Ice sheet advance, dynamics, and decay configurations: evidence from west central Scotland.

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ABSTRACT

A 3700 km^2 area adjacent to the Firth of Clyde, Scotland, is examined to constrain the development and dynamics of the western central sector of the last British and Irish Ice Sheet. Results from geomorphological mapping, lithostratigraphic investigations, three-dimensional geological modelling and field observations are combined to produce an empirically constrained, five-stage conceptual model of ice sheet evolution. (A) Previously published dates on interstadial organic deposits and mammalian fossils suggest that the Main Late Devensian (MLD) (MIS 2) glaciation of central Scotland began after 35 ka cal BP. During build-up, ice advanced from the western Scottish Highlands into the Clyde and Ayrshire basins. Glaciomarine muds and shelly deposits scavenged from the Firth of Clyde were redeposited widely as shelly tills and glacial rafts. Ice advance against reverse slopes generated, and subsequently overtopped, ice-marginal sediment accumulations. We hypothesise that some of these formed pre-cursor ridges which were moulded into suites of ribbed moraine during the glacial cycle. (B) Sustained stadial conditions at the Last Glacial Maximum (LGM) (c. 30 - 25 ka cal BP) resulted in development of a major dispersal centre over the Firth of Clyde and Southern Uplands. This dispersal centre locally preserved previously-formed subglacial bedforms, and fed a wide corridor of fast-flowing ice east towards the Firth of Forth. (C) Initial deglaciation promoted a substantial re-configuration of the ice surface, with enhanced westward drawdown into the outer Firth of Clyde and eastward migration of an ice divide towards the Clyde-Forth watershed. (D) Renewed ice sheet thickening over the Firth of Clyde may have accompanied growth of the Irish Ice Sheet during the Killard Point Stadial (c. 17.1 - 15.2 cal ka BP); it was associated with limited bed modification. Subsequent ice sheet retreat was characterised by substantial meltwater production, ponding and erosion. (E) Late stages of MLD ice sheet retreat were punctuated by one or more significant ice margin oscillations. Discovery of De Geer moraines at the site of a former proglacial lake in western Ayrshire allows glacier flow at the ice margin to be approximated as ≤ 290 m a^{-1} during one such oscillation. Such velocities were probably enabled by basal sliding and shallow sediment deformation. At this stage those parts of the MLD ice sheet margin that were grounded in the Firth of Clyde were extremely vulnerable to collapse. Final disintegration of glacier ice in the Clyde basin probably occurred early in the Lateglacial Interstadial (Greenland Interstadial-1), coinciding with marine incursion to c. 40 m above present day sea level.

1. Introduction

Detailed geomorphological investigations, aided by increasingly powerful remote sensing datasets, have revealed complex flow signatures from former ice sheets (e.g. Clark and Stokes, 2001; De Angelis and Kleman, 2007; McCabe, 2008; Greenwood and Clark, 2009a). Such evidence is essential in order to test and refine numerically-driven ice sheet models, which can simulate dynamic cycles and major ice flow configuration changes (e.g. Boulton and Hagdorn, 2006; Hubbard et al., 2009). The last British and Irish Ice Sheet (BIIS) is now known to have undergone substantial changes in geometry and flow during its evolution (Bowen et al., 2002; Bradwell et al., 2008; Evans et al, 2009; Greenwood and Clark, 2009b). However, terrestrial evidence is fragmentary, and coherent, time transgressive reconstructions are lacking for many key sectors. Interpretations conflict owing to the isolated nature of individual studies, and uncertainty remains over whether events identified in the geological record were local phenomena resulting from internal glacier readjustments, or ice sheet-wide events controlled by climatic response (e.g. Sissons, 1964, 1976; Paterson, 1974; McCabe et al., 2007b; Peacock et al., 2007).

This study attempts to address these issues for west central Scotland (Figs 1, 2) - an area of some 3700 km^2 which was subjected to interactions between major accumulation zones of the last BIIS. By combining geomorphological, lithostratigraphical, and three-dimensional geological modelling investigations with existing research, we reassess the palaeoglaciology of this formerly dynamic ice sheet zone. Specifically, this paper aims to: (i) identify evidence for spatially and temporally variable ice flow patterns, major geometry changes, and oscillatory events that accompanied build up and decay of the BIIS in west central Scotland; (ii) take account of published evidence from surrounding areas to test for indicators of more widespread ice sheet reorganisation(s); (iii) provide a coherent, conceptual model of ice sheet evolution for this zone of the last BIIS; and (iv) enable better understanding of the three-dimensional distribution of glacigenic, glaciolacustrine and glaciomarine sediments beneath greater Glasgow required for applied geological investigations.

2. Background

This section reviews the key published evidence relating to the Main Late Devensian (MLD) ice sheet glaciation of west central Scotland. In this paper, dates are quoted in radiocarbon years and calibrated to calendar years before present where appropriate, using the curves of Stuiver et al. (2005) and Fairbanks et al. (2005). Radiocarbon ages from marine samples are quoted assuming a 400-year reservoir age correction, except where indicated otherwise.

The western central lowlands of Scotland have been recognised as an area affected by complex ice-flow patterns since J. Geikie (1863) described it as the "debatable ground over which glaciers of the Highlands or Southern Uplands (Fig. 1) prevailed according to their contemporary strengths". Highland-sourced ice initially extended south-eastwards to the northern flanks of the Southern Uplands as suggested by the distribution of indicator erratics and the widespread occurrence of 'lower' tills of north-western provenance underlying 'upper' tills derived from the south (Price, 1975; Sutherland and Gordon, 1993). Southern Uplands ice became more dominant later, deflecting Highland ice both to the east and south-west. The easterly diverted flow left a strong imprint on the landscape of the Clyde basin, in the form of extensive, drumlin assemblages (Rose, 1987; Hall et al., 1998; Rose and Smith, 2008).

Former changes in ice flow direction are particularly apparent in central Ayrshire where evidence from erratics, stratigraphy, cross-cutting striae and roches moutonées demonstrate that ice firstly flowed onshore towards the south-east and then offshore south-westwards (Richey et al., 1930). Highland-sourced ice initially penetrated at least as far south as Nith Bridge (Fig. 2), carrying shells that were probably scavenged from the Firth of Clyde (Holden and Jardine, 1980; Sutherland, 1993). This early, on-shore flow of ice resulted in widespread deposition of shelly tills and rafts of glaciomarine clay, notably at Afton Lodge and Greenock Mains (Fig. 2) (Smith, 1898; Holden, 1977; Abd-Alla, 1988; Gordon, 1993a,b). Evidence for a later, onshore readvance of Highland-sourced ice at the latter locality (Holden, 1977) implies that active ice occupied the Firth of Clyde and Ayrshire Lowlands after the eastern central lowlands had deglaciated (Sutherland, 1984).

Most evidence for these switches in flow relates to the last, MLD glaciation when the BIIS is thought to have reached the continental shelf edge to the north-west of the British Isles and merged with ice from Fennoscandia in the North Sea Basin (Graham et al., 2007; Bradwell et al., 2008). However, there is an unusually high concentration of mammalian fossil occurrences and other organic remains within glacigenic sequences that have survived the MLD glaciation, particularly in the Ayrshire Lowlands and the lower Clyde valley (Bishop and Coope, 1977; Sutherland and Gordon, 1993). Early age determinations from bones of woolly rhinoceros from Bishopbriggs (Rolfe, 1966) and reindeer antler fragments from Sourlie (Fig. 2) (Jardine et al., 1988) yielded ages of c. 27 -30 ka ^{14}C BP. A revised age of 31.1 ka ^{14}C BP (c. 35 ka cal BP) has recently been published for the Bishopbriggs sample, following ultrafiltration pre-treatment (Jacobi et al., 2009). This age is similar to those obtained from organic remains within fully investigated interstadial profiles beneath till at Balglass (Fig.1) (Brown et al., 2007) and Sourlie (Bos et al., 2004). Collectively, these dates suggest that the MLD ice sheet did not become established in the area until after c. 35 ka cal BP, contrary to the conclusions of Bowen et al. (2002) . However, numerical modelling experiments simulate minor glacial advances into the area prior to the main sustained advance in the Late Devensian (Hubbard et al., 2009). Indeed, the moderately-weathered Lawthorn Diamicton that underlies the interstadial deposits at Sourlie, and the Ballieston Till Formation (see below), which occurs within concealed depressions beneath Glasgow, must relate to an earlier expansion of glacier ice.

There is general agreement that the MLD ice sheet withdrew towards the west and northwest during deglaciation of the area (Price, 1983; Sutherland, 1984). Ice-marginal lakes formed where ice impeded drainage, firstly on the watershed between the catchments of the Avon Water and River Irvine, in Glengavel (Fig. 2), where laminated glaciolacustrine silts occur to an elevation of at least 205 m above sea level (a.s.l.) (Nickless et al., 1978). This lake came into existence shortly after Highland and Southern Uplands-sourced ice had separated (Phemister in Richey, 1926; McLellan, 1969; Martin, 1981). Water held within the upper Avon Valley merged with a much larger lake, 'Lake Clydesdale' (Bell, 1874), which eventually occupied the Clyde valley and its tributaries upstream of Glasgow. The level of Lake Clydesdale probably dropped in stages as north-westward retreat of the ice margin in the lower Clyde Valley made available spillways to the east at progressively lower elevations of 200, 165, 102, and 85m a.s.l. respectively (Paterson et al., 1998, fig. 12). The lake finally drained eastwards via a col in the upper Kelvin valley at about 45 m a.s.l. (Forsyth et al., 1996; Hall et al., 1998).

Lake Clydesdale probably existed during the creation of the Main Perth Shoreline in the Forth estuary (Sissons and Smith, 1965; Sutherland, 1984), and possibly into the beginning of the Lateglacial Interstadial (GI-1) (Peacock, 1999, 2003). The timing, contemporary sea level and manner in which the late-glacial sea eventually invaded Lake Clydesdale is disputed (see Peacock, 2003 for review), but it is generally accepted that the transgression had occurred by 13.1 - 12.5^{14}C ka BP (based on uncorrected, reported ages from marine shells) (Peacock, 1971, 2003; Peacock et al., 1977; Browne et al., 1977; Rose, 2003). The presence of a radiocarbon plateau at this time, and uncertainty regarding the reservoir correction at the Greenland Stadial-2 (GS-2) to GI-1 boundary, preclude a more precise chronology based on ^{14}C dates alone (Peacock, 2003). Final disintegration of ice blocking the Clyde estuary resulted in the level of Lake Clydesdale falling from 45 m to no less than 40 m, the contemporary sea level (Peacock, 2003). Sea level then fell rapidly, but further discussion of the subsequent, complex sea-level history of the area is not presented here, nor discussion of events during the Loch Lomond Stadial (Greenland Stadial-1, GS-1), when ice readvanced

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from the Loch Lomond basin to the northwest of Glasgow (Rose et al., 1988; Evans, 2003; Rose and Smith, 2008).

3. Lithostratigraphy of west central Scotland

Despite the success of modern geomorphological analysis in distiguishing the relative ages of landforms, it is essential also to consider the known sequence of deposits (lithostratigraphy) in order to determine a robust event stratigraphy. Detailed lithostratigraphical knowledge also underpins sophisticated three-dimensional modelling of Quaternary deposits (e.g. J.E. Merritt et al., 2007). The lithostratigraphy for the Clyde and Ayrshire basins presented here (Fig 3) develops that of Sutherland (1999) and follows a new top-down, nationwide framework (McMillan et al., 2005, in press; www.bgs.ac.uk/lexicon).

The Clyde lithostratigraphy is based on formations proposed by Rose (1981, 1989) and Browne and McMillan (1989a). The lowermost **Ballieston Till Formation** consists of consolidated sandy silty clay diamicton with isolated boulders and pebbles. It is dark greyish brown at depth, but reddish brown at the surface possibly due to weathering (oxidation). Sections in the till revealed numerous joints, many of which were striated and polished on their surfaces. Boreholes and temporary sections from Glasgow revealed up to 15 m of consolidated, laminated, unfossiliferous, silty glaciolacustrine clays with dropstones (**Broomhill Clay Formation)** overlying the Ballieston Till (Browne and McMillan, 1989a). These authors suggested that the laminations are varves, representing 600 - 1000 years of sedimentation. Importantly, these glaciolacustrine clays occur to depths of c. 25 m below present sea level, requiring a low contemporary sea level at the time of deposition. Where observed, the Broomhill Clay is overlain by the regional till of the area, the **Wilderness Till Formation**. However, in parts of northern Glasgow, the latter rests on bedded, bouldery gravelly sands of the **Cadder Sand Formation**. These sands have yielded bones and teeth of woolly rhinoceros, (Rolfe, 1966). The Wilderness Till is described by Rose et al. (1988) as a defomation till, but it also includes tectonised thrust slices of sand and laminated clay from underlying units. In Glasgow, it is a sandy silty clay diamicton with pebbles and isolated boulders. The colour varies, depending on local bedrock. In eastern Glasgow the Wilderness Till is overlain by the **Broomhouse Sand and Gravel Formation**, much of which forms ice-contact topography (eskers, mounds, flat-topped kames and kettleholes). These deposits have been extensively removed for aggregate. The Broomhouse Sand and Gravel Formation includes deltaic sands and glaciolacustrine laminated clays (**Ross Sand** and **Bellshill Clay** members, respectively), which were deposited in an ice-dammed lake, 'Glacial Lake Clydesdale' (Bell, 1874), whilst the MLD ice sheet margin retreated from the position of eastern Glasgow.

In west central Scotland, raised glaciomarine deposits of late-glacial age are assigned to the **Clyde Clay Formation** (McMillan et al., in press) in which two principle members are recognised, the Paisley Clay and Linwood Clay. The former member generally comprises thinly laminated clays and silts with dropstones. This member is generally poor in fauna, only yielding the cold-water foraminifera *Elphidium clavatum* in significant numbers. It has been mapped in areas around the Clyde estuary up to altitudes of c. 40 m a.s.l. (Browne and McMillan, 1989a). The Linwood Clay Member is confined to western areas of the Clyde estuary where it commonly overlies the Paisley Clay. It consists of more thickly bedded silts and clays with a richer faunal assemblage. (e.g. Browne and McMillan, 1989a; Peacock, 2003).

A gravelly silty clay diamicton, the **Gartocharn Till Formatio**n, occurs around the southern shores of Loch Lomond, locally including marine foraminifera and broken marine shells entrained by erosion of units from the Clyde Clay Formation (Rose et al., 1988). Plant detritus found beneath the Gartocharn Till has been radiocarbon dated at 10.6 ^{14}C ka BP (c. 12.5 cal ka BP) (Rose et al., 1988), confirming that the till was deposited as glacier ice readvanced during the Loch Lomond Stadial (GS-1).

The Quaternary lithostratigraphy of central Ayrshire follows McMillan et al. (in press) and is based mainly on the succession that was exposed in an opencast coal site at Sourlie, near Irvine (Fig. 2) (Jardine et al., 1988; Sutherland, 1999). The site was excavated into the north-western side of Sourlie Hill, one of a swarm of broadly eastward orientated drumlins. The importance of the site lay in the discovery of thin lenses of organic material (**Sourlie Organic Silt Formation**) occurring between two units of till. These lenses yielded a very rich flora and fauna deposited within a shallow pond in a treeless, low-shrub to sedge-moss tundra environment, and included bones of woolly rhinoceros and reindeer. Radiocarbon dates on antler fragments, plant debris and bulk organic matter suggest a Middle Devensian age (Bos et al., 2004).

The basal unit comprises up to 7.5 m of very stiff, dark grey, silty sandy stony clay diamicton ('lodgement till'), the **Littlestone Till Formation**, which locally encloses deformed sheets (glacial rafts?) of sand up to 7.5 m thick. The till is overlain by up to 3.5 m of unstratified, clay-rich gravel and clayey sand (**Lawthorn Diamicton Member** of the Littlestone Till Formation), interpreted as an 'ablation deposit' by Jardine et al. (1988), but probably better described today as glacigenic debris flow deposits. The Lawthorn Diamicton is overlain by up to 5.5 m of the partially cross-stratified **Armsheugh Sand and Gravel Formation** that is interpreted by Jardine et al. (1988) to have formed as glaciofluvial outwash. The Middle Devensian Sourlie Organic Silt Formation occupies shallow depressions within the surface of the sand and gravel. The organic deposits are overlain by up to 3.5 m of pinkish brown, very stiff, pebbly sandy silty clay diamicton containing clasts of local sandstone, mudstone, coal and dolerite, 'far travelled sedimentary, igneous and metamorphic rocks' and shell fragments, including sparse paired valves of marine molluscs yielding Late Devensian amino acid ratios (Jardine et al., 1988). This unit, the **Eglinton Shelly Till Member** of the Wilderness Till Formation, is correlated here with other widespread occurrences of shelly till in Ayrshire (Smith, 1898; Sutherland and Gordon, 1993), that elsewhere contains rafts of cold-water marine silts and clays, notably at Afton Lodge, near Ayr (Gordon, 1993a). The uppermost glacigenic unit at Sourlie comprises up to 12 m of stiff, dark grey 'lodgement till', the **Auchenwinsey Till Member** of the Wilderness Till Formation. It forms most of the drumlin into which the opencast site was excavated.

4. Methods

4.1. Remote sensing evidence

Remote sensing datasets were interrogated within ESRI Arc Map 9.2. Digital surface models (DSMs) and georectified 1:10,000 monoscopic aerial photographs were analysed to identify glacial landforms in the study area. The surface models, built from NEXTMap® Britain topographic data (1.5 m vertical and 5 m horizontal resolution) were illuminated from the NW and NE to ensure capture of landforms with differing alignments. The DSM was analysed at several scales, ranging from 1:10,000 to 1:200,000. During larger-scale analyses, horizontal resolution was reduced to 50 m. Within the Glasgow area two additional versions of elevation data were used. One was a hill-shaded digital terrain model (DTM), for which data are processed to smooth abrupt surface features (e.g. buildings) allowing clearer (but less accurate) visualisation of the ground surface. The second was the unprocessed, orthorectified radar data, which was effective in picking out glacial landforms within built up areas.

4.2. Three-dimensional geological evidence

Over 60,000 borehole records exist for Glasgow and the surrounding area. The British Geological Survey (BGS) is currently creating a suite of three-dimensional Quaternary and bedrock models, based on the borehole data (J.E. Merritt et al., 2005; J.E. Merritt et al., 2007) using the modelling software tool, GSI3D (Sobisch, 2000; Kessler et al., 2006). In this study, outputs from these three-dimensional geological models were used for two purposes: (i) to confirm the basic composition of landforms, thereby enabling more confident discrimination between true glaciogenic features and bedrock controlled features; and (ii) to aid interpretation and identification in areas where subglacial landforms are masked by younger deposits (Fig. 4), or modified at surface due to urban development. A comprehensive UK database containing borehole records (BGS Borehole Geology) was interrogated throughout the investigation to provide additional information about the surface and subsurface sediments in the mapping area.

4.3 Field evidence

The greater Clyde basin area was resurveyed in the field at 1:10,000 scale over a 5 year period in the 1980s. The programme also included investigation of sedimentological, geotechnical and palaeontological characteristics of sixteen cored boreholes, and studies from numerous temporary sections. This work resulted in production of Quaternary geological maps of the region (Browne and McMillan, 1989b). These field data provide important constraints for the work presented here.

4.4 Data compilation

All landforms and mapped sediment distributions were captured in a spatially attributed ArcGIS© database. Landforms mapped include: ribbed moraine, streamlined bedforms, meltwater channels, moraine ridge complexes and narrow transverse ridges. Existing Quaternary geological maps of the area (e.g. British Geological Survey, 1987, 1993a, 1993b, 2002; Browne and McMillan, 1989b) were consulted throughout the study. A recently compiled 2-D digital geological map of Britain at 1:50:000 scale (DiGMapGB 50) was interrogated in the GIS and forms the basis for mapped distributions of glaciofluvial, deltaic, and raised marine sediments, and areas where bedrock occurs at or near the surface.

5. Results

A glacial geological and geomorphological map of the greater Clyde basin is shown in Figure 5. Some detail is lost reproducing the map at this scale, and numerous smaller features such as minor meltwater channels and individual moraine crests, are not shown. It is intended that a detailed, larger format glacial geological and geomorphological map will be available elsewhere. The morphological, spatial and, where known, basic sedimentological characteristics of landform assemblages are described below.

5.1. Ribbed moraine:

Suites of southwest to northeast aligned, broad transverse ridges occupy the Clyde and Ayrshire basins up to an elevation of c. 200 m a.s.l. These ridges are 0.4-1.2 km in width, 0.4-6.5 km in length and up to 40 m in height. On a morphological basis, the ridges can be described as ribbed moraine, and their dimensions are entirely consistent with those reported by Dunlop and Clark (2006). Occupying areas principally underlain by Carboniferous sedimentary rocks and basalts, the ribbed moraine maintain a long-axis alignment that does not concord with variations in bedrock strike. Three-dimensional geological models (e.g. Fig. 4) in the Clyde basin indicate that these ridges commonly consist of glacial sediments assigned to the Wilderness Till Fm; thus their form is not considered to be controlled by bedrock structure.

The ribbed moraine are extensively remoulded with development of, or modification into, elongate streamlined bedforms (described below). A 50-m-long temporary section within a broader zone of drumlinised ribbed moraine ridges was described by McMillan and Browne (1983) and Browne and McMillan (1989a). Here, a 2- to 5-m thick surface carapace of red-brown sandy clayey till of the Wilderness Till Fm sharply truncates a series of underlying sediments (Fig. 6). The lower sediments comprise till, bedded gravels, sands and clay, and form a series of thrust slices dipping steeply westwards. Two normal faults occur in these lower sediments, on the western side of the thrust stack.

5.2. Streamlined bedforms:

Streamlined bedforms are well developed within the Clyde and Ayrshire basins, around the margins of the Southern Uplands, and to the southeast of the Lochwinnoch Gap (Figs. 2, 5). In the latter area they occur where till is thin and patchy, and locally are strike parallel to the gently dipping Clyde Plateau Volcanic Formation. In that locality, bedforms are probably influenced by bedrock structure. Cover of Quaternary sediments is much thicker over the basins to the north and south, which are underlain mainly by Carboniferous sedimentary rocks. That bedforms in the Clyde Basin are of glacial origin is supported by three-dimensional geological models, which show that the landforms principally consist of glacial sediments assigned to the Wilderness Till Fm (Fig. 4). Many of the streamlined bedforms in the basins are superimposed on, or consist of, re-shaped sections of the ribbed moraine (Fig. 7).

Numerous geomorphologically-based ice sheet reconstructions use the approach of grouping bedforms into coherent ‘flowsets’ or ‘swarms’ (e.g. Boulton and Clark, 1990; De Angelis and Kleman, 2007; Stokes et al., 2009; Greenwood and Clark, 2009a). The streamlined bedforms identified here can be broadly divided into six flowsets based on their geographical distribution, trend, morphology and spatial relationships with other geomorphological features (Fig 8, Table 1):

5.2.1. Flowset I

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Streamlined bedforms assigned to flowset-I comprise a suite of drumlins trending towards the east and east-northeast, and are generally confined to the northeast side of the River Clyde. They have a mean length of 826 m, and a mean elongation ratio (ER) of 3.7. Bedforms in this group have been described previously by Rose (1987) and Rose and Smith (2008).

5.2.2. Flowset II

Streamlined bedforms assigned to flowset-II comprise a more subdued assemblage of drumlins trending towards the north and north-northeast. These bedforms curve along the northwest margins of the Southern Uplands, showing a very slight divergence at the elevated ground to the north of Greenock Mains. They are distinct from flowset-I on the basis of shorter length (mean: 563 m) and of lower ERs (mean 2.9). Flowset-II tentatively includes two similarly aligned, but more isolated streamlined bedforms to the northeast of Corse Hill.

5.2.3. Flowset III

Flowset III comprises ice-moulded bedrock and crag-and-tail forms over higher elevations and drumlins in the lower basin areas. They are well-preserved, and trend in a south to west-northwest direction, forming an overall convergent pattern towards the southwest (Fig. 8). Subsets (III-a – III-e) are identifiable within flowset-III on the basis of slight variations in alignment and differences in morphological characteristics. For example, subsets III-b and III-c have a considerably longer mean length (> 1 km) (Table 1). This probably reflects thinner till cover, and local concordance with strike of the gently dipping volcanic rocks. Although some of these bedforms may be influenced by bedrock structure over higher elevations, a consistent convergent trend is maintained in the lower-lying sediment-filled basins. The transition between individual subsets is largely gradational; thus, they are all incorporated within flowset-III.

5.2.4. Flowset IV

Flowset-IV comprises a small cluster of streamlined bedforms c. 12 km south from the Blantyreferme Moraine (see below). These bedforms trend in an east to east-southeast direction. They possess the shortest lengths (mean 321 m) of all the flowsets identified (Table 1).

5.2.5. Flowset V

Flowset-V comprises a well-defined suite of streamlined bedforms trending towards the east. They are generally confined to the western side of the Kilmarnock Moraine Belt (Fig 8) (see below); only a few isolated bedforms occur on the immediate eastern side. Streamlined bedforms belonging to flowset-V overprint those of flowset-III (Figs 8, 9)

5.2.6. Flowset VI

Flowset VI comprises well-defined drumlins trending in a southeast direction. They are confined to the Clyde basin, and extend eastward as far as the Blantyreferme Moraine (see below). Overprinting of flowset-VI bedforms onto the generally longer flowset-I bedforms is apparent in parts of the Clyde basin, a characteristic described by Rose and Letzer (1977) and Rose and Smith (2008).

5.3. Glaciofluvial assemblages:

Glaciofluvial assemblages described here include both moundy, kettled, ice-contact deposits and terraced outwash spreads, both assigned to the Broomhouse Sand and Gravel Formation. It is worthy of note that glaciofluvial deposits portrayed on BGS maps may include deltaic sequences that formed in ice-marginal or proglacial lakes; only widespread, fine-grained deposits are generally identified as 'glaciolacustrine'. Major belts of glaciofluvial deposits occur principally in the valleys of the Clyde, Kelvin and Avon (British Geological Survey, 1993a, 1993b, 1994). Many of the deposits in the southeast of the area are associated with systems of north-easterly-descending ice-marginal meltwater channels, notably southeast of Eaglesham and south of Strathaven (Richey et al., 1930; Paterson et al., 1998), where they fall from about 320 to 260 m a.s.l. (Fig. 10A). Further significant belts of mounds and undulating spreads of sand and gravel occur in the upper Ayr and Nith valleys (British Geological Survey, 1982). These deposits are also associated with coherent systems of ice-marginal meltwater channels, which in the upper Ayr valley descend from about 300 to 250 m a.s.l. towards the east. Further sets of eastward descending marginal meltwater channels exist in the Nith valley at altitudes from about 300 m down to 200 m a.s.l.

5.4. Moraine ridge complexes:

Three major moraine complexes have been identified in the Ayrshire and Clyde basins, at Kilmarnock, Blantyreferme and Eaglesham respectively (Figs. 2, 5). The Kilmarnock Moraine Belt (KMB, Fig. 5) extends for approximately 14 km in a south-southwest to north-northeast direction, and ranges from 5 to 20 m in height. To the northeast of Kilmarnock, the belt reaches a maximum width of nearly 800 m, where it forms multiple crests. Borehole records indicate that at least part of the complex comprises clay and sandy clay, while records immediately northwest of the ridge reveal till interbedded with sandy clay. Flat terrain immediately southeast of the complex is underlain by up to 7 m of sands, laminated silts and clays.

The Blantyreferme Moraine (BM, Fig. 5) was first recognised by Clough et al. (1911). Forming a near-symmetrical, cross-valley ridge, it is aligned south-southwest to north-northeast extending for over 2 km, and reaching nearly 20 m in height. Field mapping has revealed the feature to be of variable lithology, comprising till, sand and gravel and also laminated clay and silt (Browne and McMillan, 1989a).

The Chapelton Moraine Belt (CMB, Fig. 5) comprises a string of ridges and mounds that lie to the east of Eaglesham on the northern slopes of Corse Hill (Richey et al., 1930; Paterson et al., 1998). These landforms, which include esker fragments, were formed at the southern margin of Highland-sourced ice early in the deglaciation of the area. They descend eastwards from about 305 to 260m a.s.l. and were described by Sissons (1963, 1964, 1967a) as part of his more widespread evidence for the supposed 'Perth Readvance' (Simpson, 1933).

A further, near-coherent chain of moraine ridges occurs in the upper Ayr valley, above Greenock Mains. Individual ridges, up to 250 m in length, extend from c. 4 km north of Greenock Mains eastward for c. 8 km, declining in altitude from 310 to 270 m a.s.l. Suites of eastward declining, marginal meltwater channels occur on the northern and southern flanks of the upper Ayr valley, at altitudes of about 305 to 270 m a.s.l. Those on the northern side merge with the moraine ridges to the north of Greenock Mains (Fig. 10B).

5.5. Glaciolacustrine assemblages:

Extensive spreads of fine-grained glaciolacustrine sediment of the Bellshill Clay Member occur at surface southeast and east of the Blantyreferme Moraine (Fig. 5). Glaciolacustrine sediments also crop out locally on the western flanks of the Kelvin valley (Browne and McMillan, 1989a; Hall et al., 1988) and along the margins of the Irvine valley downstream of Darvel (Nickless et al., 1978).

The glaciolacustrine assemblages in the Clyde basin commonly pass up and laterally into flat-topped, deltaic deposits of the Ross Sand Member (Browne and McMillan, 1989a; Martin, 1981). Formerly exposed sections revealed sands and gravelly sands forming dipping foresets of Gilbert-type deltas (Fig 11, A and B). The deltaic deposits locally exceed 20 m in thickness in the eastern Clyde basin (Browne and McMillan, 1989a).

5.6. Narrow transverse ridges:

Two suites of previously unreported, closely spaced, narrow, linear transverse ridges occupy parts of the lower Irvine valley (Fig. 12) and the southern entrance to the Lochwinnoch Gap. The former, situated on the western side of the KMB trend broadly from south-southwest to north-northeast, and the latter trend broadly west to east. Those in the lower Irvine valley occur between about 30 m and 150 m a.s.l., while those south of the Lochwinnoch Gap lie between about 35 m and 140 m a.s.l. The former have a mean width of 75 m, a mean height of 2.4 m, and generally possess a symmetrical cross-profile (Fig. 13). Many of the ridges are continuous for over 400 m, maintaining their alignment across topographic undulations of up to 20 m in height. No sections have yet been observed within any of the landforms. However, borehole evidence from the Kilmarnock area demonstrates that surface sediments in the area of these narrow ridges comprise silts, sands and till. In places, the narrow transverse ridges are clearly superimposed on streamlined bedforms assigned to flowset V (Fig 12).

5.7. Raised marine deposits:

The lithostratigraphy of raised marine deposits in the Clyde basin has been described above. Field mapping has identified deposits of the Paisley Clay Member occupying extensive areas of the Clyde basin up to c. 40 m a.s.l. (Browne and McMillan, 1989a). Three-dimensional geological modelling supports the field interpretation and reveals thick spreads of silts and clay, often partially masking the underlying, drumlinised landscape (Fig. 4). Further discussion of the distribution of raised marine deposits and associated features is not presented here.

6. Interpretation of events in west central Scotland

6.1. Pre-Late Devensian Glaciation

Evidence for a pre-Late Devensian, MIS 4 or older glacial advance-retreat cycle has briefly been discussed in section 2. It led to the deposition of the Ballieston Till Formation in the Clyde basin, and the Littlestone Till Formation in Ayrshire; it probably involved a substantial advance of ice from the northwest. The apparently weathered top of the Ballieston Till suggests that there was a significant period of exposure before deposition of the overlying, glacitectonised, thinly laminated, glaciolacustrine sediments of the Broomhill Clay Formation in the Bellshill area (Figure 11C).

6.2. Late Devensian Glaciation; Stage A. (Fig.14A; build-up to LGM)

If the laminae in the Broomhill Clay Formation are correctly interpreted as varves, at least 600 to 1000 years elapsed before emplacement of the overlying Wilderness Till Formation (Browne and McMillan, 1989a). These glaciolacustrine sediments may document ponding during the earliest stages of ice advance into the area. Their occurrence to depths of 25 m below present sea level, suggests that at the contemporary relative sea level was at least 25 m lower than present because there is no known barrier that could have prevented marine invasion.

Instances exist where streamlined bedforms from each flowset are superimposed on the ribbed moraine ridges (e.g. Fig 7). Therefore, ridge formation must have occurred prior to the earliest phases of preserved streamlining in the study area. Regional geological evidence (e.g. Price, 1975; Sutherland, 1984; Sutherland and Gordon 1993) (Fig. 15) along with numerical ice sheet models (Hubbard et al., 2009) indicate that initial MLD ice-sheet advance into the area was from the northwest, broadly perpendicular to the ribbed moraine. We suggest this was the period of ribbed moraine formation (Fig 14A), when the ice front advanced against a reverse slope, building (then overtopping) sediment ridges through folding and thrusting of proglacial sediments. A similar mechanism is invoked for the formation of 'cupola hills' elsewhere (Benn and Clapperton, 2000; Benn and Evans, 1998). The sediments formerly exposed at Holmbrae Road in Glasgow (Fig. 6) (McMillan and Browne, 1983) are consistent with this interpretation. Initial advance lead to thrusting of the gravel, sand and clay beds in the eastern side of the section. The two normal faults may have been activated during a minor ice margin retreat, prior to overriding and deposition of the upper (Wilderness) till. The concept that some ribbed moraine originate as overridden ice marginal moraines has been proposed by Möller (2006). However, rigorous investigation of the sediments is required to test this hypothesis for these particular ribbed moraine suites.

That landforms from such an early stage of glaciation could survive is supported by preservation of Middle Devensian deposits in the area, together with the widespread occurrences of shelly diamicton (Eglinton Shelly Till) and glacial rafts of glaciomarine mud that were most likely scavenged from the Firth of Clyde during this early build-up stage. Their survival was probably aided by the development of an ice divide over the Firth of Clyde during the LGM (see below), beneath which there was minimal subglacial landscape modification.

6.3. Late Devensian Glaciation; Stage B. (Fig.14B; LGM)

The drumlins of flowset-I must have begun to form after ice in the Clyde basin had become sufficiently thick to over-top the main Clyde-Forth drainage divide, allowing fast, essentially non-topographically constrained ice flow beyond. A significant dispersal centre had developed over the Southern Uplands had occurred by this stage, contributing to deflection of ice in the Clyde basin toward the east, as evidenced by the well documented Dubawnt-type train (cf. Dyke and Morris, 1988) of essexite erratics that were dispersed from their source

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near Lennoxtown (Fig. 15) into the Firth of Forth (Peach, 1909: Shakesby, 1978, Evans et al., 2005). Striae patterns (Paterson et al., 1998) (Fig. 5) also document this flow.

Streamlined bedforms belonging to flowset-II were formed by north-eastward flow towards the Firth of Forth, driven by thicker ice to the southwest. This flow would have deposited Southern Upland till, with north-orientated clasts over the lower shelly tills (possibly of the Eglington Shelly Till Member) as described at Nith Bridge (Holden and Jardine, 1980; Sutherland, 1993) (Fig 16).

6.4. Late Devensian Glaciation; Stage C. (Fig.14C)

Streamlined bedforms from flowset-III document convergent south-westward and westward flow into the Firth of Clyde, and are consistent with patterns of glacial striations in the north Ayrshire basin (Paterson et al., 1998) (Fig. 5). This evidence suggests that a major change in the ice sheet configuration had occurred, largely caused by drawdown to the west. Subsets of flowset-III probably document a transgressive phase where part of the Southern Upland dispersal centre was scavenged by this increasing drawdown as ice flowed westward over the south of Arran. Westward transportation of Ailsa Craig erratics (Sissons, 1967a) (Fig. 15) would have occurred during this flow phase. By this stage, the north-eastward flow that generated flowset-II (Fig 14B - see above) must have switched off, allowing eastward migration of the ice divide, beneath which minimal subglacial modification was occurring.

6.5. Late Devensian Glaciation; Stage D. (Fig.14D)

A further, substantial alteration in ice sheet configuration and local basal conditions is indicated by the following suite of landforms: limited east-trending streamlining (flowset-IV), eastward descending marginal meltwater channels, eastward descending suites of ice contact glaciofluvial landforms (including the Chapelton Moraine Belt), and minor eastward-pushing morainic assemblages in the Ayr valley. Collectively, they demonstrate ice-divide migration back towards the west, coupled with ice-sheet thickening in the vicinity of the Firth of Clyde. The configuration is roughly that originally proposed for the 'Perth Readvance' in central Scotland (Sissons, 1963, 1964, 1967b). Of note is the limited bedform development from this stage, with only a small patch of Group IV bedforms occurring in the east. A minor readvance is apparent in the Ayr valley near Greenock Mains where a coherent assemblage of meltwater channels and moraine ridges indicate a late, north-eastward push (Fig. 10B) (Holden and Jardine, 1980).

6.6. Late Devensian Glaciation; Stage E. (Fig.14E)

A subsequent phase of more persistent streamlining is indicated by bedforms assigned to flowset-V and VI. These bedforms are generally longer than those of flowset-IV to the east (Table 1). Abrupt termination of flowset-V and VI just beyond (to the east of) the Kilmarnock and Blantyreferme moraine complexes demonstrates that the ice flow phase that generated them extended to these areas. This flow phase probably included, or was followed by ice margin stabilisation at the moraines. Cross-cutting of drumlins in the Glasgow area (e.g. Rose and Letzer, 1977) indicates that cessation of the formation of flowset-I drumlins must have occurred prior to formation of flowset-VI bedforms.

To the east of the Kilmarnock and Blantyreferme moraine complexes (which may or may not be contemporaneous features), vast suites of glaciolacustrine and deltaic sediments in the Clyde and Irvine valleys demonstrate existence of ice-dammed lakes. The narrow, closely-spaced ridges observed in the Kilmarnock area (Figs. 5, 12) and to the west of the Lochwinnoch Gap (Fig. 5) are similar in morphology to De Geer moraines, which form at, or close to the grounding line of calving glaciers (e.g. Lindén and Möller, 2005). The scale of the landforms is consistent with that of De Geer moraines, as is their pattern trending across topographic undulations (e.g. Todd et al., 2007). Given the abundant evidence for ice-dammed lakes in the area, and local borehole records of interbedded sands, silts and till, the landforms are interpreted as De Geer moraines.

A simple estimate of calving speed can be calculated across a hypothetical calving margin similar to the one indicated by the De Geer moraines at Kilmarnock. Warren and Kirkbride (2003) described an empirical linear relationship between water depth (D_W) and calving speed (U_C) for glaciers terminating in freshwater bodies:

$$U_C = 17.4 + 2.3D_W$$

D_W can be approximated from the c. 150 m a.s.l. upper altitude of De Geer moraines (proxy for lake surface altitude) and the base of the Ayr valley (30 m a.s.l.). Assuming a similar calving margin relationship, palaeo-calving rate is calculated to have been 293 m a^{-1} across the deepest part of the lower Ayr valley. Under steady state conditions (ice front not retreating, nor advancing), ice velocity at the margin would have been c. 290 m a^{-1}. These values are comparable with those of modern glaciers terminating in proglacial lakes on the eastern side of the South Patagonian Icefield (e.g. Warren and Aniya, 1999). Ice flow velocities of this order are consistent with the local development of well-preserved streamlined bedforms assigned to flowset-V (Figs. 5, 8). The survival, locally, of extended

Devensian sequences suggests that fast flow was enabled by basal sliding, and a relatively thin, near surface deforming layer.

The preservation of both pre-MLD sediments and landforms interpreted to have developed early in the MLD glaciation is intriguing. Both the Clyde and Ayrshire basins lay directly in the path of ice sheet advance, and were subjected to more than one phase of relatively fast glacier flow (described above). Despite thick 'soft sediment' sequences occupying these basins, it seems unlikely that widespread bed deformation occurred at any one time. Rather, a mosaic (*sensu* Piotrowski et al., 2004) of deforming and stable spots (characterised by ice-bed separation and basal sliding) is envisaged, enabling some sediment/landform preservation.

7. Towards a regional synthesis

In order to put our results into a more regional context (Fig.17) we briefly compare and test our deductions with some of those published recently for surrounding segments of the former BIIS. Importantly, our history of events for west central Scotland is consistent with the paradigm of a mobile, dynamic BIIS (Bowen et al., 2002; Bradwell et al., 2008; McCabe, 2008; Evans et al., 2009; Greenwood and Clark, 2009b).

Recent numerical modelling experiments (Hubbard et al., 2009) simulate initial ice advance from the northwest into the Clyde and Ayrshire basins, accompanied by independent ice cap development over the Southern Uplands. Our hypothesis for pre-cursor ribbed moraine development partially through accumulation and over-riding of ice marginal sediments also requires Southern Uplands ice to have remained a confined, independent mass while northwest-sourced ice entered the Clyde and Ayrshire basins(Fig. 14A). Further support for this early configuration comes from erratic transport paths (Fig 15) where a distinct limit of Highland-sourced erratics has been identified (Fig. 5) (Eyles et al., 1949). Coupling of the two ice masses is unlikely to have occurred until northwest-sourced ice reached at least the south-eastern fringes of the Ayrshire basin. The subsequent development of a substantial ice divide over Arran and the Firth of Clyde with eastward flow across central Scotland by the LGM (Fig. 17B) has similarities with a recent reconstruction for northern England (Evans et al., 2009) in which ice sourced over southwest Scotland and the Lake District is driven eastwards through the Stainmore and Tyne gaps. Importantly, there is no evidence for eastward transport of Arran and Ailsa Craig erratics (Fig. 15), limiting the westernmost position of the ice divide to the vicinity of Arran.

There is no direct evidence to constrain the timing of ice divide migration towards the east following the LGM (Fig. 17C). However, there is evidence of post LGM enhanced drawdown of ice towards shelf-edge fans on the continental shelf to the northwest (Bradwell et al., 2008) and into the Irish Sea basin (Eyles and McCabe, 1988; Roberts et al., 2007). Our reconstruction of westward flow over the North Channel towards Ireland at this time is consistent with the view of Salt and Evans (2004). Post-LGM, convergent westward flow may have been similar to that of an ephemeral ice stream (cf. Stokes et al., 2009) responding to break up and calving offshore to the northwest of Ireland. This interpretation is consistent with the findings of Greenwood and Clark (2009b) that once the ice sheet was established, geometry was largely controlled by fast flow / streaming corridors, which in this instance forced the ice divide to the east.

The past two decades have seen considerable advances towards understanding the dynamics and deglacial history of the Irish sector of the last BIIS, largely through the work of McCabe and co-workers (e.g. Knight and McCabe, 1997; McCabe et al, 1998, 2005, 2007a). Bedform patterns demonstrate that an ice sheet dome existed in the vicinity of Lough Neagh for much of the glacial cycle (Fig. 17B,C,D) (Knight, 2002) and a variety of inverse ice sheet models reconstruct an ice ridge over the North Channel linking the Southern Uplands and Lough Neagh dispersal centres during, and following, the LGM (e.g Boulton et al., 1991, 2002). Two major readvances interrupted decay of the Irish Ice Sheet: the Clogher Head Readvance (c. 15.0 - 14.2 ^{14}C, 18.5 - 16.7 cal ka BP), and the Killard Point Readvance (c. 14.2 - 13.0 ^{14}C, 17.1 - 15.2 cal ka BP), the latter believed to be a direct response to Heinrich Event 1 in the North Atlantic (McCabe et al., 1998; McCabe and Clark 2003; McCabe et al., 2007a). We speculate that the strong westward ice flow during stage C (Fig. 17C) may have been in operation during deposition of a moraine at Corvish, County Donegal, during the Clogher Head Readvance (McCabe et al., 2007a).

Our interpretation of subsequent westward migration of the Forth-Clyde ice divide towards Kintyre followed by topographically constrained eastward ice flow (Fig. 17D), is consistent with aspects of the reconstruction by Salt and Evans (2004) (their stages F and G). Renewed, climatically-driven ice sheet growth over northeast Ireland during the Killard Point Stadial has been suggested by McCabe et al. (1998), and is supported by recent cosmogenic exposure ages of 15.6 ^{10}Be ka BP from moraine sequences in north-western Ireland (Clark et al., 2009). It is possible that the thickening of ice over Arran and the Firth of Clyde deduced here occurred at a similar time (Fig. 17D).

It is noteworthy that the ice limits during our Stage D (and stage 3 of Paterson et al., 1998) are consistent with the ice sheet configuration in central Scotland suggested by Sissons (1963, 1964, 1967b) during the hypothesised ‘Perth Readvance’. All require the presence of a large ice mass over the Firth of Clyde during deglaciation. Evidence for a significant readvance at Perth was questioned by Francis et al. (1970), Paterson (1974), Price (1983), and Sutherland (1984), and the concept was rejected by Sissons (1976). McCabe et al. (2007b) recently cited new evidence in support of a readvance at Perth, which they correlate with the Killard Point Stadial in Ireland, concluding that it indicated an ice-sheet wide response to North Atlantic climate forcing. However, the evidence at that location remains open to interpretation (see comments by Peacock et al., 2007 and reply from McCabe et al., 2007c). The evidence presented here cannot support nor refute that a more widespread readvance of the eastern ice margin took place at this time. However, the configuration depicted (Fig. 14D, 17D) would have had the effect of isolating ice masses on the eastern side of the Clyde-Forth drainage divide from their western source, possibly leading to development of widespread ‘ice stagnation’ glaciofluvial topography, initially cited as one piece of evidence for the readvance (c.f. Sissons 1964).

Local readvances have been proposed to have occurred during deglaciation at Blackrock Ridge, at the head of Loch Indaal, Islay (Peacock and Merritt, 1997), Stranraer (Charlesworth, 1926; Peacock and Everest, 2009), and at Armoy and east Antrim on the northeastern Irish coast (McCabe, 2008) We suggest that moraine building at Blantyreferme and Kilmarnock (Fig. 5), occurred during the same overall phase of events. Ice retreat from Loch Indaal occurred possibly only a few hundred radiocarbon years before the beginning of the Lateglacial Interstadial (GI-1) (Peacock, 2008), placing tentative chronological constraints on these late ice margin oscillations. Work in progress suggests that the outer Firth of Clyde was probably deglaciated before the opening of GI-1, at c. 14.7 cal ka BP (J.D. Peacock, personal communication), with deglaciation of the Glasgow region occurring some time after.

The Irish record suggests radically different local ice sheet geometries during build-up and decay (McCabe, 2008; Greenwood and Clark, 2009b). In contrast, the ice sheet decay geometry in west central Scotland is reconstructed to have been similar to the build-up configuration (Fig. 14A, E). This was likely a result of proximity to the western Highlands, where the ice sheet was well situated to survive rises in equilibrium altitude during initial warming. Thus, the western Highlands and parts of Argyll were able to remain an important source area nourishing late stage ice margin oscillations.

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8. Summary of regional events

Following ice sheet build up (Fig. 17A), a centre of relatively immobile ice existed over Argyll and west central Scotland (Fig.14B). Ice from this centre later linked with dispersal centres over Lough Neagh, in Ireland, and over the hills of southwest Scotland and the Lake District. Ice was driven eastward towards the Firth of Forth and through the Stainmore and Tyne Gaps (Fig. 17C) (Evans et al., 2009). Ice divides then migrated both eastwards and southwards as a result of enhanced drawdown of ice towards shelf-edge fans on the continental shelf, to the northwest, and into the Irish Sea basin (Fig.17C). This reorganisation severely reduced the power of eastward flow towards the Firth of Forth and resulted in the generally accepted, relatively early deglaciation of eastern Scotland. A reversal of ice flow also occurred within the Vale of Eden and Solway Lowlands.

A major ice-surface high and ice divide developed over the outer Firth of Clyde, possibly during the Killard Point Stadial of Ireland. The ice divide probably linked with dispersal centres over Lough Neagh and the Southern Uplands (Fig. 17D). The ice sheet surface now descended from west to east over west central Scotland. On southern fringes of the Southern uplands, ice flow became topographically constrained (Salt and Evans, 2004), extending into the Solway Firth (Evans et al., 2009).

Subsequent local readvances at east Antrim, Armoy, Islay, Stranraer, Kilmarnock and Blantyreferme punctuated late stages of ice sheet decay. Whether these ice margin oscillations were synchronous and climatically driven, or diachronous and influenced by local factors such as topography and glacier bed hydrology, is uncertain. The remaining ice mass is likely to have been extremely unstable during final retreat from the Clyde basin, with large portions of the bed below the contemporary sea level.

10. Conclusions

The main conclusions from this research are as follows:

1. Published dates on preserved interstadial organic deposits show that the Main Late Devensian (MLD) (MIS 2) glaciation of central Scotland began after 35 ka cal BP. Some deposits of an earlier glaciation (MIS 4 or older) occur locally within the Clyde and Ayshire basins.

2. During a sustained build-up phase, ice advanced from the western Scottish Highlands into the Clyde and Ayrshire basins. Glaciomarine muds and shelly deposits scavenged from the Firth of Clyde were redeposited widely across Ayrshire. Ice advance against reverse slopes

enabled the build up of marginal sediment sediment accumulations. Some of these accumulations probably formed pre-cursor ridges, moulded into suites of ribbed moraine by subsequent over-riding.

3. Sustained stadial conditions at the Last Glacial Maximum (LGM) (30-25 ka cal BP) resulted in development of a major dispersal centre over the Southern Uplands and deflection of Highland ice towards the east and northeast. Relatively immobile ice beneath an ice-surface 'high' positioned over Ayrshire and the western Clyde basin, preserved previously-formed subglacial landforms and fed a wide corridor of fast-flowing ice towards the Firth of Forth.

4. A substantial re-configuration of the ice surface over west central Scotland was caused by enhanced westward drawdown into the outer Firth of Clyde and eastward migration of an ice divide towards the Clyde-Forth watershed. This reorganisation is tentatively correlated with the Clogher Head Readvance established in the north of Ireland (c. 15.0 - 14.2 ^{14}C, 18.5 - 16.7 cal ka BP).

5. Renewed ice sheet thickening over the Firth of Clyde may have accompanied growth of the Irish Ice sheet during the Killard Point Stadial (c. 14.2 - 13.0 ^{14}C, 17.1 - 15.2 cal ka BP). Subsequent ice sheet retreat was initially characterised by substantial meltwater production, ponding and erosion.

6. One or more significant ice front oscillations occurred late during deglaciation. These were nourished by elevated source areas in the western Highlands and Argyll, which were well placed to survive initial warming. The discovery of De Geer moraines in western Ayrshire allows ice margin velocity during one such oscillation to be calculated as ≤ 290 m yr $^{-1}$. These late oscillations probably occurred close to the opening of the Lateglacial Interstadial (GI-1).

7. Once the MLD ice sheet margin had retreated into the inner Firth of Clyde, it was extremely vulnerable to collapse, which may have occurred early in GI-1. It was accompanied by marine incursion of the lower Clyde Valley up to c. 40 m above present-day sea level.

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Figure Captions

Figure 1. Regional context. Box (labelled Figs. 2, 5, 14) shows location of study area in west central Scotland. Proposed former glacier ice limits and ages from Sissons (1967b), Dawson (1982), Rose et al. (1988), Peacock and Merritt (1997), McCabe et al. (1998, 2003), Thomas et al. (2004), Ballantyne (2007), and McCabe (2008). Key place names are shown.

Figure 2. Topography of study area and place names mentioned in text. Locations of subsequent figures are shown. Hill-shaded digital surface model built from Intermap Technologies NEXTMap® Britain topographic data. Northwest illumination.

Figure 3. Simplified lithostratigraphy for the period spanning the Main Late Devensian glaciation for the Clyde and Ayrshire basins. Based on McMillan et al. (2005, in press). For clarity only formations of primary relevance to this study are included. Fm - Formation, Mbr - Member. GICC05 - Greenland Ice Core Chronology 2005 events, after Lowe et al. (2008).

Figure 4. Three-dimensional Quaternary geological models revealing basic composition of geomorphological features in Glasgow area. A. Fence diagram revealing three-dimensional geology of Erskine - Renfrew area. Note drumlins entirely comprise sediments of the Wilderness Till Formation (in blue). Cross sections are based on borehole records. Vertical sticks represent individual boreholes B. Complete three-dimensional geological model for central Glasgow area showing Paisley Clay Mbr draped over drumlins comprising Wilderness Till Fm. C. Three dimensional geological model of Paisley area. Paisley Clay Formation sediments (in green) and alluvial sediments (in yellow) are removed to more clearly reveal bedforms in the Wilderness Till Formation (blue). All images are vertically exaggerated between 5 and 10 times.

Figure 5. Glacial geomorphology and geology of the Clyde and Ayrshire basins. Erratic limits from Eyles et al. (1949) and glacial striations from Paterson et al. (1998).

Figure 6. Temporary section within zone of drumlinised ribbed moraine at Holmbrae Road, Glasgow. Ice flow direction inferred from regional streamlining and sense of compression.

Figure 7. Streamlined bedforms superimposed on ribbed moraine. See text for description. Hill-shaded digital surface models built from Intermap Technologies NEXTMap® Britain topographic data. Northwest illumination.

Figure 8. Streamlined bedforms and flowsets identified in this study. fs - flowset.

Figure 9. Cross cutting of streamlined bedforms in the Ayshire basin to the west of Kilmarnock.

Figure 10. A. Ice-marginal meltwater channels clearly descending towards the NE, to the south of Strathaven. B. Assemblage of moraine ridges and NE declining marginal meltwater channels in the vicinity of Greenock Mains. White arrows denote inferred final palaeo-ice flow direction. Hill-shaded digital surface models built from Intermap Technologies NEXTMap® Britain topographic data. Northwest illumination.

Figure 11. Deposits associated with ice-dammed lakes in the Clyde basin. Deltaic sediments of the Ross Sand Mbr revealed in former sand and gravel pits near Bishopbriggs (A) and Hamilton (B). Photos from BGS archive image base. C. Lithostratigraphy including surface and buried glaciolacustrine deposits, revealed in the BGS Bellshill borehole.

Figure 12. A. Hill-shaded digital surface model, built from Intermap Technologies NEXTMap® Britian topographic data, revealing narrow transverse ridges in the vicinity of Kilmarnock. Note overprinting of ridges on streamlined bedforms. B. Interpretation of same area.

Figure 13. Cross-profile data extracted from the digital surface model revealing dimensions of the narrow transverse ridges.

Figure 14. Reconstructed stages, showing the evolution of the last BIIS in west central Scotland. See text for discussion. Hill-shaded digital surface models built from Intermap Technologies NEXTMap® Britain topographic data. Northwest illumination.

Figure 15. Erratic transport paths in SW Scotland. From Eyles et al. (1949) and Sissons (1967a). Note, erratic paths do not imply contemporaneous flow. Rectangle delimits main study area as shown in Figs. 2 and 5.

Figure 16. Section at Nith Bridge, from Holden and Jardine, (1980).

Figure 17. Proposed evolution of the western sector of the last BIIS. Arrows denote palaeo-ice flow directions. Dashed lines indicate approximate positions of ice divides. See text for discussion.

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Table 1. Morphological characteristics of streamlined bedforms in study area. ER - elongation ratio, St Dev - standard deviation.

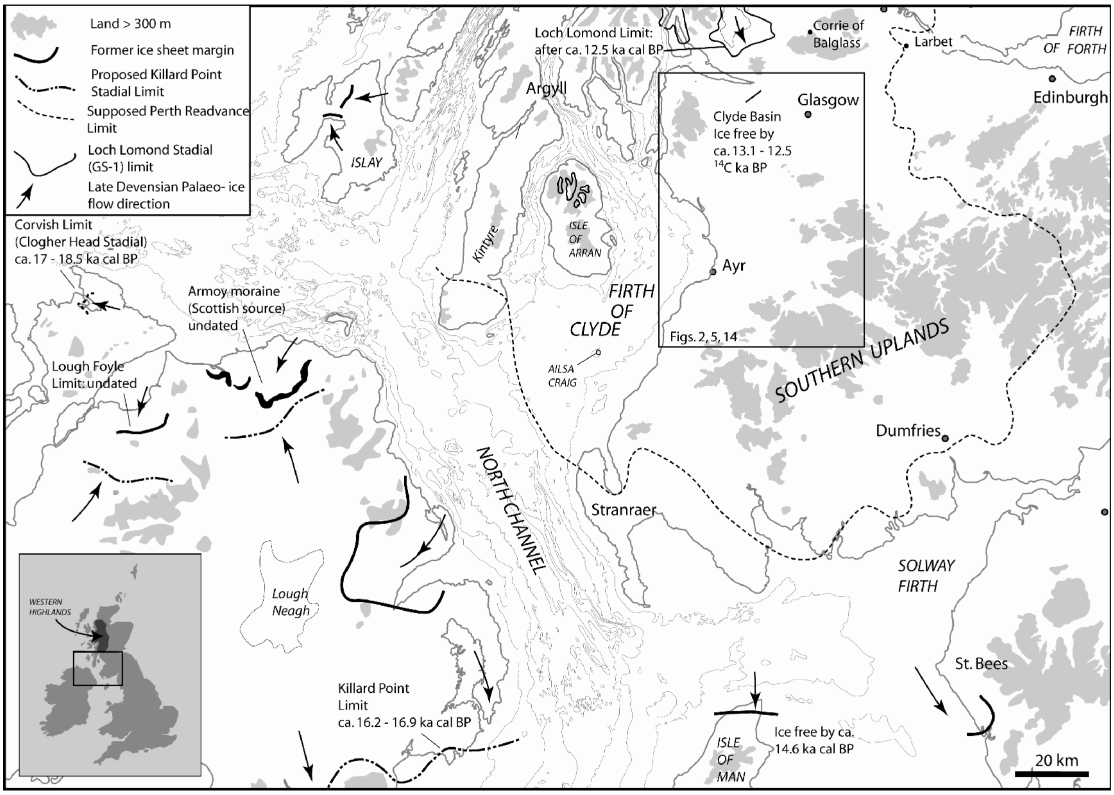
1300 

Fig. 1

1305

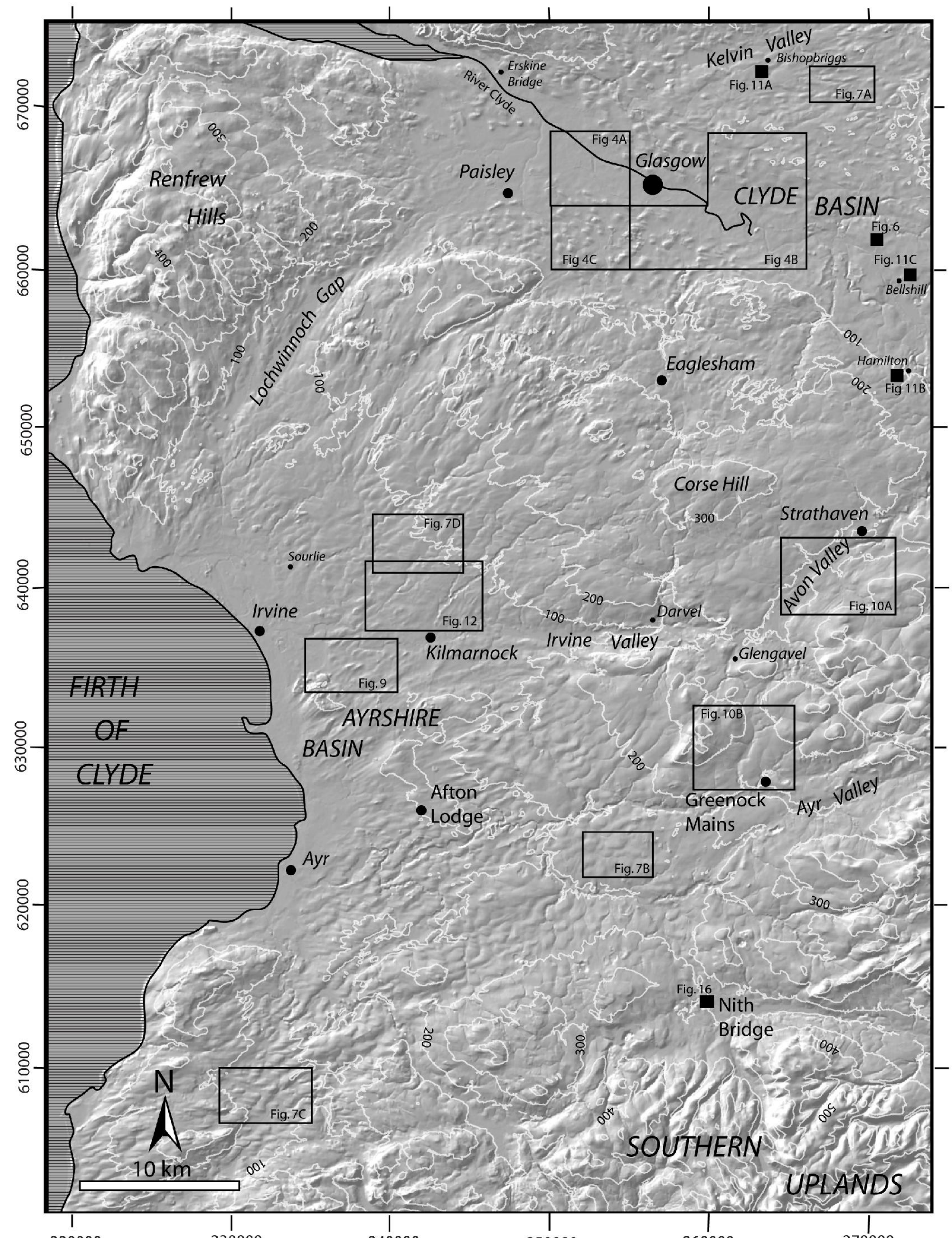

Kelvin Valley
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Fig. 11A
Fig. 7A
Fig 4A
Glasgow
Paisley
Renfrew Hills
CLYDE BASIN
Fig. 6
Fig 4C
Fig 4B
Fig. 11C
Bellshill
Lochwinnoch Gap
Eaglesham
Hamilton
Fig 11B
Corse Hill
Strathaven
Fig. 7D
Sourlie
Avon Valley
Fig. 10A
Irvine
Fig. 12
Darvel
Kilmarnock
Irvine Valley
Glengavel
FIRTH OF CLYDE
Fig. 9
AYRSHIRE BASIN
Fig. 10B
Afton Lodge
Greenock Mains
Ayr Valley
Ayr
Fig. 7B
Fig. 16
Nith Bridge
N
Fig. 7C
10 km
SOUTHERN UPLANDS
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Fig. 2

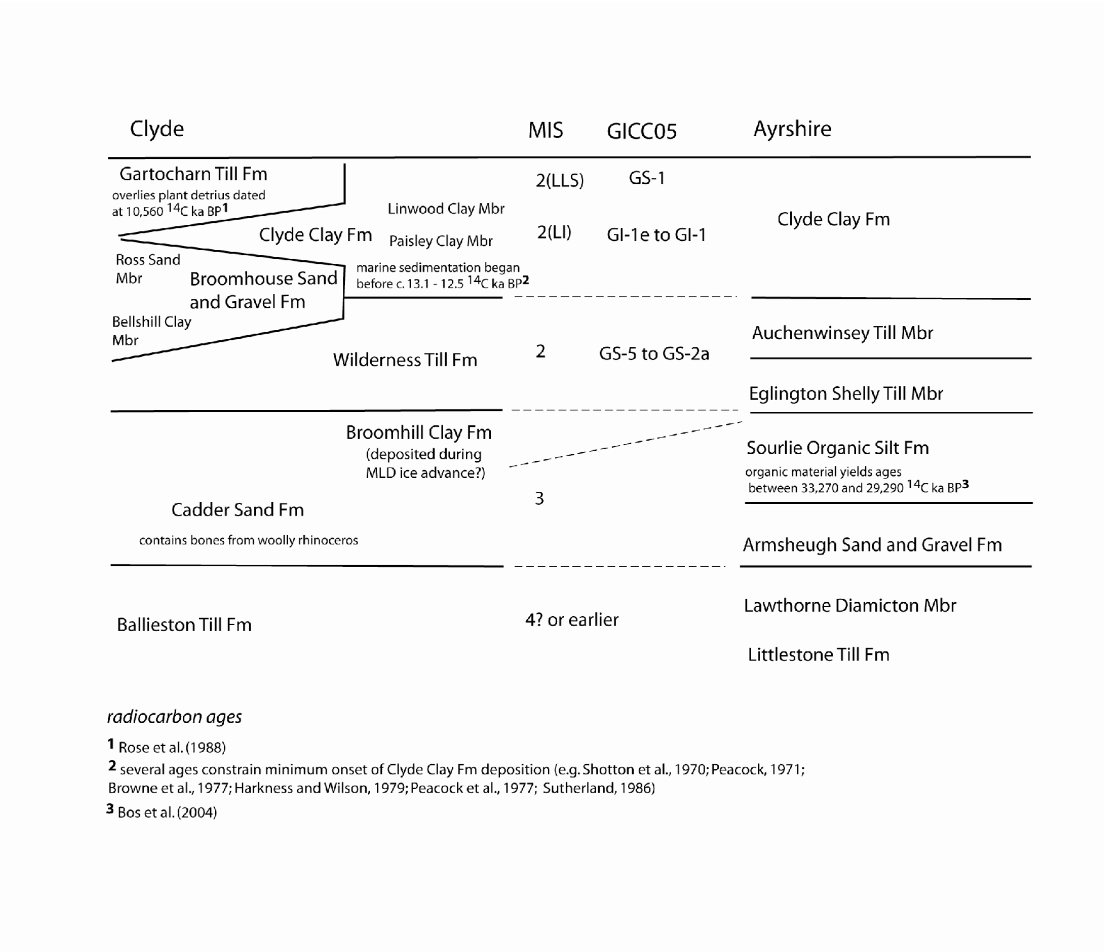
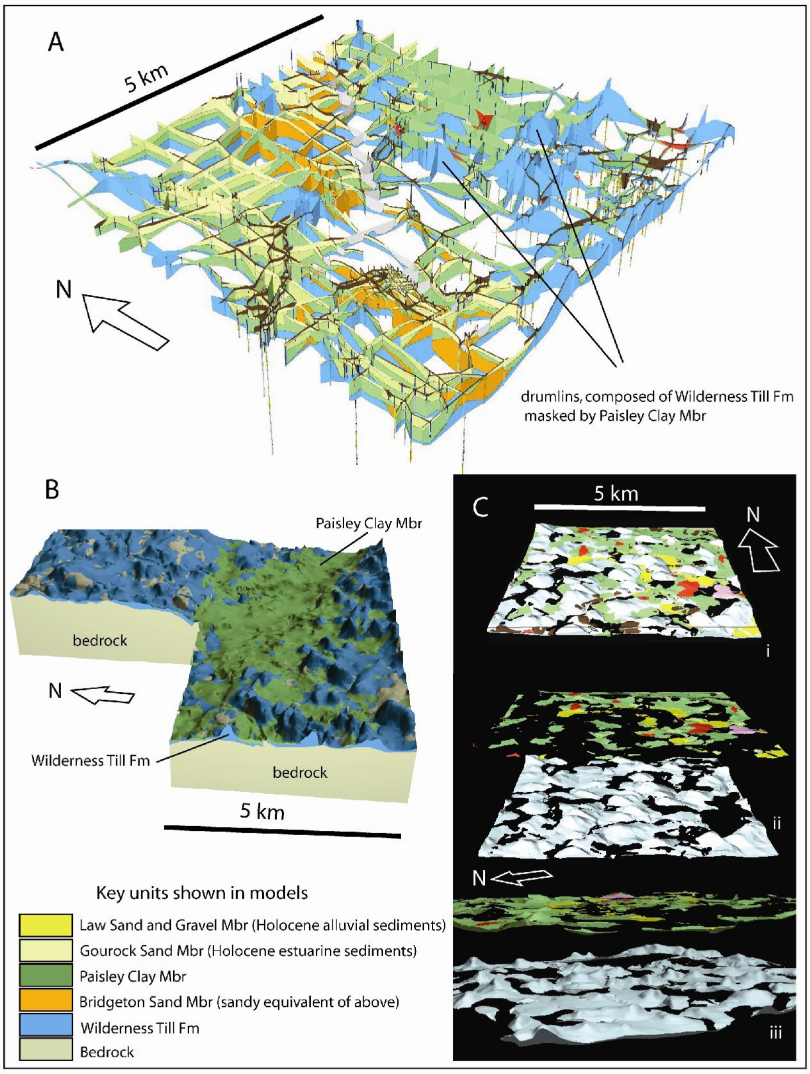

Clyde
MIS
GICC05
Ayrshire
Gartocharn Till Fm
overlies plant detrius dated at 10,560 14C ka BP1
2(LLS)
GS-1
Linwood Clay Mbr
Clyde Clay Fm
Clyde Clay Fm
Paisley Clay Mbr
2(LI)
GI-1e to GI-1
Ross Sand Mbr
Broomhouse Sand and Gravel Fm
marine sedimentation began before c. 13.1 - 12.5 14C ka BP2
Bellshill Clay Mbr
Auchenwinsey Till Mbr
Wilderness Till Fm
2
GS-5 to GS-2a
Eglington Shelly Till Mbr
Broomhill Clay Fm
(deposited during MLD ice advance?)
Sourlie Organic Silt Fm
organic material yields ages between 33,270 and 29,290 14C ka BP3
3
Cadder Sand Fm
contains bones from woolly rhinoceros
Armsheugh Sand and Gravel Fm
Ballieston Till Fm
4? or earlier
Lawthorne Diamicton Mbr
Littlestone Till Fm
radiocarbon ages
1 Rose et al. (1988)
2 several ages constrain minimum onset of Clyde Clay Fm deposition (e.g. Shotton et al., 1970; Peacock, 1971; Browne et al., 1977; Harkness and Wilson, 1979; Peacock et al., 1977; Sutherland, 1986)
3 Bos et al. (2004)

Fig. 3


A
5 km
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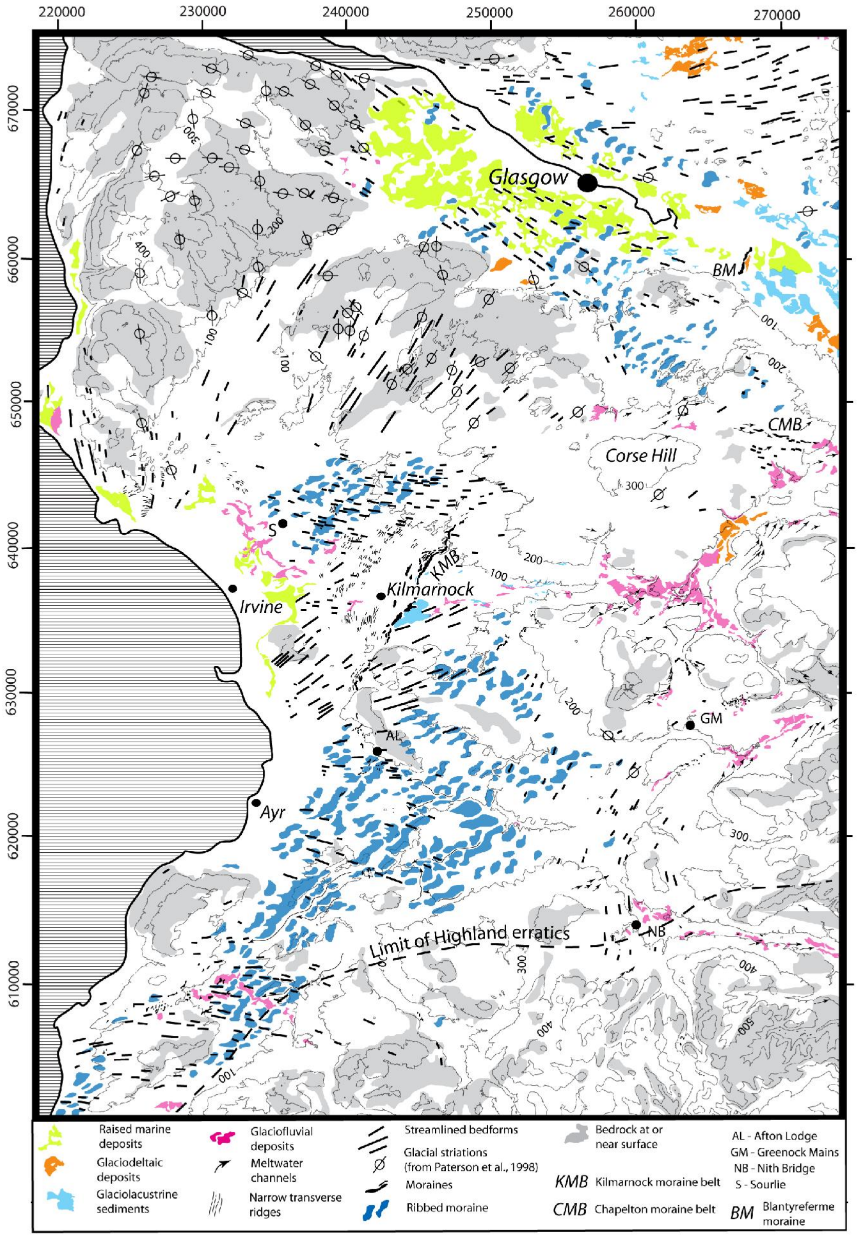
drumlins, composed of Wilderness Till Fm masked by Paisley Clay Mbr
B
Paisley Clay Mbr
bedrock
N
Wilderness Till Fm
bedrock
5 km
C
5 km
N
i
ii
N
iii
Key units shown in models
Law Sand and Gravel Mbr (Holocene alluvial sediments)
Gourock Sand Mbr (Holocene estuarine sediments)
Paisley Clay Mbr
Bridgeton Sand Mbr (sandy equivalent of above)
Wilderness Till Fm
Bedrock

Fig. 4

1315

35



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CMB
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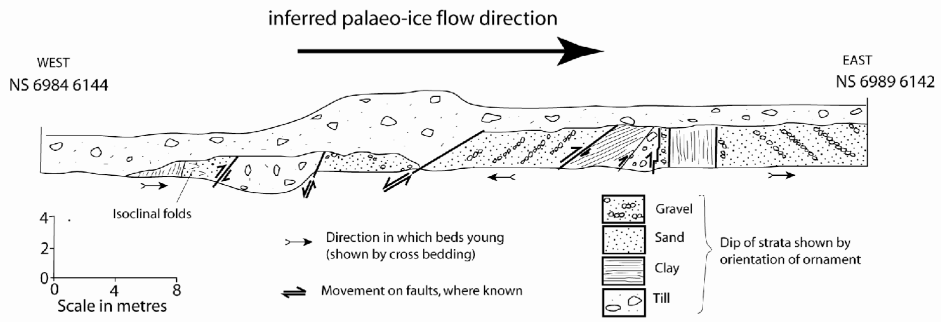
KMB
Kilmarnock
Irvine
S
AL
GM
Ayr
NB
Limit of Highland erratics
Raised marine deposits
Glaciodeltaic deposits
Glaciolacustrine sediments
Glaciofluvial deposits
Meltwater channels
Narrow transverse ridges
Streamlined bedforms
Glacial striations (from Paterson et al., 1998)
Moraines
Ribbed moraine
Bedrock at or near surface
KMB Kilmarnock moraine belt
CMB Chapelton moraine belt
AL - Afton Lodge
GM - Greenock Mains
NB - Nith Bridge
S - Sourlie
BM Blantyreferme moraine

Fig. 5


inferred palaeo-ice flow direction
WEST
NS 6984 6144
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NS 6989 6142
Isoclinal folds
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2
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Scale in metres
Direction in which beds young (shown by cross bedding)
Movement on faults, where known
Gravel
Sand
Clay
Till
Dip of strata shown by orientation of ornament

Fig. 6

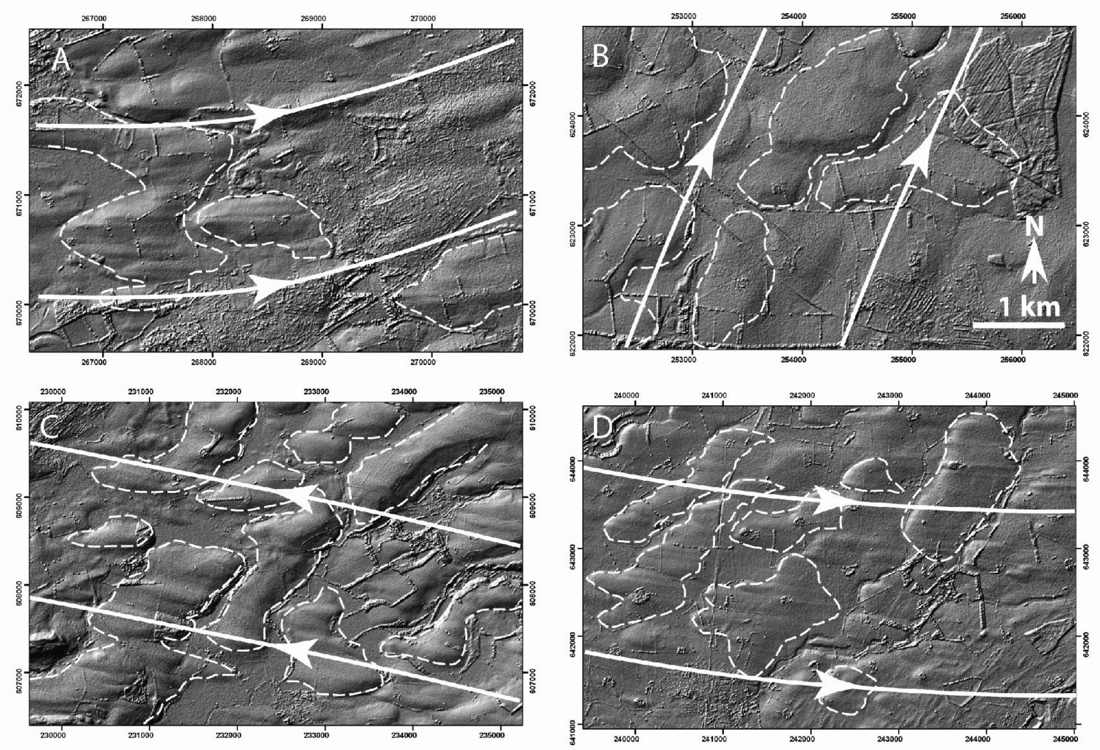
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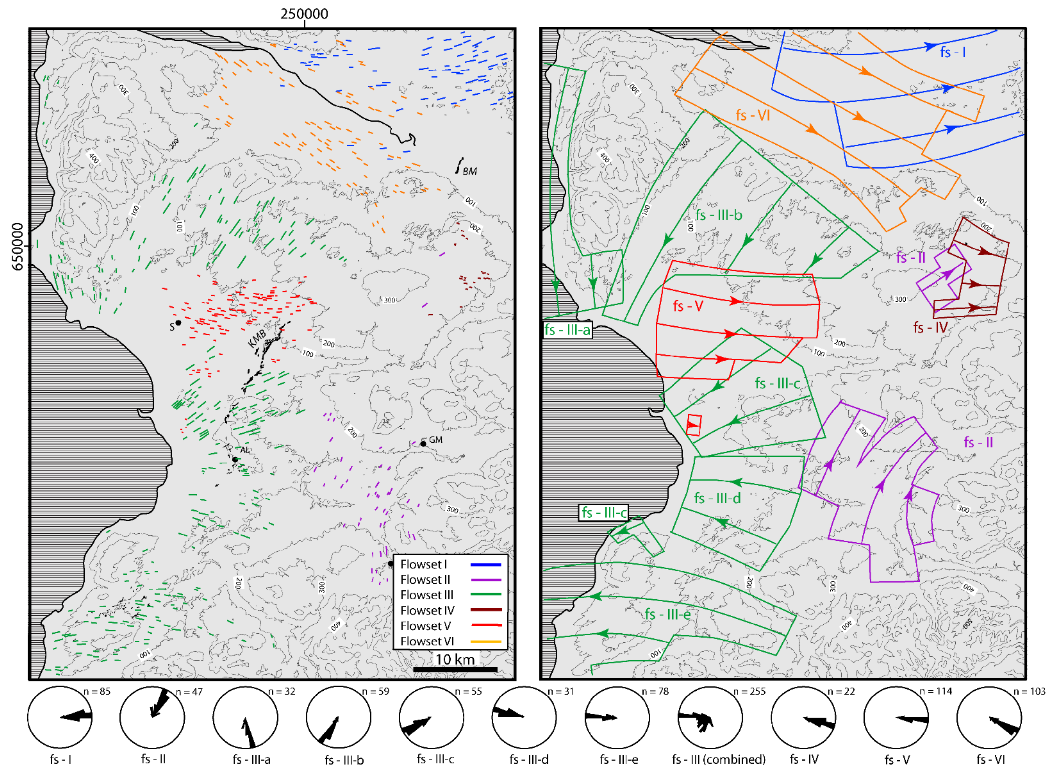
Fig. 7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fs - III-b
fs - II
fs - V
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fs - III-a
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fs - III-c
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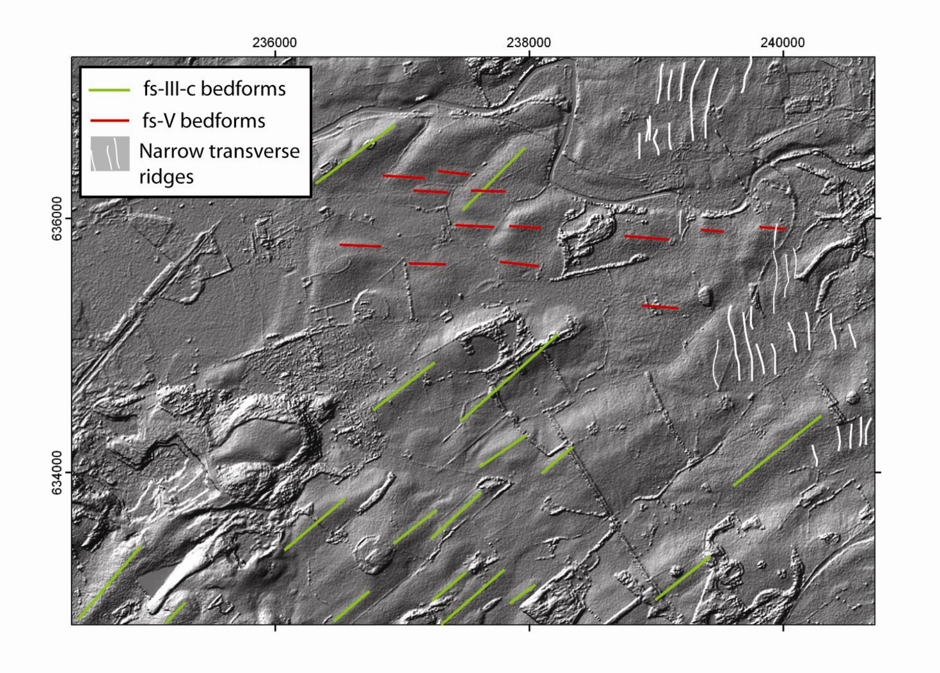
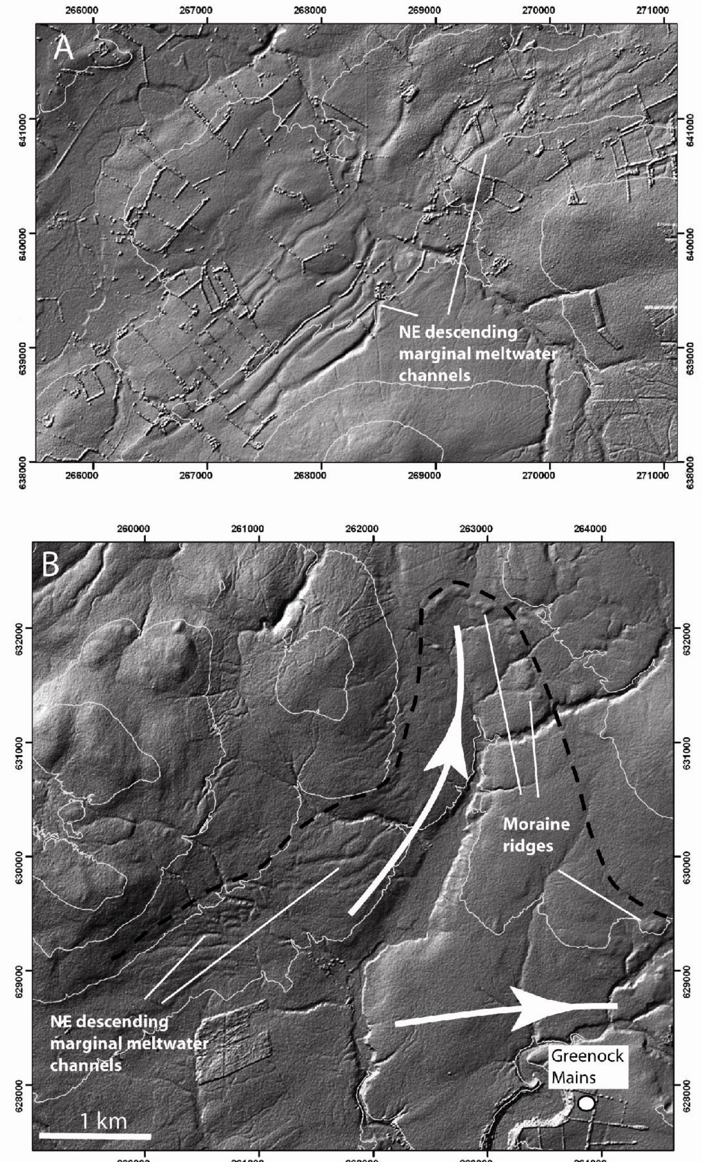
n = 85
n = 47
n = 32
n = 59
n = 55
n = 31
n = 78
n = 255
n = 22
n = 114
n = 103
fs - I
fs - II
fs - III-a
fs - III-b
fs - III-c
fs - III-d
fs - III-e
fs - III (combined)
fs - IV
fs - V
fs - VI

Fig. 8


236000
238000
240000
636000
634000
fs-III-c bedforms
fs-V bedforms
Narrow transverse ridges

1340
Fig. 9

40



1345 Fig. 10

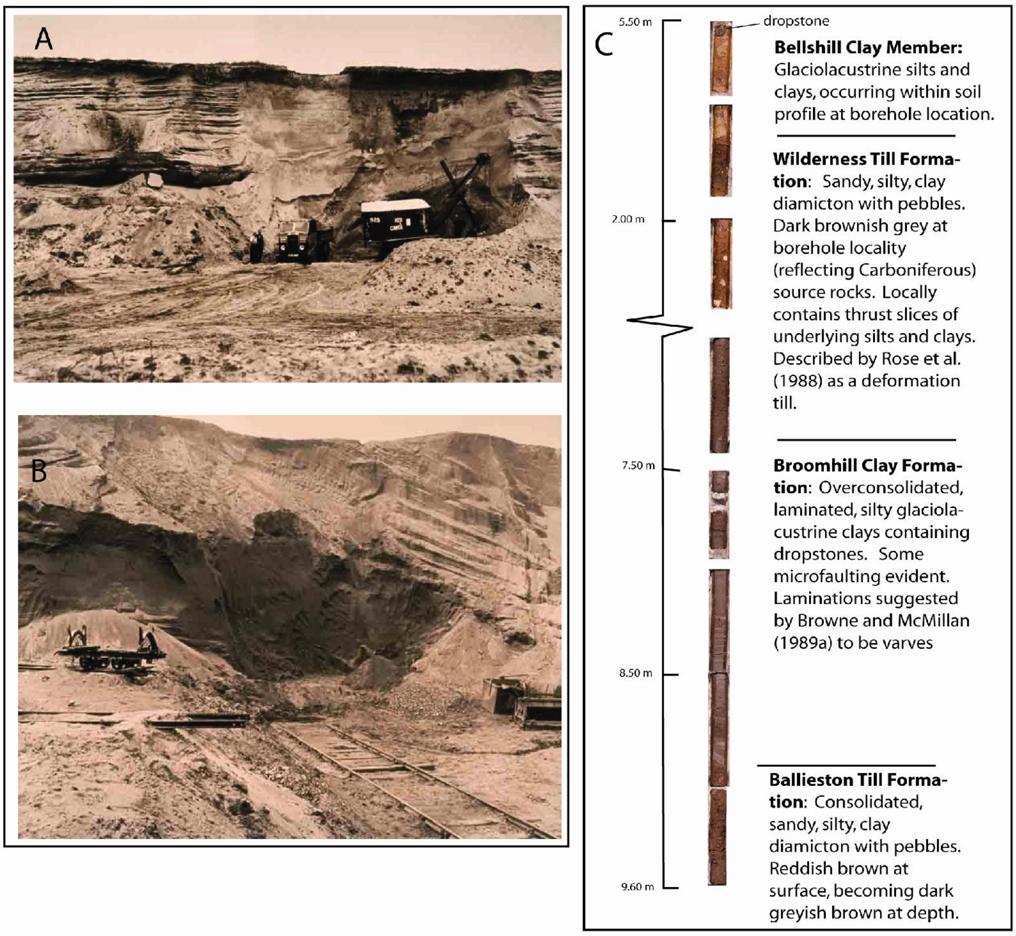

A
B
C
5.50 m
dropstone
Bellshill Clay Member: Glaciolacustrine silts and clays, occurring within soil profile at borehole location.
Wilderness Till Formation: Sandy, silty, clay diamicton with pebbles. Dark brownish grey at borehole locality (reflecting Carboniferous) source rocks. Locally contains thrust slices of underlying silts and clays. Described by Rose et al. (1988) as a deformation till.
2.00 m
7.50 m
Broomhill Clay Formation: Overconsolidated, laminated, silty glaciolacustrine clays containing dropstones. Some microfaulting evident. Laminations suggested by Browne and McMillan (1989a) to be varves
8.50 m
Ballieston Till Formation: Consolidated, sandy, silty, clay diamicton with pebbles. Reddish brown at surface, becoming dark greyish brown at depth.
9.60 m

Fig. 11

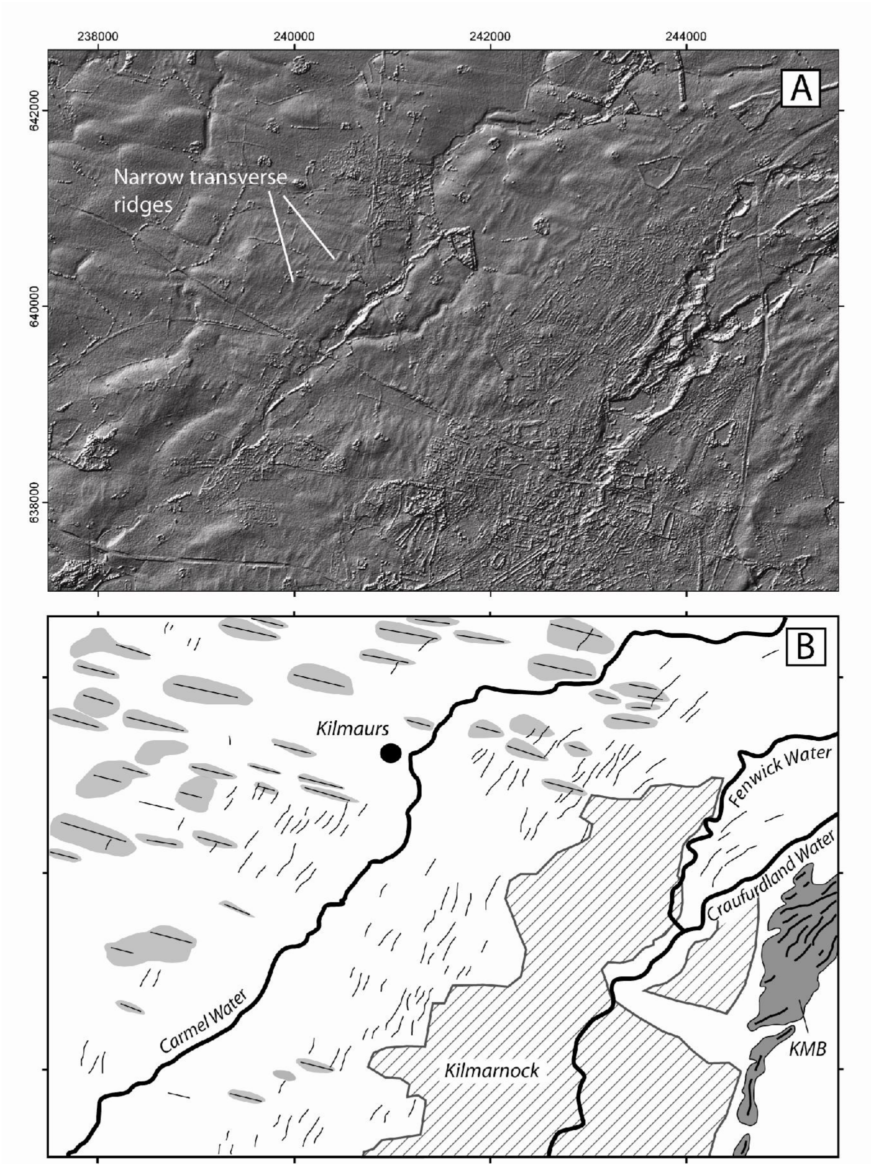

238000
240000
242000
244000
642000
640000
638000
A
Narrow transverse ridges
B
Kilmaurs
Fenwick Water
Craufurdland Water
Carmel Water
Kilmarnock
KMB

Fig. 12

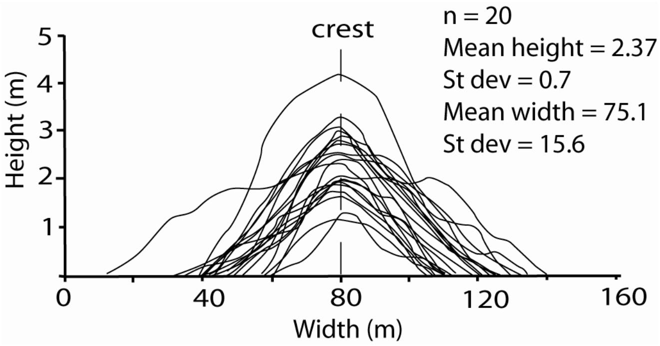

crest
n = 20
Mean height = 2.37
St dev = 0.7
Mean width = 75.1
St dev = 15.6
Height (m)
5
4
3
2
1
0
40
80
120
160
Width (m)

Fig. 13

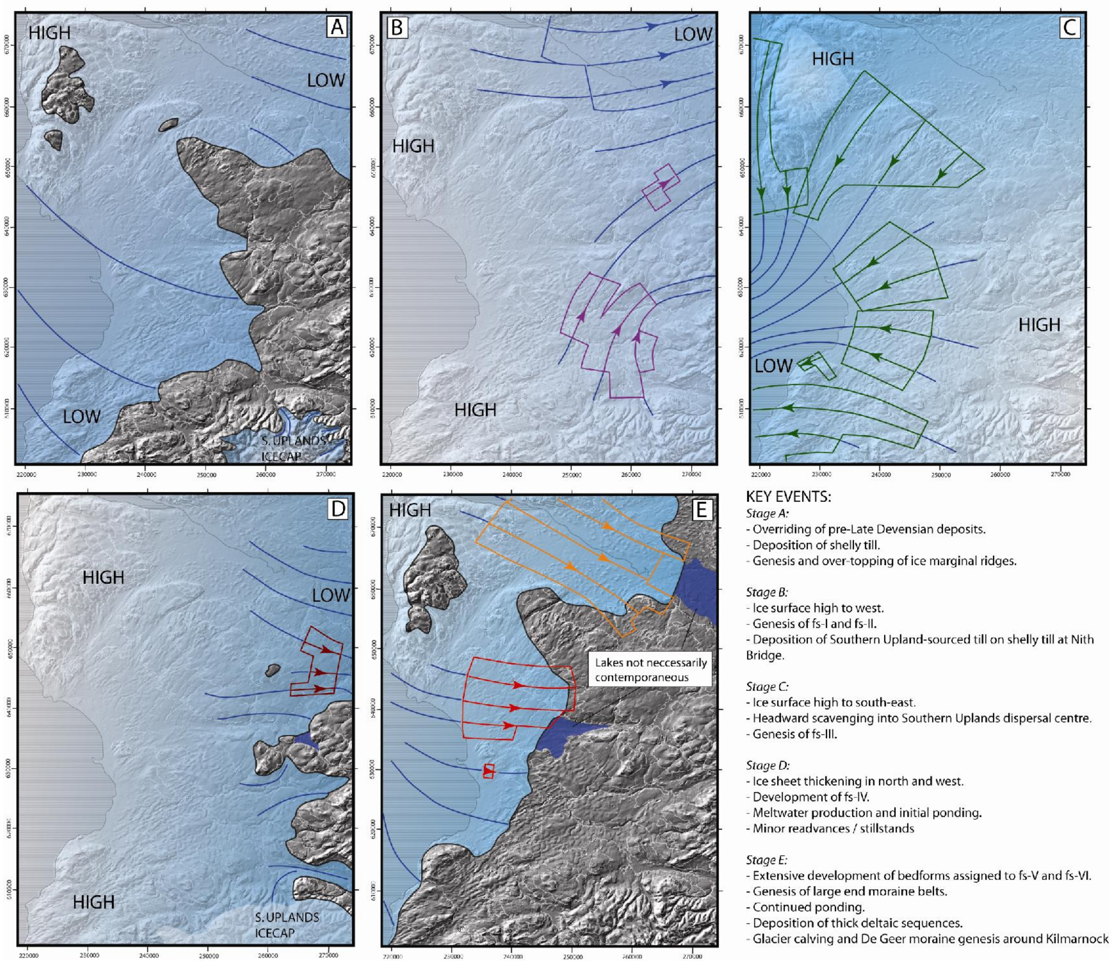
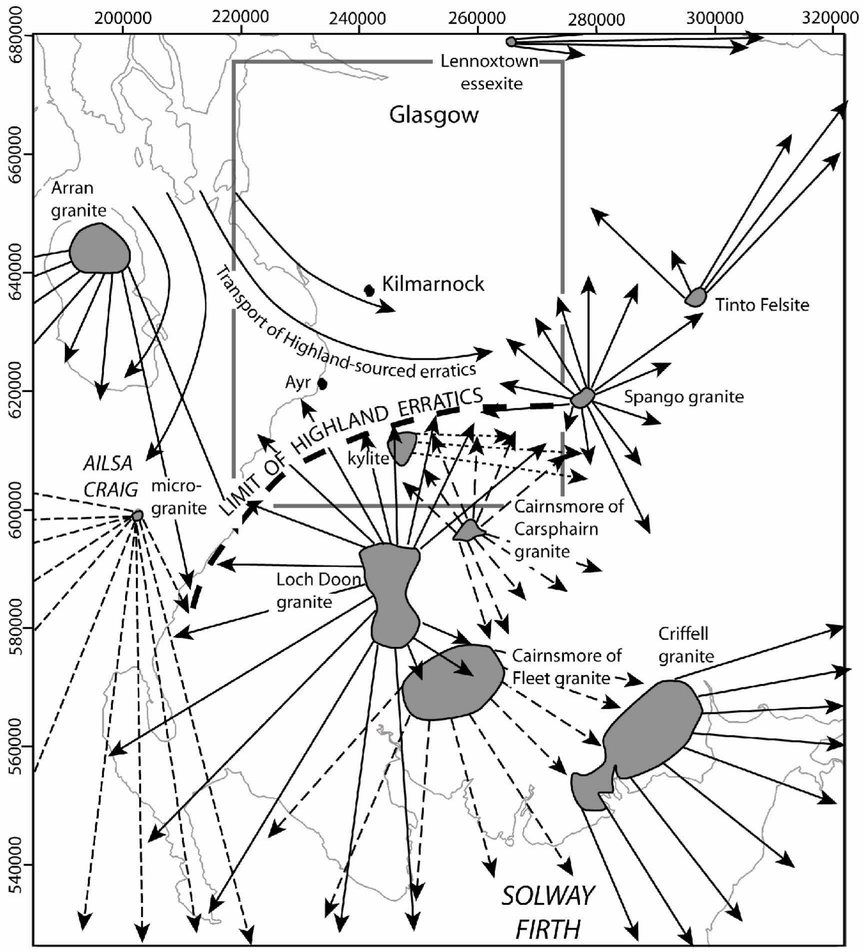
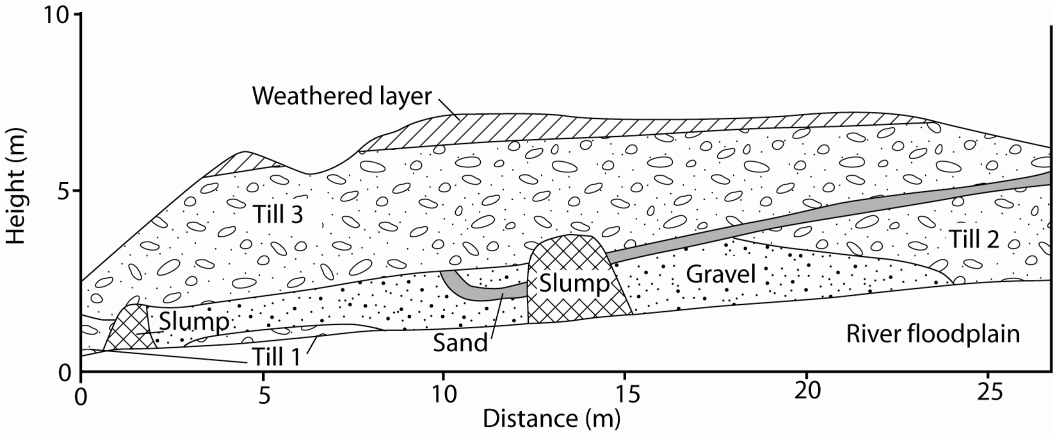

A
HIGH
LOW
LOW
S. UPLANDS ICECAP
B
LOW
HIGH
HIGH
C
HIGH
HIGH
LOW
D
HIGH
LOW
HIGH
S. UPLANDS ICECAP
E
HIGH
Lakes not neccessarily contemporaneous
KEY EVENTS:
Stage A:
- Overriding of pre-Late Devensian deposits.
- Deposition of shelly till.
- Genesis and over-topping of ice marginal ridges.
Stage B:
- Ice surface high to west.
- Genesis of fs-I and fs-II.
- Deposition of Southern Upland-sourced till on shelly till at Nith Bridge.
Stage C:
- Ice surface high to south-east.
- Headward scavenging into Southern Uplands dispersal centre.
- Genesis of fs-III.
Stage D:
- Ice sheet thickening in north and west.
- Development of fs-IV.
- Meltwater production and initial ponding.
- Minor readvances / stillstands
Stage E:
- Extensive development of bedforms assigned to fs-V and fs-VI.
- Genesis of large end moraine belts.
- Continued ponding.
- Deposition of thick deltaic sequences.
- Glacier calving and De Geer moraine genesis around Kilmarnock

Fig. 14

45



1385 Fig. 15



	Till 1	Till 2	Till 3
Mean clast orientation	193	175	351
Percentage gravel	15.07	23.21	46.52
Percentage sand	39.89	34.54	24.28
Percentage silt and clay	45.23	42.25	29.20
Percentage of local erratics	88	100	24
Percentage of S. Uplands erratics	0	0	76
Percentage of Highland erratics	6	0	0
Shell fragments	present	present	absent

Fig. 16

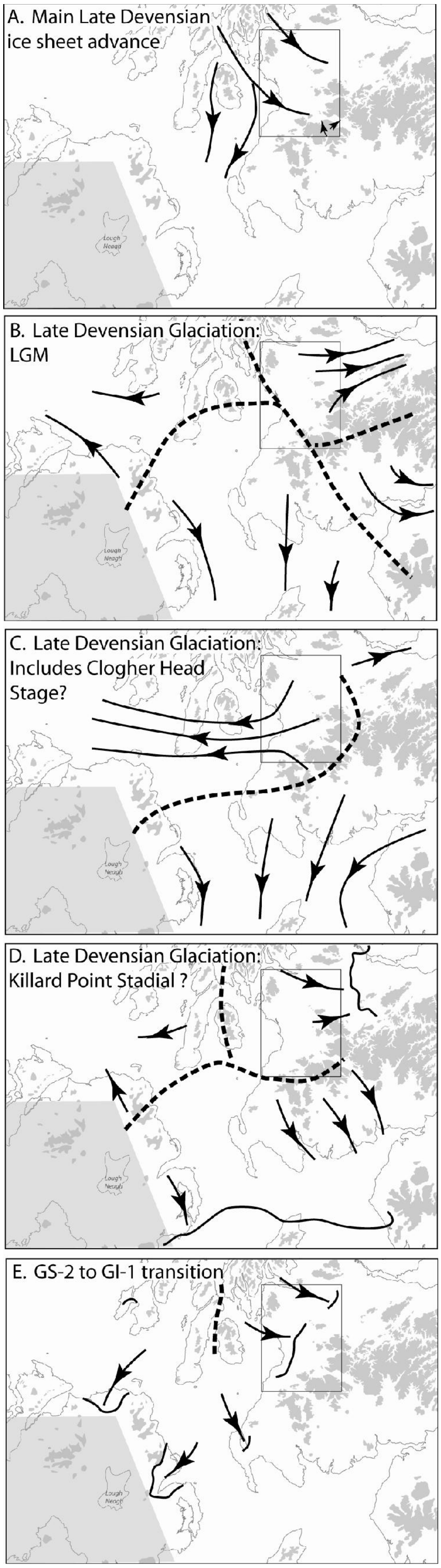

A. Main Late Devensian ice sheet advance
B. Late Devensian Glaciation: LGM
C. Late Devensian Glaciation: Includes Clogher Head Stage?
D. Late Devensian Glaciation: Killard Point Stadial ?
E. GS-2 to GI-1 transition

Fig. 17

Flowset	Length			Width			ER		
	Range	Mean	St Dev	Range	Mean	St Dev	Range	Mean	St Dev
Fs - 1	370 - 1498	826	262	75 - 468	239	87	2.1 - 7.6	3.7	1.1
Fs - II	330 - 1195	563	187	65 - 443	307	83	1.6 - 7.7	2.9	1.0
Fs - III-a	350 - 2430	706	476	76 - 500	194	103	2.0 - 5.3	3.7	1.0
Fs - III-b	379 - 2391	1017	445	79 - 590	274	108	2.3 - 6.9	3.8	0.9
Fs - III - c	400 - 2600	1063	367	100 - 467	236	78	2.5 - 8.0	4.6	1.3
Fs - III-d	449 - 1214	766	226	118 - 469	250	88	1.7 - 5.7	3.3	0.9
Fs - III-e	372 - 1700	754	244	72 - 580	197	84	2.2 - 11.6	4.3	1.5
Fs - III (combined)	350 - 2600	877	384	72 - 590	230	96	1.7 - 11.6	4	1.3
Fs - IV	204 - 487	321	86	51 - 134	76	22	3.1 - 5.8	4.3	0.9
Fs - V	167 - 1142	540	179	61 - 290	127	39	2.3 - 8.0	4.3	1.1
Fs - VI	290 - 1206	660	155	80- 467	232	79	1.7 - 6.4	3.1	0.8

Table 1