>Killard Point Stadial[1]</td><td>~ 16.8</td></tr>
<tr><td>Deglaciation (formation of Glacial Lake Carlisle)</td><td>Great Easby Clay Formation</td><td>Plumpe Sand and Gravel Formation</td><td></td><td></td></tr>
<tr><td rowspan="2">Main Glaciation</td><td rowspan="2">Greystoke Till Formation</td><td rowspan="4">Chapelknowe Till Formation</td><td>Gosforth Oscillation[2]</td><td>~ 19.5</td></tr>
<tr><td rowspan="3">Dimlington stadial</td><td rowspan="3">Main Late Devensian</td></tr>
<tr><td colspan="2">Middle Sands</td></tr>
<tr><td>Early Scottish advance</td><td>Gillcambon Till Formation</td></tr>
</table>

Table 2: Average clast lithology (%) from Blackhall Wood, Maryport and Swarthy Hill, 8-64 mm. Carrock Fell Gabbro and Skiddaw Granite are sourced from the Lake District and Criffel, Dalbeattie and Loch Doon Granite are sourced from the Southern Uplands. Lava's from the Borrowdale Volcanic Group are sourced from the Lake District.

	Clast lithology	*Blackhall Wood*			*Swarthy Hill*	*Maryport*
		LF1	*LF2*	*LF5*		
Metamorphic	Slate	1.19	0.28	3.53	1.64	0.79
	Schist	1.58	0.57	0.39	1.64	0.26
	Brown Quartzite	0.40	0.85	1.96	1.97	1.59
	Red Quartzite	0.40	0.28	0.00	0.33	0.00
	White Quartzite	0.00	0.57	0.00	2.62	2.65
Igneous	Carrock Fell Gabbro	0.00	0.00	0.39	0.00	0.26
	Skiddaw Granite	0.00	0.00	0.00	1.97	0.53
	Criffel Granite	0.40	0.85	1.18	0.00	0.00
	Dalbeattie Granite	1.98	1.98	0.78	1.97	1.59
	Loch Doon Granite	0.40	0.57	0.00	0.00	0.00
	Rhyolite	0.40	1.13	0.78	1.64	1.59
	Andesite	0.00	1.13	0.00	0.00	0.00
	Brown Andesite	0.00	0.00	0.00	0.00	0.53
	Borrowdale Volcanic Lava's	0.40	2.83	0.78	10.16	6.35
	Basalt	0.00	0.57	0.39	0.66	1.32
	Dolerite	8.30	15.30	7.06	4.26	6.61
	Diorite	0.00	0.00	0.00	0.00	2.12
	Felsite	2.37	1.13	1.18	1.31	1.85
	Porthyry	0.00	0.57	0.00	0.00	0.53
	Unidentified Granite	1.98	1.70	0.78	0.98	1.85
Sedimentary	Silurian Greywacke	30.83	30.31	40.00	14.75	23.54
	Permo-Triassic Sandstone	9.49	6.23	4.71	1.64	9.26
	Sandstone	12.65	5.38	14.51	32.13	13.49
	Quartzitic sandstone	1.58	2.27	1.18	2.95	6.35
	Old Red Sandstone	3.56	2.83	4.71	0.98	0.00
	Shale	0.79	0.28	0.00	0.00	0.00
	Magnesium Limestone	0.00	0.28	0.39	0.00	0.00
	Carboniferous Limestone	14.62	15.86	12.94	0.00	0.00
	Mudstone	3.16	1.98	0.39	0.00	1.85
	Siltstone	0.00	1.13	0.78	5.57	9.52
	Breccia	0.00	0.28	0.00	0.00	0.00
Mineral Veins	Quartz	1.98	0.28	1.18	0.00	1.59