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Re-advance of Scottish ice into the Solway Lowlands (Cumbria, UK) during the Main Late Devensian deglaciation

SJL, COC, DJAE

Abstract:

Introduction:

A paradigm shift in geomorphological understanding related to palimpsest ice flow signatures (e.g. Boulton & Clark, 1990a, b; Punkari, 1993; Clark, 1997, 1999) coupled with a greater accessibility and higher resolution of DEMs has led to a re-appraisal of the British-Irish Ice Sheet (BIIS) during the Late Devensian (marine isotope stage 2) glaciation. This has resulted in a greater appreciation of the dynamism and sensitivity of the BIIS, with evidence for major shifts and switches in ice flow, rapid streaming deep inland of the ice margin, and multiple oscillations and re-advances (Salt & Evans, 2004; Greenwood & Clark, 2008; Livingstone *et al*., 2008; Evans *et al*., 2009). The Solway Lowlands in Cumbria (Fig. 1) were situated at the heart of the BIIS, and as such, were affected by the complex interplay between multiple upland ice dispersal centres, including the north Pennines, Lake District and Southern Uplands, the migration of ice divides and the impact of the Irish Sea and Tyne Gap ice streams (Livingstone *et al*., 2008). This complexity is reproduced in the geomorphic and stratigraphic record, with previous attempts to resolve the glacial history leading to a variety of glaciologically implausible scenarios, conflicting interpretations and unanswered questions (Goodchild, 1875, 1887; Trotter, 1929; Hollingworth, 1931; Huddart, 1971). This is especially true for the Scottish re-advance, an indistinct and controversial ice flow phase which is deemed to have impinged onto the Cumbrian coast during a late stage of the Main Late Devensian deglaciation (Trotter, 1922, 1923, 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1971a, b, 1991, 1994; Huddart & Tooley, 1972; Huddart *et al*., 1977, Huddart & Clark, 1994).

In light of recent developments in geomorphological mapping, the identification of several re-advances into the Solway Lowlands (Livingstone *et al*., in prep) and the recognition of multiple ice flow phases throughout the central sector of the BIIS (Livingstone *et al*., 2008; Evans, *et al*., 2009), it now seems pertinent to re-assess the evidence related to the Scottish re-advance in the Solway Lowlands, and indeed, how this relates to reconstructions of the BIIS more broadly. This paper utilises geomorphological mapping and sedimentological and stratigraphic techniques in order to address a number of research questions, namely: (a) What geomorphological and sedimentological evidence can be correlated to the re-advance of Scottish ice?; (b) What was the maximum extent of the Scottish re-advance?; (c) Was there was a concurrent re-advance of ice from the Lake District?; and (d) What does the evidence available tell us about ice behaviour during this re-advance?

Previous research on the Scottish re-advance:

Evidence for a glacial re-advance into the Solway Lowlands was first proposed by Trotter (1922, 1923, 1929). This evidence comprised a thin upper till overlying a series of sands and laminated clays, comprised of Scottish erratics (Dixon *et al*., 1926; Trotter, 1929; Trotter & Hollingworth, 1932). Further evidence in favour of a Scottish re-advance includes a number of NW-SE trending eskers at Thursby, Cummertree and Gretna, with outwash deltas also deposited in association with the eskers at Gretna (Dixon *et al*., 1926; Charlesworth, 1926; Trotter, 1929) (Fig. 2). Although no terminal moraines were recognised, Trotter predicted that the ice limit reached up to 134 m O. D. stretching as far east as Lanercost, Brampton and Cumwhitton and as far south as Foulbridge and Bolton Low Houses (Fig. 1 & 2). The thin till facies, lack of terminal moraines, and generally undisturbed sequences were construed as evidence for a short lived and transient re-advance phase (Trotter, 1929). During, and following, the maximum extension of the Scottish re-advance Trotter (1929) envisaged a number of glacially impounded lakes and overspill channels forming at successively lower levels, as the ice front retreated westwards. At the maximum extent Lake Carlisle and Lake Lyne formed against the reverse slope of the Tyne Gap, with water draining eastwards via the Gilsland meltwater channel (Trotter, 1929). As ice started to retreat westwards lower level lakes Caldew and then Wigton began to develop (Fig. 2), with meltwater escaping westwards via overspill channels which connected with, and drained into, successively lower lakes, depositing a series of deltas (Trotter, 1929).

The Scottish re-advance concept was re-evaluated by Huddart (1970, 1971a, b, 1991, 1994), Huddart & Tooley (1972), Huddart *et al*., (1977), Huddart & Clark (1994) and Huddart & Glasser (2002) with evidence gathered from the Solway Lowlands used to support a more limited phase of ice movement (Fig. 2). Huddart (1970) argued that the Scottish ice failed to extend far beyond Carlisle, with much of the stratigraphic evidence provided by Trotter (1922, 1923, 1929) believed to be patchy or capable of re-interpretation as debris flow deposits. Instead the re-advance limits were predominantly defined by esker deposits at Thursby, a thin upper till west of Carlisle and a major glaciofluvial deltaic complex at Holme St Cuthbert (Huddart, 1970, 1991, 1994) on the Cumbrian coastal fringe (Fig. 2). Lake development was acknowledged (Huddart, 1970, 1981, 1991), but attributed to a prior episode of deglaciation, whilst ‘meltwater overspill channels’ such as at Wampool and Wiza Beck (Fig. 2) were re-interpreted as subglacial.

Evidence for a Scottish re-advance is not confined to the Solway Lowlands. The St Bees push moraine situated on the west Cumbrian coast has been assigned to the Scottish re-advance (Fig. 2) by some researchers (e.g. Huddart, 1994; Merritt & Auton, 2000), although its origins remain controversial (cf. Merritt & Auton, 2000; Williams *et al*., 2001). The Bride Moraine on the Isle of Man 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 that the Scottish re-advance correlates with the maximal extent of the Killard Point Stadial re-advance (*ca*. 14 ^{14}C ka BP). However, these correlations must be treated with caution as there is no *direct* stratigraphic evidence or well constrained dates. Furthermore, the growing realisation that the Cumbrian coastal fringe was subject to multiple re-advances (Merritt & Auton, 2000; Livingstone *et al*., in prep) precludes against a common and simple cross-basin correlation.

The re-advances that have taken place within the Solway Lowlands include the Scottish re-advance, and also an earlier ‘Blackhall Wood re-advance’ (Livingstone *et al*., in prep). This re-advance has been correlated with the Gosforth oscillation (Merritt & Auton, 2000) at *ca*. 19.5 cal ka BP, and was characterised by major pro-glacial lake development in an ice free enclave, followed by re-advance and re-connection with the Irish Sea Ice Stream (Livingstone *et al*., in prep).

Due to the rather meagre evidence for a Scottish re-advance into the Solway Lowlands some researchers have questioned its validity (Pennington, 1978; Evans & Arthurton, 1973; Sissons, 1974; Thomas, 1985). Furthermore, an alternative ‘glaciomarine model’ proposed by Eyles and McCabe (1989) envisaged rising sea levels into an isostatically depressed Irish Sea Basin causing the rapid retreat of the Irish Ice Stream and resulting in marine limits up to 140 m O. D., thick wedges of ice-contact glaciomarine sediments, raised deltas and mud drapes. This concept largely ignores or re-interprets the evidence for a terrestrial Scottish ice re-advance in the Solway region, with geomorphic and stratigraphic information such as the Holme St Cuthbert complex and upper till instead ‘fitted’ into the glaciomarine model (cf. Huddart, 1994).

Methods:

Glacial geomorphological mapping:

Geomorphological mapping involved the recognition and compilation of discrete landform assemblages from NEXTMap digital elevation model (DEM) data. This is a 5 m spatial resolution DEM derived using airborne interferometric synthetic aperture radar (http://www.neodc.rl.ac.uk/). In addition, maps of the bedrock geology and superficial deposits (DiGMapGB-625 downloaded from the BGS) were overlain onto the imagery. Glacial landforms were mapped manually using on screen digitisation. Vectors were used to digitise lineations, meltwater channels and eskers. Polygons were used to digitise hummocky moraine, ribbed moraine, glaciofluvial sediment accumulations and the break of slope exhibited by subglacial lineations (cf. Livingstone *et al*., 2008). Long profiles were derived along meltwater channels in order to assess their mode of origin. Lineations were sub-divided into ‘flow-sets’ based on their morphology, length and parallel conformity (Clark, 1997, 1999). This allowed overprinting relationships to be identified and a relative chronology of ice flow phases to be constructed (Clark, 1999; Livingstone *et al*., 2008). Ground-truthing was carried out in areas of complexity to verify the mapped glacial landforms, or to substantiate morphological features described in the literature but not observed within the spatial resolution of the DEM. Stoss and lee forms identified from the geomorphological mapping, coupled with published erratic pathways (e.g. Goodchild, 1875; Trotter, 1929; Hollingworth, 1931), were used to interpret flow directions throughout the region.

Sedimentology and stratigraphy:

Field sites were identified throughout the Solway Lowlands (Fig. 1). This included Overby and Aldoth sand pits in Holme St Cuthbert sand and gravel complex, a site on the Scottish border at Plumpe Farm in an easterly orientated flow set and a site exposed at Carleton in hummocky terrain. Texture, sedimentary structure, colour, bed geometry, contacts and inclusions were all measured and logged, from which lithofacies were identified (Evans & Benn, 2004). 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 facies could be assessed. Palaeocurrent indicators (such as ripples and imbricate gravel), clast macrofabric analysis of the a-axis orientation and dip, and clast lithologies augmented the sedimentary logging. Borehole logs (BGS) provided a less detailed but wider coverage, allowing regional stratigraphic correlations.

Results and Interpretation:

1. Holme St Cuthbert sand and gravel complex:

Geomorphology:

Holme St Cuthbert (NY 105, 470) is a 9 km^2 spread of sand and gravel situated amongst heavily lineated glacial terrain of the Cumbrian lowlands (Fig. 1 & 3). It stretches *ca*. 2 km NW-SE between Lowsay Farm and Hards Farm, *ca*. 4 km NE-SW between Aldoth and New Cowper, and reaches heights of up to 50 m O. D. (Fig. 3). The complex is superimposed over a strongly lineated NE-SW orientated flow set which wraps arcuately around the northern-most edge of the Lake District before trending out into the Irish Sea Basin (Fig. 3). The Holme St. Cuthbert complex itself consists of two distinct morphological assemblages:

1. A SE facing scarp ridge which runs between New Cowper and Round Hill (Fig. 3). This slope forms the eastern extremity of a flat topped ridge running NE-SW that is between 0.5 to 1 km in width, up to 50 m O. D. and ends on the western side with another steep scarp (figure 5).

2. To the west of this scarp is undulating terrain containing a series of depressions and two subdued and winding ridges orientated NW-SE (Fig. 3).

Site 1: Overby:

Overby sand pit is situated within the flat-topped ridge (NY 125, 471) of the Holme St Cuthbert sand and gravel complex (Fig. 1 & 3). The succession has been divided into three main lithofacies associations described and interpreted in turn below, and displayed diagrammatically in Fig. 4.

Lithofacies Association 1 (LFA OV1): Description:

LFA OV1 consists of the finest-grained sediments exposed throughout the sand pit (Fig. 3). It is characterised by laterally extensive sheets of 0.1 - 1.2 m thick fine-coarse sands which dip gently (4°) towards the east and are exposed in the bottom 12 - 15 m of the sand pit (Fig. 6a). Lithofacies which comprise LFA OV1 include, horizontally laminated fine-coarse grained sand containing occasional granule gravel, thin sheets (0.1 – 0.2) of massive sand, normally graded sand, sinusoidal ripples, type A ripple drift-cross laminations, and type A and B climbing ripples (Allen, 1968, 1973; Jopling & Walker, 1968). The type A and B ripples identified in each of the logged exposures indicate a palaeo-current direction which varies between the east and SSW (Table 1). The lithofacies tend to alternate through horizontal laminations, or massive sand lithofacies, into type A/B climbing ripples (Fig. 6a). Sinusoidal ripples and silt/clay bands are infrequently observed generally at the top of the type A/B climbing ripples. Also identified were occasional thin (less than 0.2 m) granule gravel sheets, outsized pebble gravel, and clusters of small (*ca*. 0.5 m thick by 1-2 m wide) trough and lenticular shaped, clast-supported, pebble gravel lithofacies characterised by erosional bases and upwards fining, commonly found towards the top of LFA OV1.

LFA OV1: Interpretation:

The predominance of fine-grained sediments, including rippled and laminated sand suggests that deposition occurred primarily in a low energy environment. Laminated fine sand and silt/clay bands are deposited from suspension settling, whilst coarser grained medium-coarse sand laminae and rippled sand indicate deposition by low energy traction currents or density underflows (Jopling & Walker, 1968; Reineck & Singh, 1975; Smith & Ashley, 1985). Massive and normally graded sand is likely to have been deposited via sediment gravity flows supported by turbidity currents. The complex facies architecture, consisting of cyclical assemblages, is envisaged to have resulted from a fluctuating hydrograph (Smith & Ashley, 1985). This is best exemplified by the progressive changes from coarse/medium sand and granule gravel laminae, through type A/B climbing ripples into sinusoidal ripples which marks a reduction in flow velocity and increase in deposition by suspension fall-out (Jopling & Walker, 1968; Allen, 1973; Smith & Ashley, 1985). The variety of palaeo-current orientations demonstrates changing influx points, although with flow generally moving south-easterly. Rare scour structures towards the top of LFA OV1 composed of clast-supported normally graded pebble gravel record rapid cut-and-fill processes resulting from turbidity currents (Miall, 1977, 1985; Collinson & Thompson, 1989; Nemec, 1990) and caused by a hydraulic jump.

Lithofacies Association 2 (LFA OV2): Description:

LFA OV2 comprises 14 m of steeply dipping planar cross-beds of stratified medium-coarse sand with granule gravel and some out-sized pebble gravel, situated stratigraphically above LFA OV1 in exposures along the entire 60 m NE face of the sand pit (Fig. 3 & 6b). A series of distinctive units up to 10 m thick are observed within LFA OV2, identified by sharp erosional lower surfaces and dips which range between 6 - 40° towards the SE (with the exception of one unit which dips gently (6°) towards the NW). Some of these units also exhibit slight fining upward sequences from very gravelly coarse sand and granule gravel into medium-coarse sand, whilst it was observed that the angle of dip generally decreased towards the base of the LFA (Fig. 6b).

LFA OV2: Interpretation:

The dipping, stratified lithofacies of gravelly medium-coarse sand are typical of foreset beds formed by cohesionless debris flows down a slope face (Smith & Ashley, 1985; Nemec, 1990; Nemec *et al*., 1999). The presence of distinctive units characterised by fining upwards, re-activation surfaces and a variety of dips documents changes in the sediment delivery over various timescales either related to changes in the discharge regime or changes in the sediment influx point (e.g. Gustavson *et al*., 1975; Clemmensen & Houmark-Nielsen, 1981). The general trend of the foreset structures indicates sediment progradation to the SE. The unit dipping in the opposite direction, down to the NW, could be a backset bed formed by very turbulent water (Reineck & Singh, 1975) causing transient reversals of flow (e.g. Clemmensen & Houmark-Nielsen, 1981; Nemec *et al*., 1999).

LFA OV3: Description

The top-most succession (LFA OV3) is 0.5 – 1.5 m thick and exhibits a series of matrix supported cross-stratified, planar stratified and massive coarse sand and granule gravel lithofacies (Fig. 3). Each lithofacies is no more than 0.5 m thick, with erosional lower contacts often related to thin (< 0.1 m), discontinuous pebble-gravel lags (Fig. 6c). The cross-stratified coarse-sand and granule gravel lithofacies dip south-easterly in the same direction as the foreset structures (LFA OV2). At the base of LFA OV3 are a number of coarse sand dykes, trending down into LFA OV2, whilst sheared-up blocks of fine-medium sand have also been observed in LFA OV3 (Fig. 6c). The top 1 - 2 m of both LFA OV2 and LFA OV3 also show evidence of deformation, with occasional small scale convolutions recorded.

LFA OV3: Interpretation:

The lithofacies of LFA OV3 are interpreted to be fluvial in origin, with the cross-bedded stratified sand characteristic of dune or bar forms (Miall, 1977, 1985; Smith, 1985; Collinson & Thompson, 1989). The planar stratified coarse sand and granule gravel were deposited by high energy traction carpets (Miall, 1977, Allen, 1984) possibly as bar forms, whilst the massive coarse sand, granule and pebble gravel lithofacies were deposited either during a high energy, high density event (Collinson & Thompson, 1989) or as a debris flow. The pebble gravel lags mark channel floor deposits, with the erosional reactivation surfaces within LFA OV3 demonstrating discrete depositional events possibly related to a migratory fluvial system (Miall, 1977). The strong erosive contact and marked shift in grain size between LFA OV2 and OV3 indicates a distinct shift in fluvial regime. The dykes of coarse sand are inferred to have been produced by hydrofracturing (e.g. Rijsdijk *et al*., 1999; Van der Meer *et al*. 1999, 2009), with the downward tapering and infilling from LFA OV3 demonstrating that water escaped and dissipated into LFA OV2.

Site 2: Aldoth:

Aldoth sand pit (NY 148, 483) is situated in the north-eastern corner of the sand and gravel complex in a small wedge of sand (up to 42 m O. D.) which protrudes from the main ridge (Fig. 1, 3). The succession has been divided into three main lithofacies associations described and interpreted in turn below, and displayed diagrammatically in Fig.7.

LFA A1: Description

The 5 - 7 m thick LFA A1 is exclusively found in the north-eastern corner of the sand pit, and is composed of laterally extensive 0.1 – 1.5 m thick sheets of massive and normally graded clast-supported gravel lithofacies interbedded with thinner (0.1 – 0.5 m) sheets of horizontally laminated, massive and cross-stratified sand (Fig. 6d & 7). The sand lithofacies become more prevalent towards the top of the LFA. The gravel is generally pebble sized with some rare cobbles and sub-rounded to rounded clasts (Fig. 6d). The beds have sharp contacts and are horizontally organised.

LFA A1: Interpretation:

The poorly sorted gravel lithofacies are thought to have been deposited as cohesionless debris flows (Shanmugam, 2000), whereas the normally graded gravel beds probably formed by high density turbidity currents (Lowe, 1982; Collinson & Thompson, 1989). The interbedded sand lithofacies represent waning flow conditions leading to the development of dune forms (Miall, 1977), traction carpets associated with high density turbidity currents (Miall, 1977, Allen, 1984) and debris flows. Indeed the propensity for high density turbidity currents and debris flows would suggest that LFA A1 was deposited in a subaqueous environment (cf. Shanmugam, 2000).

LFA A2: Description

LFA A2 was observed towards the southern end of the sand pit, with the approximately 3 m thick sequence consisting of thin (< 0.5 m) sheets of diffusely stratified granule gravel and sand, sinusoidal ripples, infrequently observed and isolated type A ripples, fine sand laminations and occasionally imbricated, normal and inversely graded granule-pebble gravel (Fig. 6e & 7). The bottom 0.3 cm is comprised mainly of fine-medium sand characterised by wavy bedding, ripples and pockets of granule-pebble gravel, which grade up into granule-gravel and gravelly sand dominated beds (Fig. 6e). The beds are generally horizontal to sub-horizontal with both sharp and diffuse boundaries observed and evidence of deformation in the form of centimetre-scale convolutions, and sheared, displaced blocks of sand (Fig. 6e).

LFA A2: Interpretation:

LFA A2 is envisaged to have formed by tractional deposition related to high and low density turbidity flows in a subaqueous environment (e.g. Lowe, 1982; Shanmugam, 2000). Stratified granule gravel and sand, with evidence of both inverse and normal grading, demonstrates deposition by a traction carpet (Lowe, 1982; Kneller, 1995; Shanmugam, 2000), whilst the sinusoidal ripples and massive sand lithofacies are formed via suspension rain-out (Smith & Ashley, 1985). Imbricate gravel gives a south-easterly palaeo-current, whilst the type A ripples are orientated both towards the SE and the NW. The isolated examples of flow reversal demonstrated by the ripple structures could relate to backsets produced in very turbulent water (Reineck & Singh, 1975). Displaced blocks of sand are attributed to scouring related to a hydraulic jump, followed by rapid deposition (cf. Russell & Arnott, 2003).

LFA A3: Description

LFA A3 is composed of bands of massive and normally graded fine-coarse sand, silt and clay, with infrequent examples of trough cross-bedded sand and pebble gravel lithofacies (Fig. 6f & 7). The bands of sand are sub-horizontally orientated, laterally extensive (> 10 m) and between 0.1 – 0.5 m thick (Fig. 6f). They also exhibit fining towards the southern end of the quarry. This LFA, which reaches thicknesses of up to 3.0 m, is observed in the south-western corner of the quarry and also overlying LFA A1 and A2. Centimetre-scale convolutions, faults and sheared boundaries in conjunction with a 1.5 m pillar structure composed of heavily faulted sand detail evidence of deformation.

LFA A3: Interpretation:

Massive bands of sand, silt and clay probably reflect deposition by suspension rain-out and density underflows in a low energy environment (Reineck & Singh, 1975; Smith & Ashley, 1985). Normally graded sand bands reflect waning flow conditions resulting in a loss of competence whilst massive sand bands reflects rapid deposition by a loss of capacity (Reineck & Singh, 1975). Lateral variations in grain size are attributed to downstream fining. The trough cross-bedded pebble-gravel lithofacies indicate rapidly formed cut-and-fills produced by transient high energy scouring (Miall, 1977, 1985).

<u>Synopsis: environment of deposition:</u>

Holme St Cuthbert sand and gravel complex is interpreted as an ice-contact, Gilbert style, delta (e.g. Price, 1973) which evolved from a subaqueous grounding line fan (e.g. Powell, 1990; Nemec *et al*., 1999; Thomas & Chiverrell, 2006).

The LFA's of Aldoth sand pit are envisaged to have been deposited as a subaqueous outwash fan (e.g. Rust & Romanelli; Cheel & Rust, 1982; Powell, 1990; Gorrell & Shaw, 1991; Plink-Björklund &

Ronnert, 1999; Russell & Arnott, 2003; Winsemann *et al*., 2007). The exposures are located in a wedge of sand and gravel jutting out from the main body of the delta, with a morphology not dissimilar to an esker-bead (e.g. Bannerjee & McDonald, 1975; Gorrell & Shaw, 1991). Coarse-grained sand and gravel (LFA A1) deposited by traction carpets and cohesionless debris flows in the zone of flow establishment (ZFE) (Russell & Arnott, 2003) grade rapidly into better sorted and finer grained lithofacies (LFA A2) deposited by episodic turbidity currents (Cheel & Rust, 1982; Postma *et al*., 1983) and found in the ZFE or zone of transition flow (ZTF) (Russell & Arnott, 2003). LFA A3 documents more distal sedimentation, found in the ZTF or zone of established flow (Russell & Arnott, 2003), with suspension rain-out now the dominant depositionary mechanism.

The LFAs at Overby sand pit exhibit all the classical genetic features of a Gilbert-style delta (cf. Smith & Ashley, 1985). The fine grained sediments of LFA OV1 are typical of bottomsets, deposited in a low energy environment (Jopling & Walker, 1968; Gustavson *et al*., 1975; Cohen, 1979; Clemmensen & Houmark-Nielsen, 1981) distal to the influx point, by both underflows (ripple structures) and suspension rain-out (clay/silt bands and fine sand laminations). The rhythmicity exhibited by the deposits is a function of fluctuating discharges (Smith & Ashley, 1985), possibly on an annual scale (Gustavson *et al*., 1975). LFA OV2 displays many of the features of Gilbert-style deltaic foresets, including dipping beds of gravitationally deposited coarse sands and granule gravel, which become easier angled lower down in the sequence (Nemec *et al*., 1999), and distinctive units which demonstrate shifts in the influx point (Smith & Ashley, 1985). The fluvial sediments of LFA OV3 are attributed to topsets, deposited on the delta surface by a migratory river system (Nemec *et al*., 1999).

An ice-contact origin for the delta has been proposed, based on the following lines of evidence: (a) the undulatory terrain to the NW of the main ridge, which is interpreted as kame and kettle topography composed of glaciofluvial sediments (Cook, 1946; Holmes 1947; Paul 1983) and comprising a series of NW-SE trending eskers and kettle holes (e.g. Price, 1969); (b) the steep NE-SW orientated ridge which runs along the edge of the kame and kettle topography demarcating the position of the ice-margin; and (c) the formation of a subaqueous fan which must have emanated from a grounding line along the ice front (Powell, 1990).

Palaeo-current indicators, clast lithological analysis and the glacial geomorphology all suggest that ice must have flowed south-easterly out of Scotland and across the Solway Firth. Ripple structures (Table 1) are orientated towards the E-SW, whilst the subaqueous outwash fan fines southwards. Clast lithological analysis contains a strong signal from the Southern Uplands (11% Criffell and Dalbeattie granite and 56% Silurian greywacke), whilst the position of the ice contact slope in relation to the kame and kettle topography indicate that ice must have been situated to the NW of the delta. This interpretation is consistent with the genesis of a glacial lake which must have formed as westerly drainage of meltwater into the Irish Sea Basin and Solway Firth became blocked by ice.

2. Stratigraphy:

Site 1: Plumpe Farm

The exposures at Plumpe Farm (NY 334 681) are situated just to the east of the Scottish border amongst subdued and low-lying (< 40 m O. D.) glacial lineations orientated W-E (Fig. 1). The succession has been divided into two main lithofacies described and interpreted in turn below, and displayed diagrammatically in Fig. 8. These lithofacies are underlain by a lower red-brown diamicton not exposed at Plumpe Farm (Phillips *et al*., 2007).

Lithofacies (LF PL1): Description:

The basal lithofacies comprises up to 3.2 m of compact, red fine-medium sand interbedded with a series of silt and clay bands (Fig. 8 & 9a). The sand is generally massive or normally graded, with individual beds up to 1.3 m thick (Fig. 9a). Towards the top of the lithofacies the unit becomes locally laminated, exhibiting alternating clay and fine sand bands up to 2 cm thick (Fig. 9a). Exposure 2 displays little in the way of deformation, with some shearing between boundaries, evident in the upper half of the unit. However, exposure 1 displays a series of centimetre-scale deformation structures including normal and reverse faults, sheared contacts and convoluted clay and silt bands (Fig. 9b). The extent of deformation is noted to increase upwards through the lithofacies.

LF PL1: Interpretation:

The sediments of LF PL1 are thought to have been deposited in a glaciofluvial or more likely, given the fine-grained nature of the deposits, glaciolacustrine environment, formed by suspension rain-out (clay/silt bands) and density underflows (sand facies) (Reineck & Singh, 1975). Evidence of deformation, typified by an up-profile increase in shear strain and comprising simple shear and vertical displacement (loading) structures (cf. McCarroll & Rijsdijk, 2003), suggests that some minor glaciotectonisation occurred (e.g. Hart & Boulton, 1991; Evans, 2000). This was mainly concentrated in the upper 0.5 m of the lithofacies and across the boundary with LF PL2. The lack of any major glaciotectonisation structures implies that conditions at the ice-bed interface acted to reduce the transmission of shear strain down into the sequence, possibly due to a water lubricated surface (cf. Phillips *et al*., 2007).

LF PL2: Description:

The uppermost lithofacies consists of a dark red-brown (5.0 YR 3/3) matrix-supported diamicton up to 0.9 m thick. The diamicton contains a high frequency of sub-rounded to sub-angular clasts with some striae. The lower contact with LF PL1 is gradational, with sheared, fine-grained sands incorporated into the lower 20 cm of the diamicton. Sand and granule gravel pods are also evident within the diamicton, whilst the upper 20 cm is heavily mottled. The principle lithologies are greywacke (59%), siltstone (18%), quartzitic sandstone (7%) and the local Permian and Triassic sandstone (6%). Clast macro-fabric data (Fig. 8) reveals a W-E orientation, with an S_1 eigenvalue of 0.59.

LF PL2: Interpretation:

LFA PL2 is interpreted as a subglacial traction till (*sensu* Evans *et al*., 2006) based on the following evidence: (a) clasts that show evidence of striae and contain a far-travelled component derived primarily from the Southern Uplands; (b) the fabric orientation is consistent with the flow-set within which the exposure is situated; (c) the high S_1 value which indicates streamlining of clasts (Benn, 1995; Evans, 2000); and (d) the diamicton is underlain by deformed glaciofluvial/glaciolacustrine sand, silt and clay which has been sheared up and incorporated into the diamicton (Hicock & Dreimanis, 1989; Benn & Evans, 1996).

Site 2: Carleton

Carleton gravel pit (NY 428 527) is located in the central zone of an arcuately organised tract of hummocky terrain, to the south east of Carlisle in the Petteril Valley (Fig. 1). The mound that the pit is situated within has a slight SW-NE trend, is 0.75 km long by 0.5 km wide and is 60 m O. D. (the exposure is situated at 50 m O. D.). The succession has been divided into two main lithofacies associations and one lithofacies described and interpreted in turn below, and displayed diagrammatically in Fig.10.

LFA CA1: Description:

LFA CA1 is composed of NNE dipping (22°) foresets composed of clast-supported granule-cobble gravel (max. diameter 30 cm) and coarse sand with some granule gravel (Fig. 9c & 10). Each foreset ranges from 2 – 15 cm and consists of either a coarsening upwards sequence from coarse sand with granule gravel to pebble-cobble gravel or alternations of open-work pebble gravel and closed-work granule-pebble gravel (Fig. 9c). Rare balls of diamicton are also observed. LFA CA1 is up to 2.2 m thick and is situated in the SSW corner of the gravel pit.

LFA CA1: Interpretation:

The dipping beds of LFA CA1 are interpreted as deltaic foresets, formed by cohesionless grain flows on a steep slope (Nemec, 1990; Nemec *et al*., 1999). The inverse grading is produced from dispersive pressures (Bagnold, 1956; Lowe, 1982), whilst openwork pebble-cobble gravels result from clogging in the upper layers by fines during fluctuating discharges (cf. Smith, 1985). The foreset orientations suggest that water flowed NNE, whilst the presence of diamicton balls indicate the delta must have been either in contact with, or near to, the ice front.

LFA CA2: Description:

LFA CA1 is overlain by a synclinal, 3.5 m thick by 15 m wide LFA (CA2) composed of interbedded clast-supported gravel lithofacies (up to 1. 0 m thick), infrequently observed thin (< 5 cm) clay/silt lithofacies and a 0.7 m bed of fine-grained sediments (Fig. 9d). The gravels, which dominate the LFA, range from pebble-cobble sized, and are either massive or display coarsening followed by fining upward sequences. The bed of fine-grained sediments is composed of a lower (*ca*. 0.3 m) band of silt capped by massive and normally graded fine sand with some sinusoidal ripple structures and a clay band. This fine-grained unit is heavily disturbed, with evidence of shearing, silt and clay stringers, flame structures, convoluted bands of clay, centimetre-scale reverse faults and a series of normal faults, with offsets of up to 0.3 m, which extend into the gravel lithofacies (above and below). The principle lithologies of LFA CA1 & CA2 are quartzitic sandstone (26%), greywacke (24%), Carboniferous Limestone (13%), Permian and Triassic sandstone (8%) and andesite, rhyolite & basalt (8%).

LFA CA2: Interpretation:

The laterally extensive, horizontally bounded and clast-supported gravel lithofacies are interpreted as longitudal bars (Boothroyd & Ashley, 1975; Rust, 1975; Miall, 1977, 1985; Smith, 1985; Marren, 2001) formed by tractional deposition in a glaciofluvial environment. The range of coarse-grained material, from granule-cobble size, suggests that there were significant fluctuations in discharge, with very high energies required to move the largest cobbles. The fine-grained sediments are typical of abandonment, or low-flow conditions, leading to the rapid deposition of the suspended sediment load (Miall, 1977; Smith, 1985).

LF CA3: Description:

At the very top of the succession is a thin (0.4 m thick), laterally discontinuous red-brown (5.0 YR 4/3) diamicton (Fig. 9e). The diamicton is massive, clast rich, with rounded to sub-rounded clast shapes. No striations were observed on clast surfaces. A macro-fabric taken from the diamicton reveals a NW-SE orientation (Fig. 10) with an S_1 eigenvalue of 0.46. The principle lithologies are andesite, rhyolite & basalt (31%), greywacke (30%), dolerite (12%), quartzitic sandstone (5%) and mudstone (4%).

LF CA3: Interpretation:

The diamicton of LF CA3 is characterised by a low S_1 eigenvalue, lack of striae, discontinuous nature, thin geometry and predominantly rounded-sub-rounded clast morphology. Thus LF CA3 is interpreted as a subaerial debris flow deposit (e.g. Lawson, 1981, 1982).

<u>Synopsis: environments of deposition:</u>

The exposure at Plumpe Farm forms part of a tripartite sequence consisting of an upper and lower till separated by fine-grained glaciofluvial or glaciolacustrine sediments. The deposition of an upper till which partially glaciotectonised the underlying sand, silt and clay lithofacies implies that ice underwent a re-advance phase. Macrofabric and clast provenance data support the interpretation of a Scottish re-advance emanating from Dumfries-shire and flowing eastwards into the Solway Basin (e.g. Dixon *et al.*, 1926; Trotter, 1929; Phillips *et al.*, 2007).

The gravel and sand lithofacies at Carleton demonstrate deltaic and glaciofluvial deposition (e.g. Marren, 2001). Extensive deformation, including normal faults, sheared boundaries, convolutions and flame structures, is characteristic of high water pressures and vertical collapse typically found in a proglacial environment (McDonald & Shilts, 1975; McCarroll & Rijsdijk, 2003), whilst the presence of diamicton balls support this inference. The clast provenance and palaeo-current indicators suggest that ice was situated to the S/SW of Carleton, with the gravels mainly derived from the local bedrock. The subaerial debris flow deposit (LF CA3), which caps the sequence, has a mixed provenance. However, geochemical analysis (unpublished work) indicates a similar geochemical signature to that of Plumpe Farm. Thus LF CA3 is likely to have been deposited during the Scottish re-advance.

3. Glacial geomorphology:

<u>A. flow sets related to the Scottish re-advance:</u>

Glacial geomorphological mapping within the Solway Lowlands has identified a series of lineations, meltwater channels, eskers, glaciofluvial deposits, hummocky terrain and ribbed moraine and has allowed the grouping of lineations into discrete flowsets (Fig. 1). These flowsets have been used to construct a relative chronology based on identifiable cross-cutting relationships between glacial features (cf. Livingstone *et al*., 2008). This has allowed a number of late-stage ice flows to be identified, which could relate to a Scottish re-advance.

Livingstone *et al*. (in prep) assigned the arcuate belt of lineations, hummocky terrain, and ribbed moraine of the Solway Lowlands along with the NE-SW orientated drumlins located in Bewcastle Fells and at Glasson (Fig. 1) to the “Blackhall Wood re-advance”. This re-advance flowed south-westwards into the Irish Sea Ice Basin as a tributary of the Irish Sea Ice Stream (Roberts, *et al*., 2007; Livingstone *et al*., in prep). The Blackhall Wood re-advance phase is subsequently cross-cut by Holme-St Cuthbert deltaic sequence (Fig. 1) which has been assigned to the Scottish re-advance (Huddart, 1970, 1981, 1991), and also a northerly orientated set of lineations moving out of the Vale of Eden and into the Solway Lowlands (Fig. 1) which must relate to a late-stage oscillation of Lake District ice. A thick band of diamicton towards the margin of this lobate re-advance (cf. Livingstone *et al*., in prep) could mark the position of an associated moraine.

The W-E orientated flowset of lineations marking the flow of ice down the Annan Valley and into the Solway Lowlands (Fig. 1) has been correlated with the similarly orientated Tyne Gap flowset which demarcates the easterly movement of ice across the Pennines (cf. Livingstone *et al*., 2008, sub). However, the stratigraphic sequence at Plumpe Farm (Fig. 8) suggests a similarly-orientated Scottish re-advance which deposited a thin till sheet across the region around Gretna. This may or may not have acted to remould or partially remould the Tyne Gap flowset leading into the Solway Lowlands.

B. Glacial meltwater channels:

Many of the meltwater channels incised into lineations formed during the ‘Blackhall Wood re-advance’ (Livingstone *et al*., in prep; Fig. 1 & 11) have caused controversy. This is especially true for the Wiza Beck meltwater channels which have been variously interpreted as ice-marginal channels associated with the Scottish re-advance (Dixon *et al*., 1926; Trotter, 1929) and subglacial channels (Arthurton & Wadge, 1981; Huddart, 1970). Huddart (1970) suggests that these channels formed prior to the Scottish re-advance, during ice recession. Similarly the Wampool meltwater channel has been both associated with the retreat of the ‘Main Glaciation’ (Huddart, 1970) and the subsequent ‘Scottish re-advance’ (Trotter, 1929; Trotter & Hollingworth, 1932).

Cutting across the western corner of the Lias plateau around Biglands (Fig. 1) are NE-SW orientated channels roughly 4 km long. The channels have slight up and down profiles (about 2 m), which dip gently down to the SW and are below 30 m O. D. (Fig. 11)

Wampool meltwater channel is a broad-bottomed channel (up to 0.5 km wide) over 11 km in length that trends westwards from the River Caldew out towards the Irish Sea coast (Fig. 1). The channel geometry is ‘U’ shaped, with a gentle long profile (Fig. 11) which decreases westwards (from 40 - 18 m O. D.). A number of the Wiza Beck channels form confluences along Wampool’s southern bank (Fig. 1).

Wiza Beck meltwater channels, on the northern-edge of the Lake District, form an arcuate network trending southwards from Wampool meltwater channel and then westwards towards the coast (Fig. 1). The channels are broadly aligned and are characterised by undulatory, and up-and-down long profiles which are deeply incised into the underlying drumlins (Fig. 11). The arcuate channels dip steeply towards Wampool meltwater channel, and gently down towards the coast. Channels which are orientated W-E dip (sometimes steeply) towards the coast. Channel lengths range between 1 and 16 km, with heights of between 10 – 110 m O. D. (Fig. 11).

C. Eskers:

Despite being below the spatial resolution of the NEXTMap DEM, ground-truthing has confirmed the existence (cf. Dixon *et al*., 1926; Huddart, 1970, 1981) of a number of subdued (generally < 5 m tall) eskers (correlated with the Scottish re-advance) superimposed over the top of lineations associated with the Blackhall Wood lineations (Fig. 1 & 5). At Thursby two discontinuous eskers were observed, running W-E with a series of right-angled kinks (Fig. 1 & 5). Another esker at Sowerby Wood, just north of Dalston (Fig. 1) on the Lias plateau is orientated SE-NW and branches northwards into two strands.

D. Flat-topped hills:

A number of flat-topped hills have been mapped to the east of Carlisle (Fig. 1). The two most distinctive flat-topped hills overlap at Wetheral (NY 468, 548) and Warwick-Bridge (NY, 455, 563), with the former being 1.8 x 0.8 km in dimension and 54 m O. D. and the latter being 2.0 x 1.0 km and 43 m O. D. (Fig. 1). Two further flat-topped hills have been recorded at Crosby-on-Eden (NY 458, 602) and Cam Beck (NY 499, 620), which are 30 m and 60 m O. D. respectively (Fig. 1). These have been identified as deltas due to their morphology (Fig. 1) and sedimentology (borehole data (BGS); Trotter & Hollingworth, 1932, Fig. 5), plus the presence of fine-grained lacustrine deposits throughout the region (Jackson, 1979; BGS borehole data).

These deltaic complexes must therefore constrain the evolution of a major glacial lake, formed against the reverse slope of the Tyne Gap, with meltwater drainage westwards blocked by an ice mass situated in the Solway Lowlands. The various levels of the flat-topped hills indicate a general trend of falling water levels westwards towards Carlisle. This glacial lake 'Glacial Lake Carlisle' has been previously identified, although the timing of its formation is controversial. It could either have formed during the 'Scottish re-advance' (Trotter, 1929), or during the main phase of deglaciation (Huddart, 1970, 1981, 1991).

Discussion:

The following discussion aims to answer the four research questions posed at the beginning of the paper. This will be based on a critical review of previous research regarding the Scottish re-advance and in light of new evidence presented in this paper. The primary objective is to ascertain which landforms and sediment assemblages can be confidently correlated to the Scottish re-advance. This needs re-evaluating in light of new evidence which demonstrates the existence of an earlier ('Blackhall Wood') re-advance in the central sector of the BIIS. A thorough examination of the evidence pertaining to the 'Scottish re-advance' is crucial if its configuration, impact and behaviour during the deglaciation of the BIIS are to be resolved.

Research Question 1:

What geomorphological and sedimentological evidence can be correlated to the re-advance of Scottish ice?

The Holme St Cuthbert ice-contact deltaic complex provides compelling evidence for a Scottish re-advance at a point subsequent to the Blackhall Wood oscillation (Livingstone *et al*., in prep), during a late stage of deglaciation. A glacial lake developed as ice situated along the Cumbrian coast impounded meltwater attempting to drain westwards into the Irish Sea Basin. The lakes extent is hard to surmise, although borehole evidence (BGS) of infrequently observed thin clay facies (capping the sequence) have been identified between Aspatria and Wigton. Despite this sporadic sedimentological evidence and absence of observable strand-lines, the height of the foreset structures (identified at Overby sand pit) at *ca*. 49 m O. D. indicate that the lake could potentially have extended over much of the Solway Lowlands.

Plumpe field site offers the best stratigraphic evidence for a re-advance of Scottish ice into the Solway Lowlands. The tripartite sequence clearly demonstrates re-advance of ice over glaciofluvial/glaciolacustrine sediments, from a Scottish sourced ice flow. This conclusion is in agreement with work carried out by Dixon *et al*. (1926), Trotter (1929), Phillips *et al*. (2007) and Stone *et al*., (in press). They observed a number of exposures within the Gretna region extending west to as far as Redkirk point (Fig. 1) characterised by tripartite divisions (Dixon *et al*., 1926; Trotter 1929). Clast macrofabrics and provenance data demonstrate that the till was deposited by an easterly flowing ice mass (see also Phillips *et al*., 2007). Similarly, at Carleton, geochemical analysis distinguishes a discontinuous, thin diamicton envisaged to have been deposited as a debris flow during the re-advance of ice out of Scotland.

The esker deposits at both Thursby and Sowerby Wood are also interpreted to have been deposited during the Scottish re-advance. Although these features have been identified within the field (see Fig. 5), this interpretation has been derived from data collated from previous research (cf. Dixon *et al*., 1926; Huddart, 1970, 1981). Clast lithological analysis reveals a Scottish provenance (Dixon *et al*., 1926; Huddart, 1970), whilst the esker at Thursby is also characterised by a series of small hills (Dixon *et al*., 1926; Huddart, 1970), not easily picked out on the NEXTMap DEM. These hills are characterised by steep slopes along their western margin giving way to lobate forms on their eastern side (Huddart, 1970). They are interpreted as fan/delta deposits which constrain the position of the ice margin during temporary still-stands, with water discharging eastwards into a subaqueous environment from subglacial tunnels (e.g. Banerjee & McDonald, 1975; Rust & Romanelli, 1975; Thomas, 1984; Warren & Ashley, 1994).

<u>Research Question 2:</u>

What was the maximum extent of the Scottish re-advance?

South of Carleton the tripartite sequence, comprising two tills separated by sand, gravel and laminated clay/silt lithofacies, has been apportioned to the ‘Blackhall Wood’ re-advance due to its stratigraphic position underneath lineations associated with flow out into the Irish Sea ice Basin (Livingstone *et al*., in prep). This tripartite sequence has previously been correlated to the Scottish re-advance, and as such, has resulted in an over-exaggeration of its maximal margins (Dixon *et al*., 1926; Trotter 1929; Fig. 2a). Indeed, recent work carried out by Livingstone *et al*. (in prep) suggests that the region south of Carlisle is the sole preserve of the ‘Blackhall Wood’ till sequence.

The Wiza Beck meltwater channels are interpreted to have formed during the retreat stage of the 'Blackhall Wood' flow phase (e.g. Huddart, 1970), when ice was both downwasting and shrinking eastwards back into the Vale of Eden. Therefore, these meltwater channels do not constrain the southerly limits of the Scottish re-advance, as has been previously postulated (*cf*. Dixon *et al*., 1926; Trotter, 1929). This interpretation is based on their complex morphology, characterised by sharp bends and up-down long-profiles. Their morphology indicates either a subglacial origin (Sisson, 1960, 1961; Sugden *et al*., 1991) or genesis as a result of two separate stages of development, with the W-E orientated channel sections parallel configuration characteristic of an ice sub-marginal genesis (Sissons, 1961; Dyke, 1993). This ice sub-marginal interpretation is supported by a number of small sand and gravel deposits (up to 70 m O. D.) at the western end of the perched meltwater channels around Wigton (Eastwood *et al*., 1968), which probably relate to delta/fan deposits formed as water debouched into a subaerial or subaqeous environment. Both interpretations require west-east ice movement, which is contrary to the movement of the Scottish re-advance.

To the east of Carlisle, the biggest glacial feature is Brampton kame belt (Fig. 1) situated in the lee of the Pennines and comprised of glaciofluvial and glaciolacustrine deposits up to 25 m thick which overlie bedrock or in some instances diamicton (Jackson, 1979; Huddart, 1970, 1981; Livingstone *et al*., sub). This feature is likely to have formed during the retreat of the ice out of the Tyne Gap, prior to, or during the Blackhall Wood re-advance (Livingstone *et al*., sub). The absence of upper till or overprinted landforms limits the Scottish re-advance to the west of the kame belt.

Stratigraphic evidence, east of Carlisle, indicates the existence of a discontinuous and thin diamicton sheet (< 1.0 m) overlying the Crosby Moor delta (Huddart, 1970), observed at Botcherby clay pit (Trotter, 1929; Carruthers, 1932) and Eden Bridge (Huddart, 1994), and exposed in river sections of the Irthing, King Water and Cam (Trotter & Hollingworth, 1932). This diamicton sheet caps a series of glaciolacustrine deposits, which become exposed at the surface at Lanercost, Boothby Bank and Great Easby (Fig. 1). The discontinuous nature and thin geometry, similar to LF CA3, is indicative of debris flows (Lawson, 1981, 1982) deposited subaerially (*cf*. Huddart, 1970), although a subglacial origin cannot be ruled out. Their stratigraphic position is consistent with a Scottish re-advance genesis (Trotter & Hollingworth, 1932b).

The glaciolacustrine deposits (BGS boreholes; Trotter & Hollingworth, 1932; Huddart, 1970) and deltas underlying the diamicton lithofacies signifies the presence of a major glacial lake ('Glacial Lake Carlisle') impounded against ice to the west of Carlisle. It is envisaged that the lake formed against the reverse slope of the Tyne Gap, with the westerly retreating ice margin resulting in falling lake levels. Thus, it is likely that 'Glacial Lake Carlisle' formed before the Scottish re-advance, during gradual retreat of either the 'Main Glaciation' or 'Blackhall Wood' oscillation. This is supported by the sections at Plumpe Farm and Carleton, both of which suggest significant glaciofluvial/glaciolacustrine activity during a period preceding the Scottish re-advance.

The geometry and long profile of Wampool meltwater channel is indicative of a proglacial meltwater channel (cf. Benn & Evans, 1998), whereby large volumes of water were discharged westwards. It could therefore be associated with either (or both) the rapid over-spilling of Lake Carlisle (Huddart, 1970) during eastwards retreat of the Blackhall Wood oscillation; or functioned as a major proglacial channel throughout the late stage lobate re-advance of Vale of Eden ice into the fringe of the Solway Lowlands (see research question 3).

Overall, it is thought that the re-advance of Scottish ice into the Solway Lowlands was less extensive than some previous models (e.g. Trotter, 1929) and more in agreement with Huddart (1970). With the discovery of the 'Blackhall Wood' re-advance much of the evidence pertaining to the southerly expansion of the Scottish re-advance can be rejected, while several of the morphological features which have previously been correlated to the Scottish re-advance are probably relicts of the deglaciation of the 'Blackhall Wood' re-advance. The lack of exposures east of Carlisle has made it hard to verify the existence (or not) of a subglacial till extending east as far as Lanercost (e.g. Trotter, 1929). However, the discontinuous diamicton, exposed at Carleton, and interpreted as a debris flow suggests that the presence of an extensive subglacial till is unlikely.

Research Question 3:

Was there was a concurrent re-advance of ice from the Lake District?

Evidence of a flowset of south-north orientated subglacial lineations trending out of the Vale of Eden into the fringe of the Solway Lowlands demarcates a late stage ice flow which occurred after the 'Blackhall Wood' re-advance. This lobate re-advance may have occurred concurrently with the Scottish re-advance, although it cannot be conclusively proved, and may alternatively represent a re-adjustment of the Lake District ice mass to internal forcing.

Research Question 4:

What does the evidence available tell us about ice behaviour during this re-advance?

In general, the stratigraphic evidence for a re-advance of Scottish ice into the Solway Lowlands is 'patchy'. Diamicton correlated to the Scottish re-advance is generally less than 2.0 m thick and discontinuously exposed throughout the region. At Glasson (NY 259 609), for example, an exposure through a NE-SW orientated drumlin associated with the 'Blackhall Wood' re-advance is shown to exhibit no evidence of the subsequent Scottish re-advance despite being well within its bounds. Furthermore, the W-E orientated flowset, north of the Solway Firth, which is traceable through the Tyne Gap (Livingstone *et al*., 2008), has seemingly not been affected by the re-advance of Scottish ice. Indeed the till cap (< 2.0 m) probably formed a veneer over the already drumlinized landscape. Thus it can be concluded that neither depositional (of a till sheet) or erosional (remoulding features) processes played a major role during the Scottish re-advance. One explanation is that the high water contents effectively buffered the transition of stresses into the underlying sediment, with the ice able to move rapidly across this lubricated layer (e.g. Piotrowski, *et al*., 2001). This effect is most apparent at Plumpe Farm (cf. Phillips *et al*., 2007), and offers an explanation for the lack of till, glacial lineations and moraine throughout the region. An alternative explanation is that it was ice-sediment poor.

The esker are characterised by simple morphologies, sharp bends and a predominance of glaciofluvial sediments (e.g. Dixon *et al*., 1926; Huddart, 1970) all of which are analogous of concertina eskers, which have been related to surge events along active ice margins (e.g. Knudsen, 1995; Evans & Twigg, 2002). This would tend to indicate a short-lived and rapid advance into the Solway Basin.

Model of ice re-advance and wider implications:

Based on evidence presented in this paper, it has been possible to produce an up-to date model for the re-advance of Scottish ice into the Solway Lowlands (Fig. 12):

Holme St Cuthbert marks the most recognisable still-stand of Scottish re-advance ice within the Solway Lowlands (Fig. 12). The dead ice morphology and size of the deltaic sequence indicate a stillstand of significant duration or rapid deposition. A number of other features identified along the coast seemingly correlate with this stage, including the glacial outwash deltas and eskers at Gretna (cf. Stone *et al*., in prep) at 30 m O. D., the St Bees moraine complex (*cf*. Huddart, 1994; Merritt & Auton, 2000) and the proglacial sandur deposits at Harrington (*cf*. Huddart, 1994). Thus it is envisaged that the main stillstand of the Scottish re-advance impinged onto the Cumbrian coast, blocking drainage and subsequently leading to lake formation within the Wigton region (Fig. 12). Ice almost certainly advanced up to the Lias Plateau thus preventing water draining northwards into the Gretna region (Fig. 12). The foresets at Holme St Cuthbert indicates a lake level of approximately 49 m O. D. (Fig. 12).

The eskers at Thursby and Sowerby Hill can be envisaged to demarcate the most advanced position of ice within the Solway Lowlands. The small, subdued nature, and 'concertina' style of the eskers demonstrate a rapid, transient 'surge-like' ice flow (e.g. Knudsen, 1995), leading to the development of small lakelets, into which the subglacial tunnels discharged (e.g. Banerjee & McDonald, 1975). This would explain the lack of corresponding evidence throughout the region of Wigton and north and east of Carlisle, with disparate diamicton patches occasionally observed along riverbanks as far east as the Irthing and Cam (e.g. Dixon *et al*., 1926; Trotter *et al*., 1929; Trotter & Hollingworth, 1932; Huddart, 1970, 1994). The 'patchy' till suggests either that the ice was sediment poor, or that the erosion capacity of the ice was buffered by high water pressures (e.g. Piotrowski *et al*., 2001). Indeed, the evidence supports a model of thin lobate advance into the lowlands, more akin to Huddart's 60 m ice limit than Trotter's proposed 134 m high (Fig. 12).

Although the Scottish re-advance can be traced down the west Cumbrian coast (*cf*. Huddart, 1994; Merritt & Auton, 2000), evidence presented in this paper implies that it was a relatively minor and short-lived fluctuation with little erosive power. Indeed most of the evidence for the Scottish re-advance is limited to glaciolacustrine/glaciofluvial deposits indicative of wide-spread lake development, dammed up between ice in the Celtic Basin and upland regions inland. Conversely, the Killard Point stadial, which has been correlated with the Scottish re-advance (McCabe *et al*., 1998; McCabe & Clark, 1998), was thought to be more extensive. This is typified by the coeval advance of ice onto the Isle of Man (Bride moraine) and the drumlinisation and formation of rogen moraine throughout NE Ireland (McCabe *et al*., 1998). Although the Scottish re-advance lacks accurate dating controls making correlations hard, the identification of an earlier 'Blackhall Wood' re-advance (Livingstone *et al*., in prep) supports work in NE Ireland, that the Celtic Basin was subject to two re-advances (Clogher Head and Killard Point re-advances), with the second one correlated with Heinrich event 1 (McCabe *et al*., 2007). This is also in agreement with work carried out in western Cumbria (Gosforth and Scottish re-advance) (Merritt & Auton, 2000). Given the regional response of the Scottish re-advance (McCabe *et al*., 1998), which indicates an external forcing, the lobate re-advance of ice in the Vale of Eden (and sourced in the Lake District) is likely to have been coeval (Fig. 12).

Conclusions:

The complex, palimpsest landscape of the Solway Lowlands had hindered attempts to reconstruct and interpret individual ice flow phases. For the Scottish re-advance, a paucity of well-constrained landforms, sediments and dates has inevitably led to debate over its existence, limits and dynamic behaviour. This paper has combined both glacial geomorphic and sedimentological information to critically review the evidence for and against a late stage Scottish re-advance into the Solway Lowlands, and to produce an up-to-date model summarising these interpretations (Fig. 12). Several key points can be been drawn from this study, and these are documented below:

1. Re-advance of Scottish ice into the Solway Lowlands did occur at a stage subsequent to the Main Glaciation and Blackhall Wood re-advance. This is supported by disparate stratigraphic evidence of an upper till, a series of eskers along the edge of the Lias Plateau and the Holme St Cuthbert deltaic sequence.

2. A major still-stand recorded by the Holme St Cuthbert delta sequence and traced along the Cumbrian coast marks the most obvious limit of the Scottish re-advance. At this time a large ice-contact lake formed in the vicinity of Wigton. Subdued eskers and some evidence of an upper till points towards a transient ice advance further inland (Fig. 12) that is not thought to relate to the Holme St Cuthbert still-stand. A lobate re-advance of ice down the Vale of Eden may have been concurrent with the Scottish re-advance.

3. Evidence at Plumpe Farm and Carleton, plus the absence of glacial till throughout the region (e.g. Glasson and east of Carlisle) and lack of glacial lineations/moraines related to the Scottish re-advance can be reconciled with a passive, transient and thin ice advance.

4. Other glacial features that have in the past been assigned to the Scottish re-advance, including the Wampool and Wiza Beck meltwater channels, an upper till south of Carlisle and Lake Carlisle are re-interpreted here as belonging to earlier to flow phases.

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References:

Allen, J.R.L. 1968. Current Ripples. *North-Holland, Amsterdam*, 433 pp

Allen, J.R.L. 1973. A classification of climbing-ripple cross-lamination. *Journal of the Geological Society*, 129(5); 537-541.

Allen, J.R.L. 1984. Sedimentary Structures - their Character and Physical Basis. *Elsevier,Amsterdam*.

Bagnold, B. 1956. The flow of cohesionless grains in fluids. Philosophical *Transactions of the Royal Society of London*, A249; 235-297.

Banerjee, I. & McDonald, B.C. 1975. Nature of esker sedimentation. *In*, Jopling, A. V. and McDonald, B. C. *Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists Special Publication No. 23.*

Benn, D.I. 1995. Fabric signature of subglacial till deformation, Breiðamerkurjökull, Iceland. *Sedimentology*, 42(5); 279-292.

Benn, D.I. & Evans, D.J.A. 1996. The interpretation and classification of subglacially deformed materials. *Quaternary Science Reviews*, 15; 23-52.

Benn, D.I. & Evans, D.J.A. 1998. Glaciers and Glaciation. *Arnold: London*

Boothroyd, J.C. & Ashley, G.M. 1975. Processes, bar morphology, and sedimentary structures on braided outwash fans, North-eastern Gulf of Alaska. *In*, Jopling, A. V. and McDonald, B. C. *Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists Special Publication No. 23;* 193-222.

Boulton G.S. & Clark C.D. 1990a. The Laurentide ice sheet through the last glacial cycle: the topology of drift lineations as a key to the dynamic behaviour of former ice sheets. *Transactions of the Royal Society of Edinburgh: Earth Sciences* 81, 327-347.

Boulton G.S. & Clark C.D. 1990b. A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations. *Nature* 346, 813-817.

Carruthers, R.G. 1939. On Northern glacial drift: some peculiarities and their significance. *Quarterly Journal of the Geological Society*, 95; 299-333.

Cheel, R.J. & Rust, B.R. 1982. Coarse-grained facies of glacio-marine deposits near Ottawa, Canada. In Davidson, A.R., Nickling, W. & Fahley, B.D. (eds). *Research in Glacial, Glaciofluvial, and Glaciolacustrine systems: Guelph, Ontario, Canada, University of Guelph, Guelph symposium on Geomorphology, Proceedings*, 279-295.

Clemmensen, L.B. & Houmark-Nielsen, M. 1981. Sedimentary features of a Weichselian glaciolacustrine delta. Boreas, 10; 229-245.

Collinson, J.D. & Thompson, D.B. 1989. Sedimentary Structures. *Unwin Hyman, London*.

Cook, J.H. 1946. Kame-complexes and perforation deposits. *American Journal of Science*, 24; 573-583.

Charlesworth, J.K. 1926. The glacial geology of the southern uplands of Scotland, west of Annandale and upper Clydesdale. *Transactions of the Royal Society of Edinburgh*, 55; 1-23.

Clark C.D. 1997. Reconstructing the evolutionary dynamics of former ice sheets using multi-temporal evidence, Remote Sensing and GIS. *Quaternary Science Reviews* 16, 1067-1092,

Clark, C. D. 1999. Glaciodynamic context of subglacial bedform generation and preservation. *Annals of Glaciology* 28, 23-32.

Cohen, J.M. 1979. Deltaic sedimentation in glacial lake Blessington, County Wicklow, Ireland. *In* Schluchter, C. (ed). Moraines and Varves. *Rotterdam, A. A. Balkema*, p. 357-367.

Dixon, E.E.L., Maden, J.M., Trotter, F.M., Hollingworth, S.E. & Tonks, L.H. 1926. Geology of the Carlisle, Longtown and the Silloth District. *Memoir of the British Geological Survey, HMSO, London.*

Dyke, A.S. 1993. Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography*, 17; 223-247.

Eastwood, T., Hollingworth, S.E., Rose, W.C.C. & Trotter, F.M. 1968. Geology of the country around Cockermouth and Caldbeck. *Memoir of the Geological Survey of Great Britain. HMSO, 298pp.*

Evans, D.J.A. 2000. A gravel outwash/deformation till continuum, Skalafellsjokull, Iceland. *Geografiska Annaler*, 82A(4); 499-512.

Evans, D.J.A. & Twigg, D.R. 2002. The active temperate glacial landsystem: a model based on Breiðamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews*, 21; 2143-2177.

Evans, D.J.A. & Benn, D.I. 2004. A practical guide to the study of glacial sediments. *Arnold, London.*

Evans, D.J.A., Phillips, E.R., Hiemstra, J.F. & Auton, C.A. 2006. Subglacial till: formation, sedimentary characteristics and classification. *Earth Science Reviews*, 78; 115-176.

Evans, D.J.A., Livingstone, S.J., Vieli, A. & Ó Cofaigh C. 2009. The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling. *Quaternary Science Reviews*, 28; 735-757.

Evans, W.B. & Arthurton, R.S. 1973. North-west England. *In* Mitchell, G.F., Penny, L.F., Shotton, F.W. & West, R.G. (eds). *A correlation of Quaternary deposits in the British Isles*. Geological Society, London, Special Reports, 4, 28-36.

Eyles, N. & McCabe, A.M. 1989. The Late Devensian (<22,000 BP) Irish Sea Basin: the sedimentary record of a collapsed ice sheet margin. *Quaternary Science Reviews*, 8; 307-351.

Goodchild J.G. 1875. Glacial phenomena of the Eden Valley and the western part of the Yorkshire-Dale District. *Quarterly Journal of the Geological Society of London* 31, 55-99.

Goodchild, J.G. 1887. Ice work in Edenside and some of the adjoining part of North West England. *Transactions of the Cumberland and Westmorland Advancement of Literature and Science*, 12; 111-167

Gorrell, G. & Shaw, J. 1991. Deposition in an esker, bead and fan complex, Lanark, Ontario, Canada. *Sedimentary Geology*, 72; 285-314.

Greenwood, S.L. & Clark, C.D. 2008. Subglacial bedforms of the Irish Ice Sheet. *Journal of Maps*, v2008; 332-357.

Gustavson, T.C., Ashley, G.M. & Boothroyd, J.C. 1975. Depositional sequences in glaciolacustrine deltas. *In*, Jopling, A.V. and McDonald, B.C. *Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists Special Publication No. 23;* 264-280

Hart, J K. & Boulton, G.S. 1991. The interrelation of glacitectonic and glaciodepositional processes within the glacial environment. *Quaternary Science Review*, 10; 335-350.

Hicock, S.R. & Dreimanis, A. 1989. Sunnybrook drift indicates a grounded early Wisconsin glacier in the Lake Ontario basin. *Geology*, 17; 169-172.

Hollingworth S.E. 1931. Glaciation of West Edenside and adjoining areas and the drumlins of Edenside and the Solway Plain. *Quarterly Journal of the Geological Society of London* 87; 281-357.

Holmes, C.D. 1947. Kames. *American journal of Science*, 245; 240-249.

Huddart D. 1970. *Aspects of glacial sedimentation in the Cumberland Lowland.* Unpublished PhD Thesis, University of Reading.

Huddart, D. 1971a. A relative glacial chronology from the tills of the Cumberland lowland. *Proceedings of the Cumberland Geological Society*, 3; 21-32.

Huddart, D. 1971b. Textural distinction between the Main Glaciation and Scottish re-advance tills in the Cumberland Lowland. *Geological Magazine*, 108; 317-324.

Huddart, D. 1981. Fluvioglacial systems in Edenside. *In* Field Guide to Eastern Cumbria. Boardman, J. (ed). *The Quaternary of Cumbria: Field Guide*, pp 151-167. Quaternary Research Association, London, 128 pp.

Huddart, D. 1991. The glacial history and deposits of the North and West Cumbrian Lowlands. *In* Ehlers, J., Gibbard, P. L. & Rose, J. (eds). *Glacial deposits in Britain and Ireland*, p 151-167. Rotterdam: Balkema Press.

Huddart, D. 1994. The late Quaternary glacigenic sequence: landforms and environments in coastal Cumbria. *In* Boardman, J. & Walden, J. (eds). *The Quaternary of Cumbria: Field Guide*. Quaternary Research Association, Oxford, 59-77.

Huddart, D. & Tooley, M.J. (Eds) (1972). Field Guide to the Cumberland Lowland. *Cambridge: Quaternary Research Association.*

Huddart, D. & Glasser, N.F. 2002. Quaternary of Northern England. Geological Conservation Review Series, No. 25. *Joint Nature Conservation Committee, Peterborough, 745pp.*

Jackson, I. 1979. The sand and gravel resources of the country around Brampton, Cumbria: description of 1:25,000 resources sheet NY 55 and part of NY 56. *Mineral Assessment Report for the Institute of Geological Sciences,* No. 45.

Jopling, A.V. & Walker, R.G. 1968. Morphology and origins of ripple-drift cross lamination, with examples from the Pleistocene of Massachusetts. *Journal of Sedimentary Research,* 38(4); 971-984.

Kneller, B. 1995. Beyond the turbidite paradigm: physical models for deposition of turbidites and their implications for reservoir prediction. *In* Hartley, A. J. And Prosser, D. J. (eds) Characterisation of deep-marine clastic systems, *Geological Society Special Publication*, vol. 94; 31-49.

Knudsen, O. 1995. Concertina eskers, Brúarjökull, Iceland: An indicator of surge-type glacier behaviour. *Quaternary Science Reviews*, 14; 487-493.

Lawson, D.E. 1981. Distinguishing characteristics of diamictons at the margin of the Matanuska Glacier, Alaska. ANNALS OF Glaciology, 2; 78-83.

Lawson, D.E. 1982. Mobilisation, movement and deposition of active subaerial sediment flows, Matanuska Glacier, Alaska. *Journal of Geology*. 90; 279-200.

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A. 2008. Glacial geomorphology of the central sector of the last British-Irish Ice Sheet. *Journal of Maps*, 2008; 358-377.

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A., sub. A major ice drainage pathway of the last British-Irish Ice Sheet: the Tyne Gap, northern England. *Journal of Quaternary Science*.

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A., Hopkins, J. sub.The Brampton Kame Belt and Pennine Escarpment meltwater channel system (Cumbria, UK): Morphology, Sedimentology and Formation. *Proceedings of the Geologists Association*.

Livingstone, S.J., Ó Cofaigh, C. & Evans, D.J.A. in prep. Glaciolacustrine sedimentation in the Solway Lowlands (Cumbria, UK): evidence for a major glacial oscillation during Late Devensian deglaciation.

Lowe, D.R. 1982. Sediment gravity flows; II, Depositional models with special reference to the deposits of high-density turbidity currents. *Journal of Sedimentary Research*, 52(1); 279-297.

Marren, P.M. 2001. Sedimentology of proglacial rivers in eastern Scotland during the Late Devensian. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, 92; 149-171.

McCabe, A.M. & Clark, P.U. 1998. Ice-sheet variability around the North Atlantic Ocean during the last deglaciation. *Nature*, 392; 373-377.

McCabe, A.M., Knight, J. & McCarron, S. 1998. Evidence for Heinrich event 1 in the British Isles. *Journal of Quaternary Science*, 13; 549-568.

McCarroll, D. & Rijsdijk, K.F. 2003. Deformationstyles as a key for interpreting glacial depositional environments. Journal of Quaternary Science, 18(6); 473-489.

McDonald, B.C. & Shilts, W.W. 1975. Interpretation of faults in glaciofluvial sediments. *In*, Jopling, A. V. and McDonald, B. C. *Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists Special Publication No. 23;* 123-131.

Merritt, J.W. & Auton, C.A. 2000. An outline of the lithostratigraphy and depositional history of Quaternary deposits in the Sellafield District, West Cumbria. *Proceedings of the Yorkshire Geological Society*, 53(2); 129-154.

Miall, A.D. 1977. A review of the braided-river depositional environment. *Earth Science Reviews*, 13; 1-62.

Miall, A.D. 1985. Architectural-element analysis: a new method of facies analysis applied to fluvial deposits. *Earth Science Reviews*, 22; 261-308.

Nelson, A.E., Willis, I.C. & Ó Cofaigh, C. 2005. Till genesis and glacier motion inferred from sedimentological evidence associated with the surge-type glacier, Brúarjökull, Iceland. *Annals of Glaciology*, 42; 14-22.

Nemec. W. 1990. Aspects of sediment movement on steep delta slopes. *In* Colella, A. & Prior, D. B. (eds.). Coarse-drained Deltas. *International Association if Sedimentologists Special Publication 10,* 29-73.

Nemec, W., Lønne, I. & Blikra, L. 1999. The Kregnes moraine in Gauldalen, west-central Norway: anatomy of a Younger Dryas proglacial delta in a palaeofjord basin. *Boreas*, 28; 454-476.

Paul, M.A. 1983. Supraglacial landsystem. *In* Eyles N. (eds). *Glacial Geology, 91-90, Pergamon Press, Oxford.*

Pennington, W. 1978. Quaternary Geology. *In*, Mosley, F. (ed). *The Geology of the Lake District. Yorkshire Geological Society Occasional Publication*, 3, 207-225.

Phillips, E., Merritt, J., Auton, C. & Golledge, N. 2007. Microstructures in subglacial and proglacial sediments: understanding faults, folds and fabrics, and the influence of water on the style of deformation. Quaternary Science Reviews, 26, 1499-1524.

Piotrowski, J.A., Mickelson, D.M., Tulaczyk, S., Krzyszkowski, D. & Junge, F. 2001. Were deforming subglacial beds beneath past ice sheets really widespread? *Quaternary International,* 86; 139-150.

Plink-Björklund, P. & Ronnert, L. 1999. Depositional processes and internal architecture of Late Weichselian ice-margin submarine fan and delta settings, Swedish west coast. *Sedimentary Geology*, 46; 215-234.

Postma, G., Roep, T.B. & Ruegg, G.J.H. 1983. Sandy.. gravelly mass-flow deposits in an ice-marginal lake (Saalian, Leuvenumsche Beek Valley, Veluwe, The Netherlands), with emphasis on plug-flow deposits. Sediment. Geol, 134: 59-82_

Powell, R.D. 1990. Glacimarine processes at grounding-line fans and their growth to ice-contact deltas. *Geological Society, London, Special Publications*, 53; 53-73.

Price, R.J. 1969. Moraines, sandar, kames and eskers near Breidamerkurjokull, Iceland. *Transactions of the Institute of British Geographers*, 46; 17-43.

Price, R.J.(1973. Glacial and fluvio-glacial landforms.

Punkari M. 1993. Modelling of the dynamics of the Scandinavian ice sheet using remote sensing and GIS methods. In, Aber J.S. (ed.), *Glaciotectonics and Mapping Glacial Deposits*. Balkema, Rotterdam, 232-250.

Reineck, H.E. & Singh, I.B. 1975. Depositional Sedimentary Environments. *Springer Verlag, Berlin.*

Rijsdijk, K.F., Owen, G., Warren, W.P., McCarroll, D. & van der Meer, J.J.M. 1999. Clastic dykes in over-consolidated tills: evidence for subglacial hydrofracturing at Killiney Bay, eastern Ireland. *Sedimentary Geology*, 129; 111-126.

Roberts, D.H., Dackombe, R.V. & Thomas, G.S.P. 2007. Palaeo-ice streaming in the central sector of the British-Irish Ice Sheet during the Last Glacial Maximum: evidence from the northern Irish Sea Basin. *Boreas* 36; 115-129.

Russell, H.A.J. & Arnott, R.W.C. 2003. Hydraulic-jump and hyperconcentrated flow deposits of a glacigenic subaqueous fan: Oak Ridge moraine, southern Ontario, Canada. Journal of Sedimentary Research, 73(6); 887-905.

Rust, B.R. 1975. Fabric and structure in glaciofluvial gravels. *In* Jopling, A.V. & McDonald, B.C. (eds). *Glaciofluvial and Glaciolacustrine Sedimentation: Society of Economic Paleontologists and Mineralogists, Special Publication*, 23; 239-248.

Rust, B.R. & Romanelli, R. 1975. Late Quaternary subaqueous deposits near Ottawa, Canada. *In* Jopling, A.V. & McDonald, B.C. (eds). *Glaciofluvial and Glaciolacustrine Sedimentation: Society of Economic Paleontologists and Mineralogists, Special Publication*, 23; 177-192.

Salt K.E. & Evans D.J.A. 2004. Scottish Landform Example 32: Superimposed subglacially streamlined landforms of southwest Scotland. *Scottish Geographical Journal* 120, 133–147.

Shanmugan, G. 2000. 50 years of the turbidite paradigm (1950s—1990s): deep-water processes and facies models—a critical perspective. Marine and Petroleum Geology, 17(2); 285-342.

Sissons, J.B. 1960. Some aspects of glacial drainage channels in Britain. Part I. *Scottish Geographical Magazine*, 76, 131–146.

Sissons, J.B. 1961. Some aspects of glacial drainage channels in Britain. Part II. *Scottish Geographical Magazine*, 77; 15-36.

Sissons, J.B. 1974. The Quaternary in Scotland – a review. *Scottish Journal of Geology*, 10; 311-337.

Smith, N.D. 1985. Proglacial fluvial environment. *In* Ashley, G.M., Shaw, J. and Smith, N.D. (eds). *Glacial Sedimentary Environments.*

Smith, N.D. & Ashley, G. 1985. Proglacial lacustrine environment. *In* Ashley, G. M., Shaw, J. and Smith, N. D. (eds). *Glacial Sedimentary Environments.*

Stone, P., Lawrence, D., Merritt, J.W. and Millward, D. in press. *British Regional Geology*: Northern England (Third Edition). British Geological Survey, Nottingham.

Sugden, D. E., Denton, G. H. & Marchant, D. R. (1991). Subglacial meltwater channel systems and ice sheet overriding, Asgard Range, Antarctica. *Geografiska Annaler*, 73A;109-121.

Thomas, G.S.P. 1984. Sedimentation of a sub-aqueous esker-delta at Strabathie, Aberdeenshire. *Scottish Journal of Geology* , 20; 9-20.

Thomas, G.S.P. 1985. The Quaternary of the northern Irish Sea basin. *In* Johnson, R. H. (ed). *The geomorphology of North-west England*. Manchester University Press, Manchester, 143-158.

Thomas, G.S.P., Chiverrell, R.C. & Huddart, D. 2004. Ice-marginal depositional responses to readvance episodes in the Late Devensian deglaciation of the Isle of Man. *Quaternary Science Reviews*, 23; 85-106.

Thomas, G.S.P. & Chiverrell, R.C. 2006. A model of subaqueous sedimentation at the margin of the Late Midlandian Irish Ice Sheet, Connemara, Ireland, and its implications for regionally high isostatic sea-levels. *Quaternary Science Reviews*, 25; 2868-2893.

Trotter F.M. 1922. Report from the Cumberland District. *In*: Summary of Progress of the Geological Survey of Great Britain for 1921, 46-48.

Trotter F.M. 1923. Report from the Cumberland District. *In*: Summary of Progress of the Geological Survey of Great Britain for 1922, 61-63.

Trotter F.M. 1929. The Glaciation of East Edenside, the Alston Block and the Carlisle Plain. *Quarterly Journal of the Geological Society of London* 85; 549-612.

Trotter, F.M. & Hollingworth, S.E. 1932a. The glacial sequence in the North of England. *Geological Magazine*, 69; 374-380.

Trotter, F.M. & Hollingworth, S.E. 1932b. The geology of the Brampton district. *Memoir of the Geological Society, HMSO, London.*

Warren, W.P. & Ashley, G.M. 1994. Origins of the Ice-contact Stratified Ridges (Eskers) of Ireland. *Journal of Sedimentary Research*, 64A; 433-449.

Williams, G.D., Brabham, P.J., Eaton, G.P. & Harris, C. 2001. Late Devensian glaciotectonic deformation at St Bees, Cumbria: a critical wedge model. *Journal of the Geological Society, London*, 158; 125-135.

Winsemann, J. Asprion, U., Meyer, T. & Schramm, C. 2007. Facies characteristics of Middle Pleistocene (Saalian) ice-marginal subaqueous fan and delta deposits, glacial Lake Leine, NW Germany. *Sedimentary Geology*, 193; 105-129.

Van der Meer., J.J.M., Kjaer, K.H. & Krüger, J. 1999. Subglacial water-escape structures and till structures, Sléttjökull, Iceland. *Journal of Quaternary Science*, 14(3); 191-205.

Van der Meer., J.J.M., Kjaer, K.H., Krüger, J., Rabassa, J. & Kilfeather, A.A. 2009. Under pressure: clastic dykes in glacial settings. *Quaternary Science Reviews*, 28;708-720

http://www.neodc.rl.ac.uk/

Figures:

Fig. 1: (a) topographic map showing fieldsites (red boxes) and key locations; and (b) NEXTMap DEM showing the flowsets and glacial features. NEXTMap Britain data from Intermap Technologies Inc were provided courtesy of NERC via the NERC Earth Observation Data Centre (NEODC).

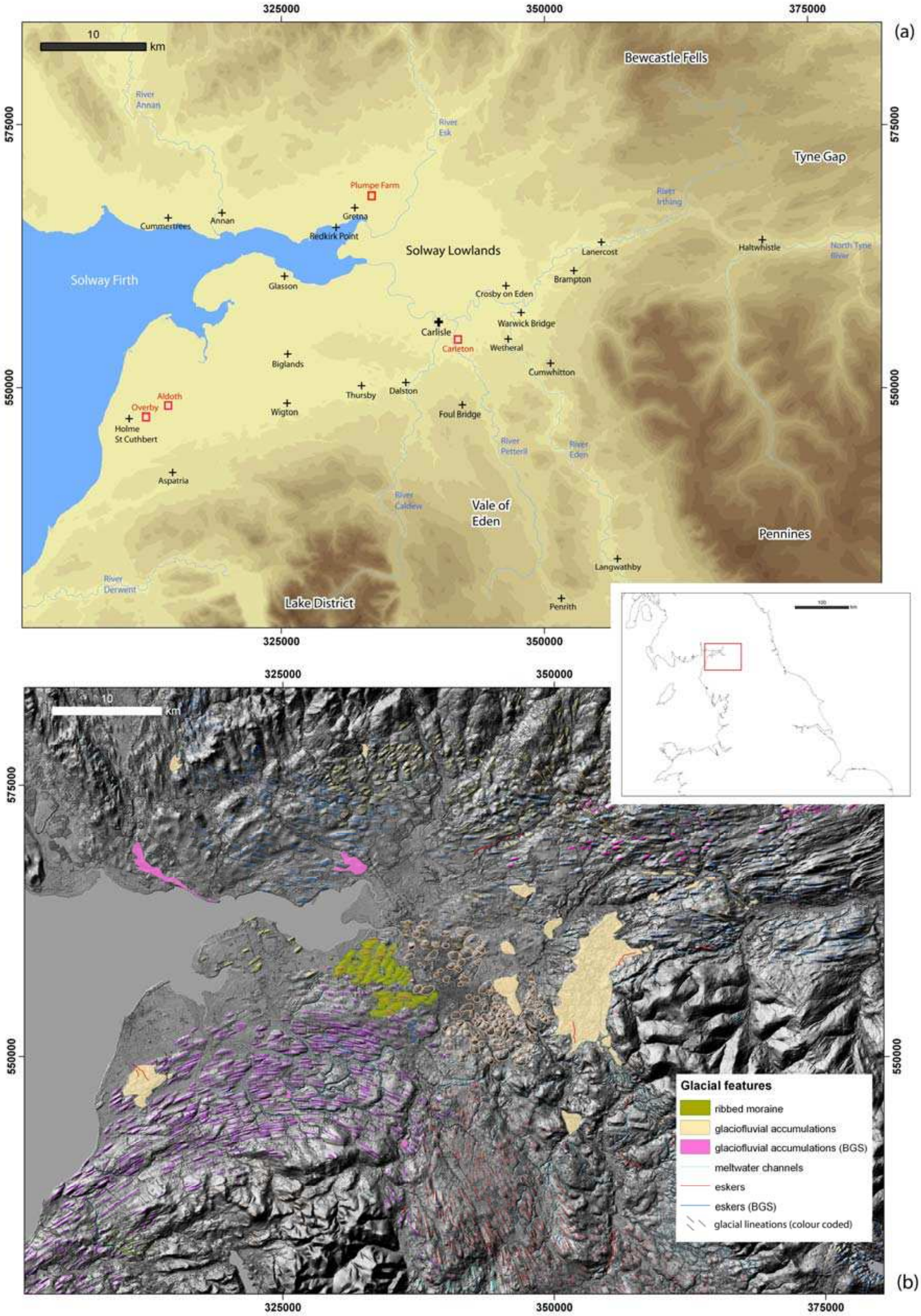


Fig. 2: Map of the various models of maximum extent for the Scottish re-advance: (a) Trotter, (1929); and (b) Huddart, 1994).

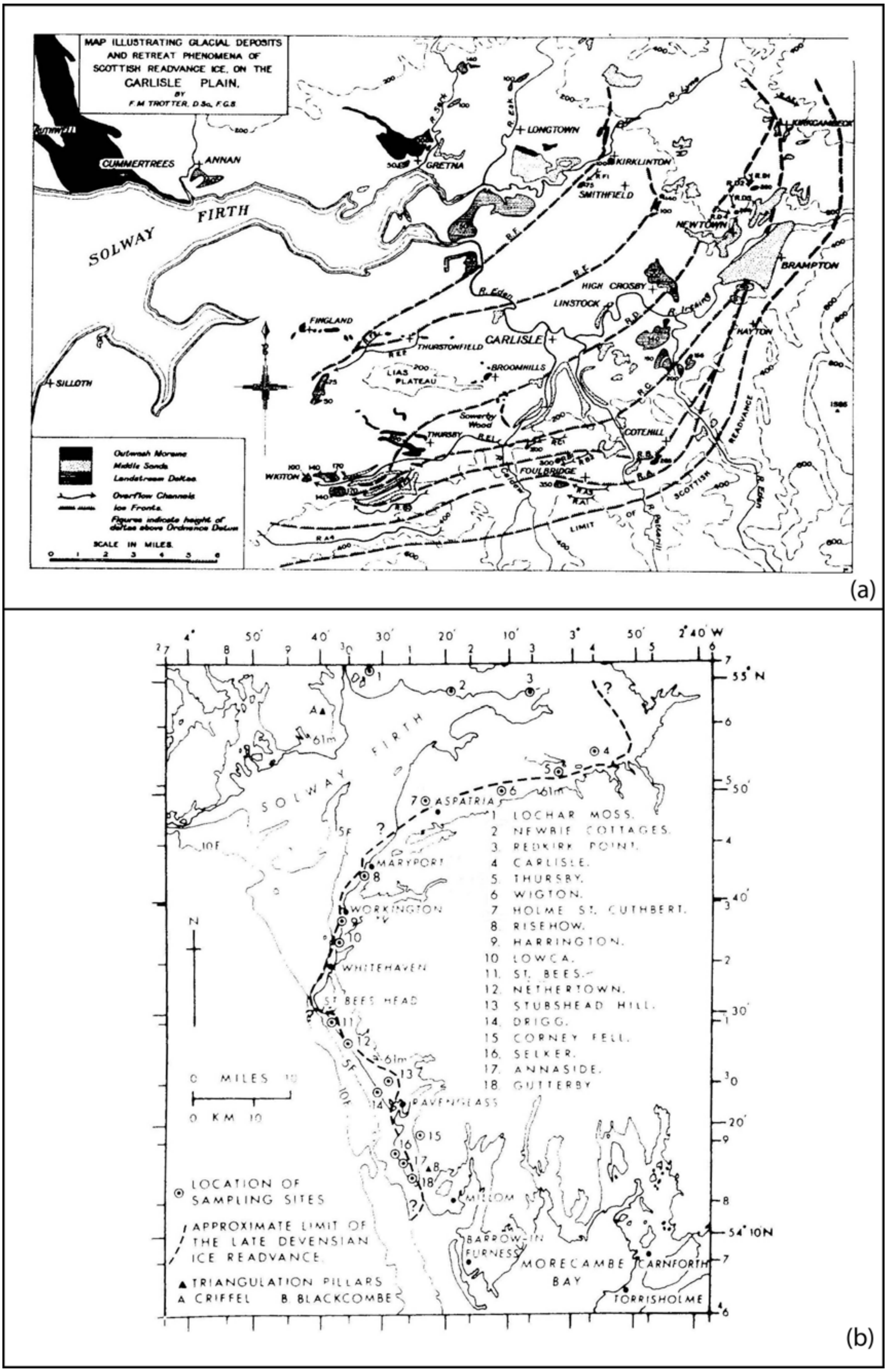


Fig. 3: NEXTMap DEM of Holme St Cuthbert glaciofluvial complex superimposed over an arcuate set of lineations. NEXTMap Britain data from Intermap Technologies Inc were provided courtesy of NERC via the NERC Earth Observation Data Centre (NEODC).

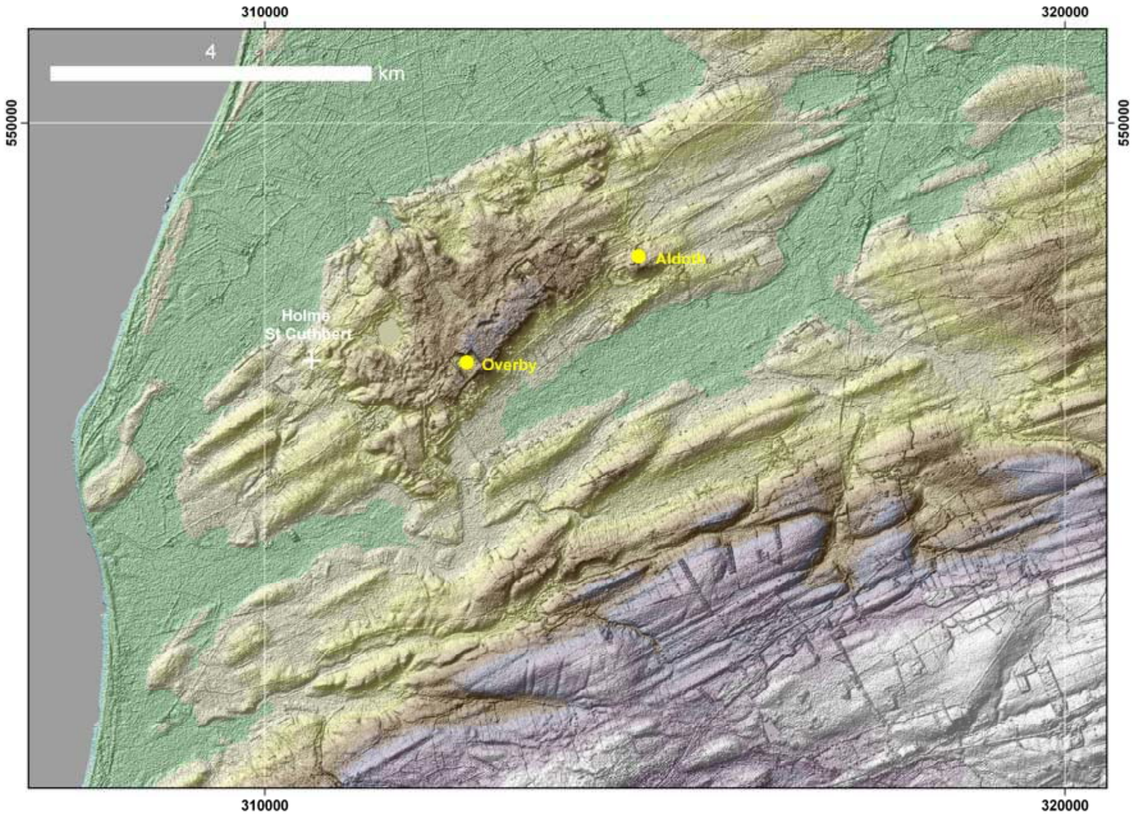


Fig. 4: Stratigraphic logs of Overby sand pit

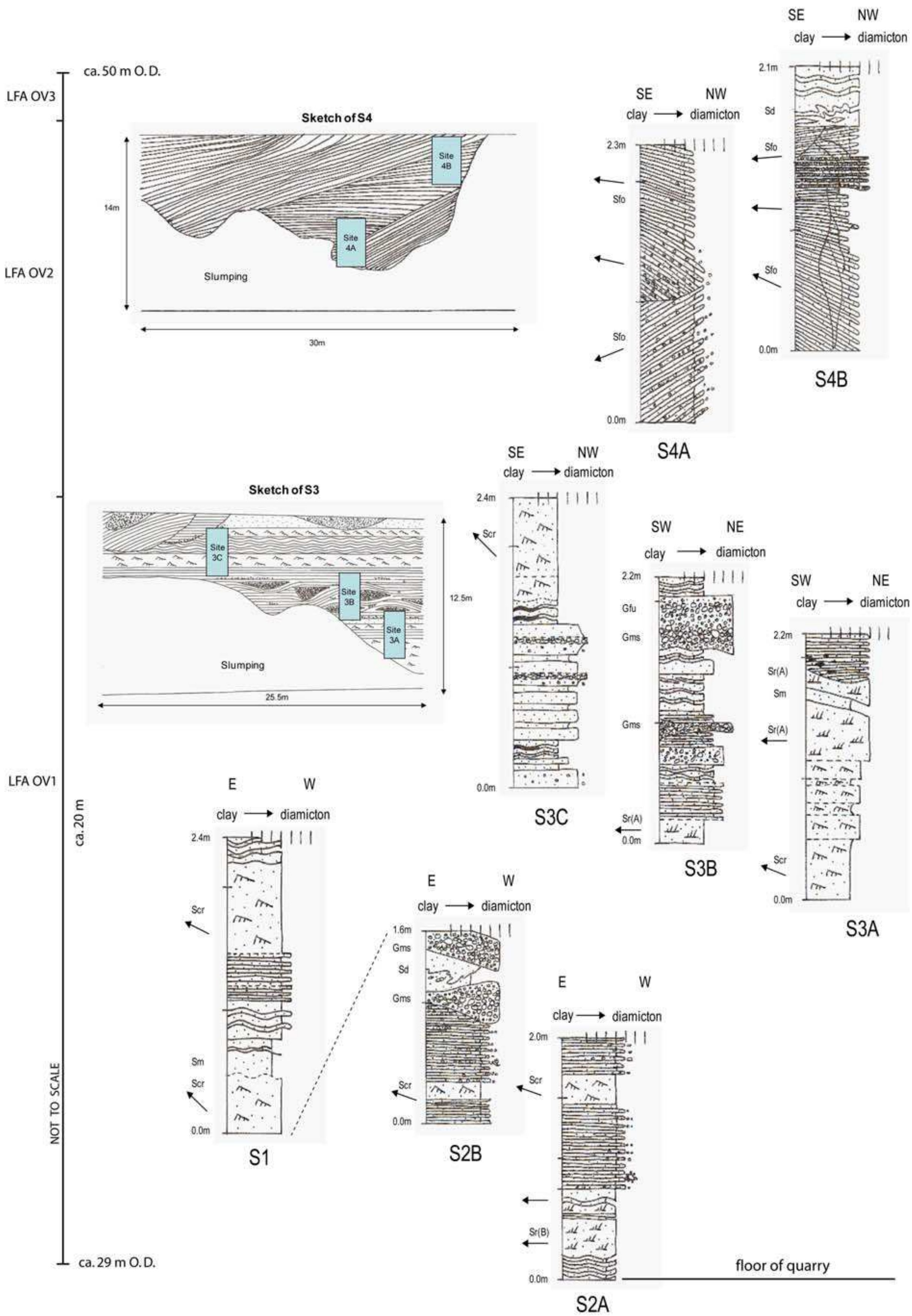


Fig. 5: Photographs:

a. Looking SW down the ice-contact slope at Overby

b. Subdued esker near Thursby (with sharp bends)

c. Trotter & Hollingworth (1932): picture of delta sedimentology at Warwick Bridge delta (foresets and topsets)

(a)

(b)

(c)

Fig. 6: Photographs: Overby and Aldoth sand pits: (a) LFA OV1: fine grained deposits (including ripple structures); (b) LFA OV2: foresets; (c) LFA OV3: topsets; (d) LFA A1; (e) Composite of LFA A2; and (f) LFA A3

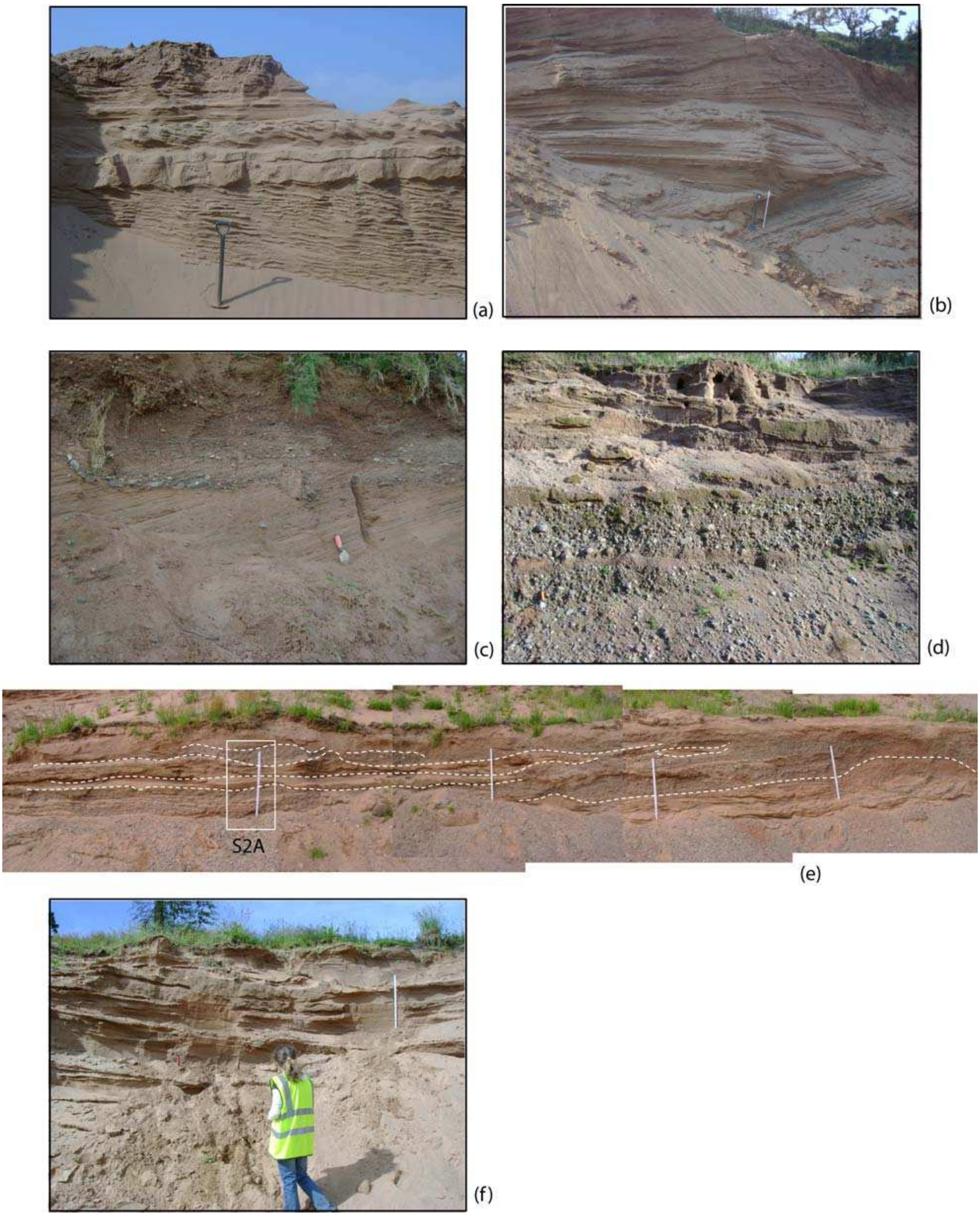


Fig. 7: Aldoth sand quarry – stratigraphic logs

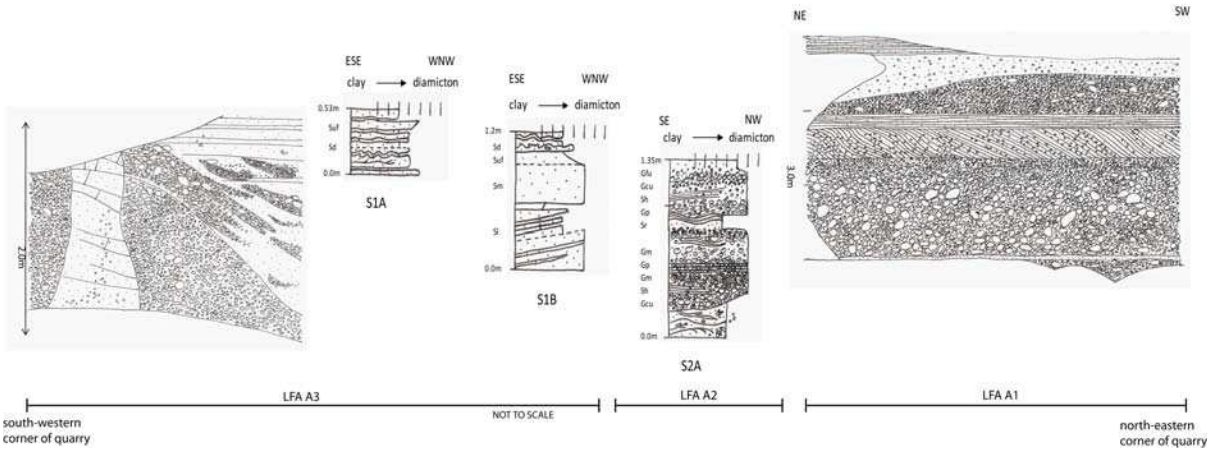


Fig. 8: Plumpe Farm – stratigraphic logs and clast macrofabric

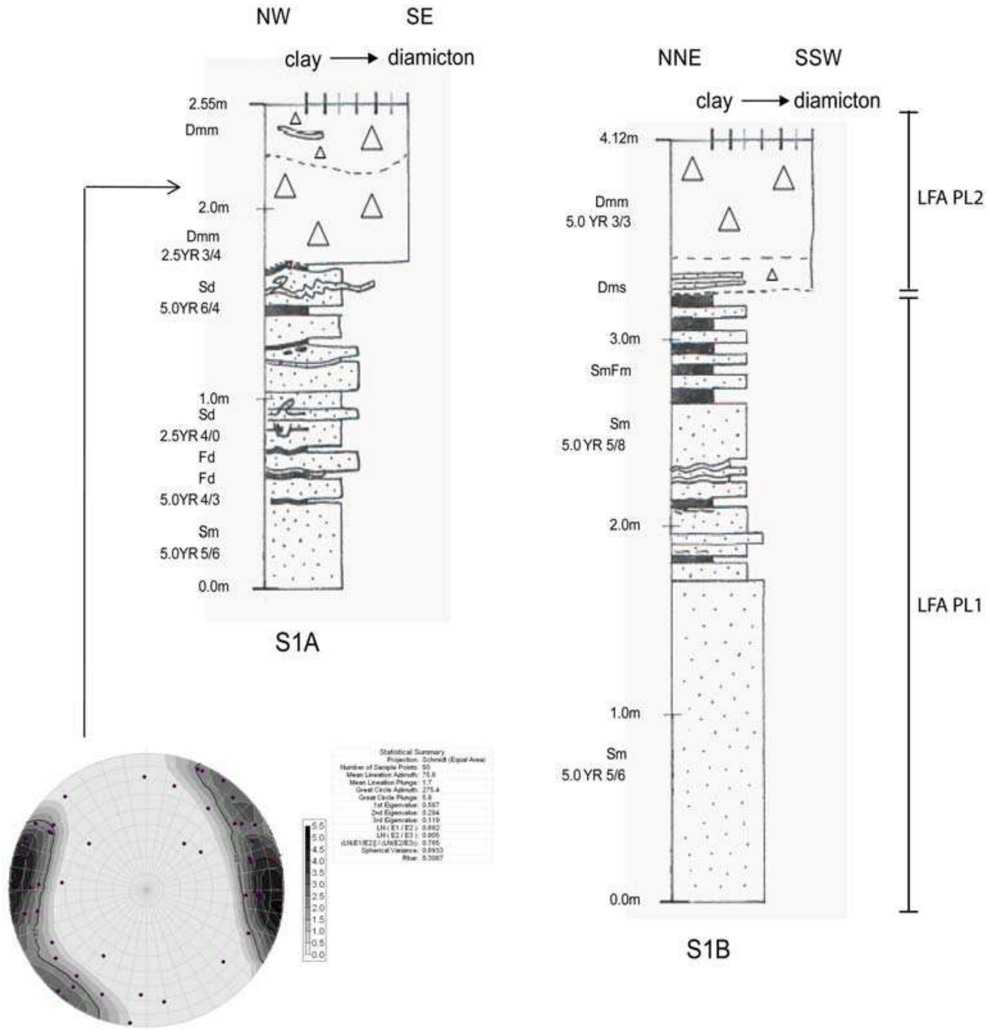


Fig. 9: Photographs: Plumpe Farm and Carleton: (a) LFA PL1; (b) LFA PL1: deformation; (c) LFA CA1; (d) LFA CA2; and (e). LFA CA3.

Fig. 10: Carleton – stratigraphic logs and clast macrofabric

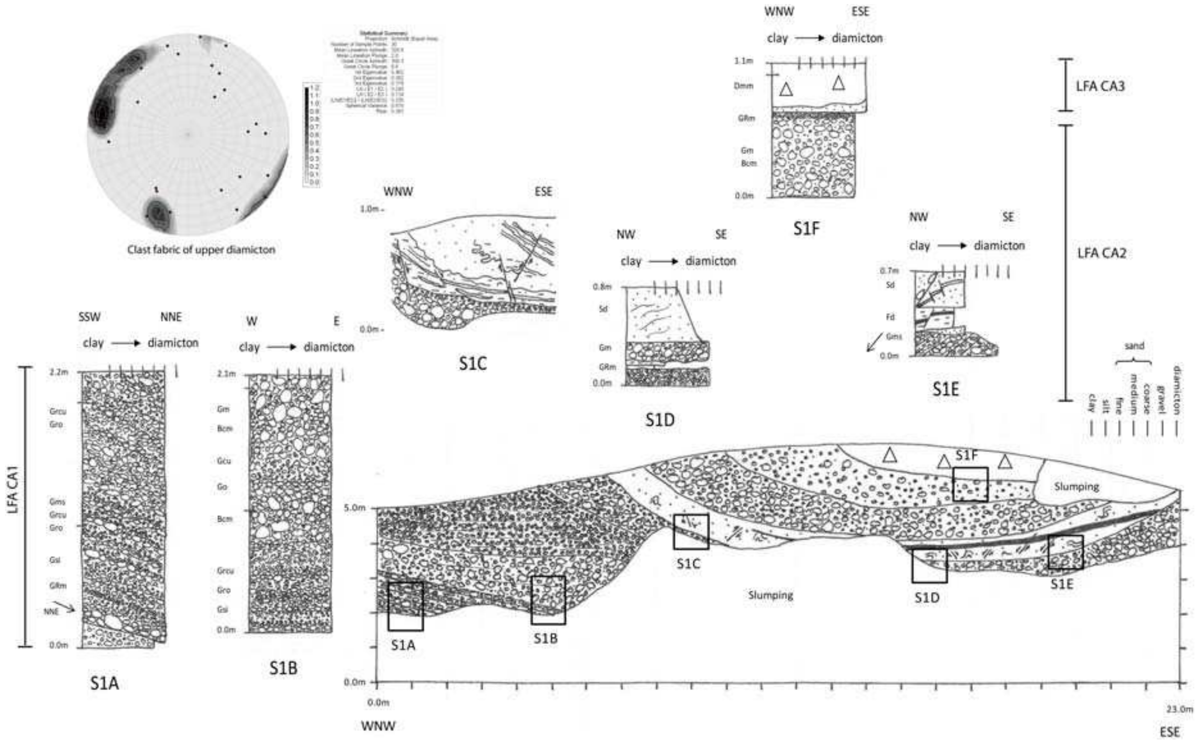


Fig. 11: (a) 3D NEXTMap image of the glacial meltwater channels of Biglands, Wiza Beck and Wampool incised into an arcuate flowset trending round the northern margin of the Lake District; (b) meltwater channel long-profiles. Biglands meltwater channels (long profiles taken NW-SE); Wiza Beck/Wampool meltwater channels (long profiles taken W-E). NEXTMap Britain data from Intermap Technologies Inc were provided courtesy of NERC via the NERC Earth Observation Data Centre (NEODC).

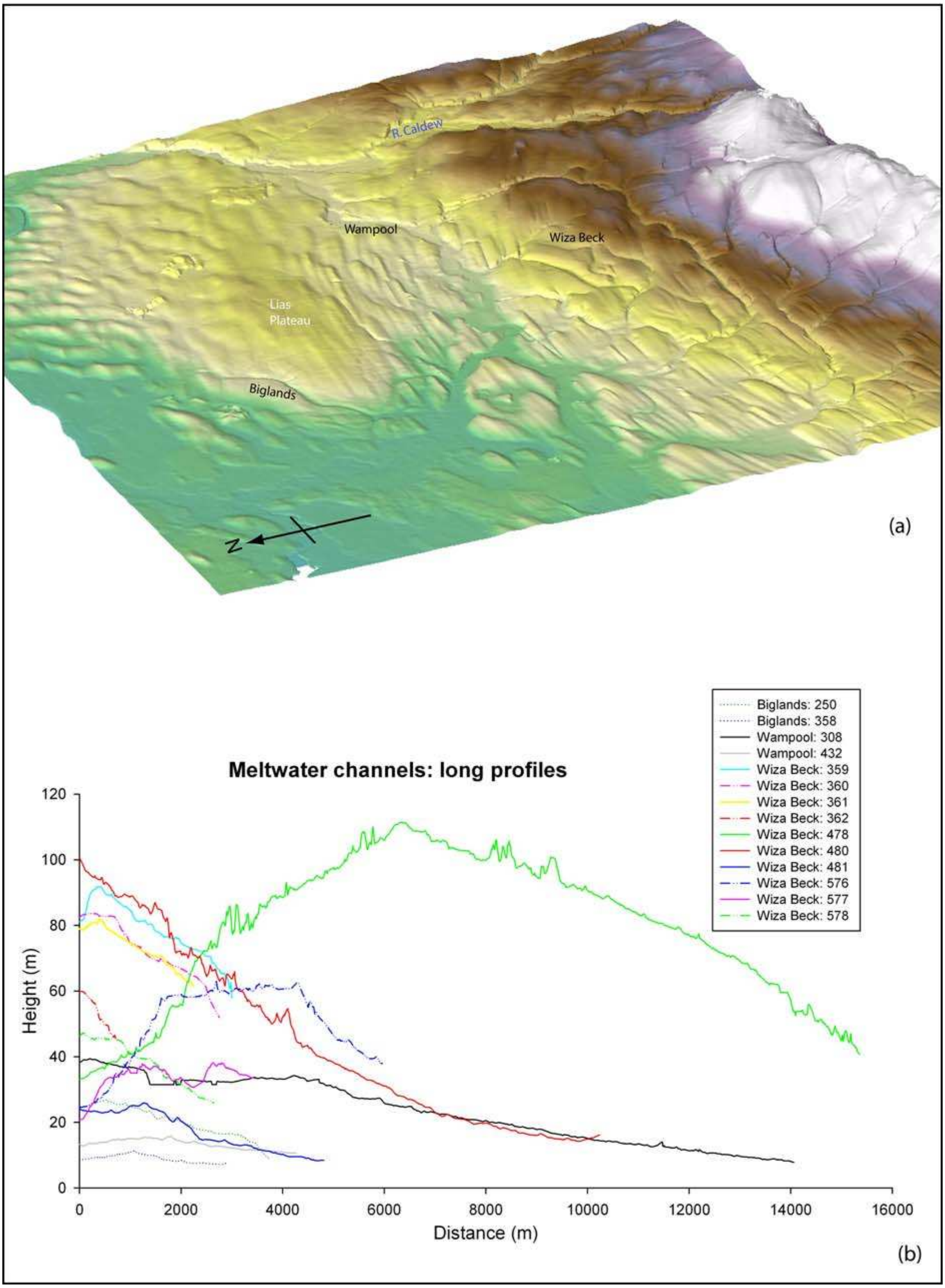
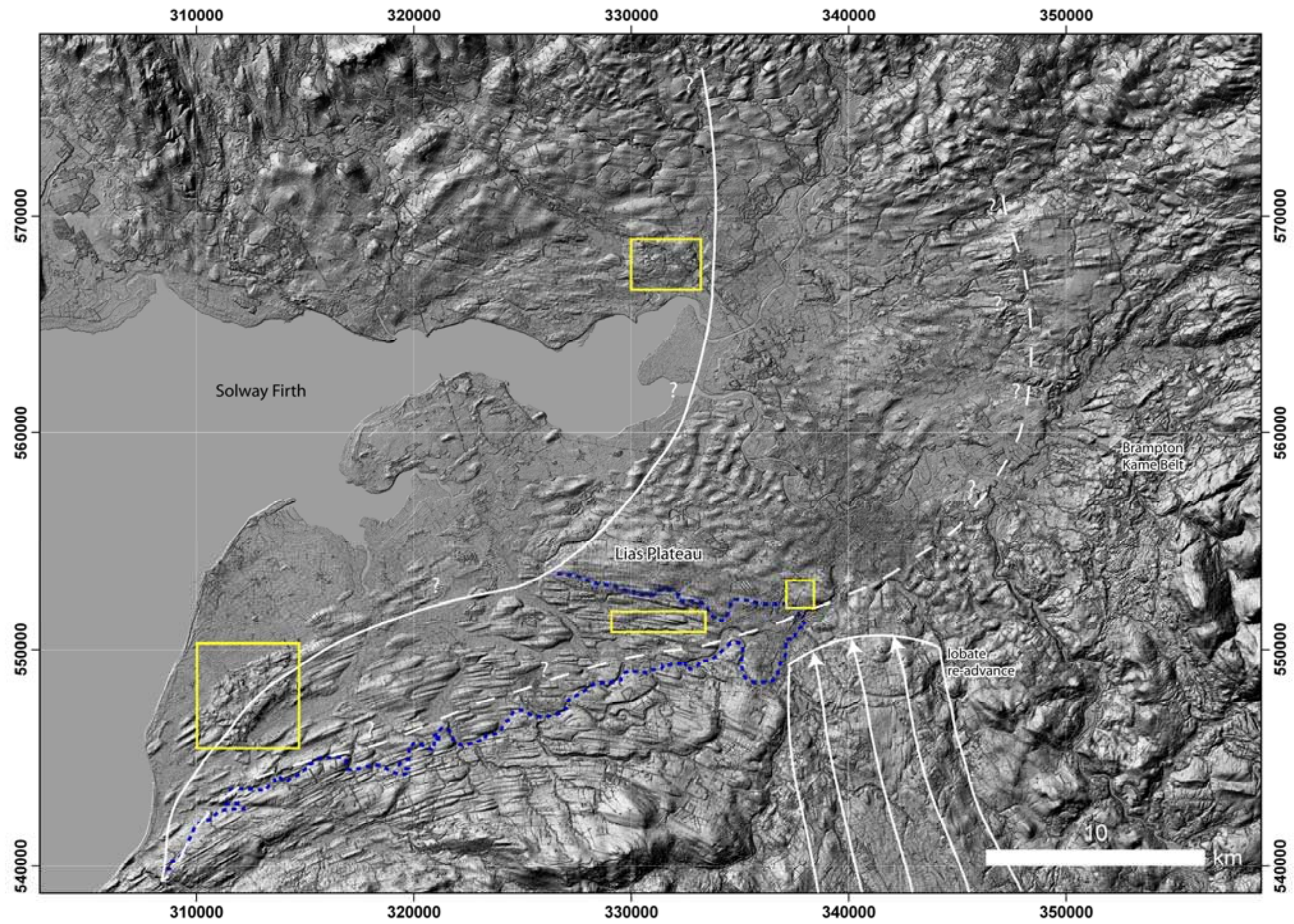

R. Caldew
Wampool
Wiza Beck
Lias Plateau
Biglands
N
(a)
Meltwater channels: long profiles
Height (m)
Distance (m)
Biglands: 250
Biglands: 358
Wampool: 308
Wampool: 432
Wiza Beck: 359
Wiza Beck: 360
Wiza Beck: 361
Wiza Beck: 362
Wiza Beck: 478
Wiza Beck: 480
Wiza Beck: 481
Wiza Beck: 576
Wiza Beck: 577
Wiza Beck: 578
(b)

Fig. 12: Reconstruction of Scottish re-advance into the Solway Lowlands. The yellow boxes refer to glacial landforms (glacial outwash delta and eskers at Gretna, Sowerby Wood and Thursby eskers and Holme St Cuthbert deltaic sequence) which can be confidently correlated with the Scottish re-advance. The white line demarcates the major stillstand position (see text), while the white dotted line refers to the rapid, transient lobe of thin ice, which spread out across the Solway Basin. The blue-dotted line extent of the lake associated with the Holme St Cuthbert delta (delineated by the height of the foresets). The white arrows refer to the lobate advance of Vale of Eden ice into the Solway Lowlands. NEXTMap Britain data from Intermap Technologies Inc were provided courtesy of NERC via the NERC Earth Observation Data Centre (NEODC).



Tables:

Table 1: Ripples structures

Exposure	Location in section (m's from bottom of exposure)	Type of ripple	Angle	Palaeo-current	Contact	Texture
2A	0 – 0.5	A	Transverse, sinusoidal, out of phase	W to E	Grades into ripples from sinusoidal structures	Fine-medium sand
	1.47 – 1.67	A	16º	W to E		Fine-medium sand
2B	0.2 – 0.4	A	16º	W to E		Fine-medium sand
1	0 – 0.48	B	36º – 40º	W to E	Grades up into coarse sand	Fine-medium sand
	1.5 – 2.2	A	20º – 24º	W to E	Erosional	Fine-medium sand
3A	0 – 1.15	A	6º – 12º	NNE to SSW	Grades between sand	Grades upwards from fine to coarse sand
	1.15 – 1.65	A	Transverse, sinusoidal, out of phase	NNE to SSW	Erosional	Fine-coarse sand
3B	0 – 0.22	A	Transverse, sinusoidal, out of phase	NNE to SSW	Grades into laminations	Fine-medium sand
3C	1.65 – 1.85	A	20º	NNW to SSE	Grades up from sinusoidal structures	Fine-medium sand
	1.85 – 2.4	B	20º	NNW to SSE	Grades from ripples (1.65 – 1.85)	Fine-medium sand