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Evans, D.J.A., Clark, C.D. and Mitchell, W.A. (2005) The last British Ice Sheet: A review of the evidence utilised in the compilation of the Glacial Map of Britain. Earth-Science Reviews, 70 (3-4). pp. 253-312. ISSN 0012-8252

https://doi.org/10.1016/j.earscirev.2005.01.001

The last British Ice Sheet: A review of the evidence utilised in the compilation of the Glacial Map of Britain

David J.A. Evans [a,] *, Chris D. Clark [b], Wishart A. Mitchell [a]

[a] *Department of Geography, University of Durham, South Road, Durham, DH13LE, England, UK*

[b] *Department of Geography, University of Sheffield, Sheffield, S10 2TN, England, UK*

* *Corresponding author. Present address: Department of Geography, University of Durham, South Road, Durham, DL1 3LE, England, UK. E-mail address: d.j.a.evans@durham.ac.uk*

Abstract

This paper reviews the evidence presently available (as at December 2003) for the compilation of the Glacial Map of Britain (see [Clark C.D., Evans D.J.A., Khatwa A., Bradwell T., Jordan C.J., Marsh S.H., Mitchell W.A., Bateman, M.D., 2004. Map and GIS database of glacial landforms and features related to the last British Ice Sheet. Boreas 33, 359–375] and http:// www.shef.ac.uk/geography/staff/clark_chris/britice.html) in an effort to stimulate further research on the last British Ice Sheet and promote a reconstruction of ice sheet behaviour based on glacial geology and geomorphology. The wide range of evidence that has been scrutinized for inclusion on the glacial map is assessed with respect to the variability of its quality and quantity and the existing controversies in ice sheet reconstructions. Landforms interpreted as being of unequivocal ice-marginal origin (moraines, ice-contact glacifluvial landforms and lateral meltwater channels) and till sheet margins are used in conjunction with available chronological control to locate former glacier and ice-sheet margins throughout the last glacial cycle. Subglacial landforms (drumlins, flutings and eskers) have been used to demarcate former flow patterns within the ice sheet. The compilation of evidence in a regional map is crucial to any future reconstructions of palaeo-ice sheet dynamics and will provide a clearer understanding of ice sheet configuration, ice divide migration and ice thickness and coverage for the British Ice Sheet as it evolved through the last glacial cycle.

1. Introduction

It is apparent that the quality and the density of detailed research into the glacial geomorphology of Britain vary significantly, and before reconstructions of the former British Ice Sheet begin to approach the accuracy of those produced in other regions (e.g. Nordkalott Project, 1986/87; Dyke and Prest, 1987; Boulton and Clark, 1990 a,b; Kleman and Borgstrom, 1996; Clark, 1997) this variability needs to be addressed. Reconstructions of the British Ice Sheet during the last glacial cycle have evolved over a long period. Regional assessments, because they rarely have been attempted more than once, are characterized by the predominant methodologies at the time of study. Consequently, past reconstructions of the ice sheet have been compiled from a variety of types of evidence, much of which was extracted from the existing literature (e.g. Boulton et al., 1977, 1985). Additionally, ice sheet reconstructions in other regions such as Fennoscandinavia and North America have identified a number of stillstand or readvance phases, based upon prominent moraine systems, that have then been used in the construction of morphostratigraphies. This approach has been validated most recently by the identification of evidence for significant regional scale palaeoclimatic changes and concomitant palaeo-ice sheet oscillations in ice core and ocean core records. In contrast, major moraine systems in the British Isles remain under-utilized as palaeoclimatic indicators. We review here for the first time the nature of the glacial geomorphological evidence for the former British Ice Sheet, in order to highlight the gaps in our knowledge and the need for a systematic mapping programme for the British Isles as a glaciated region.

We have produced three outputs:

(a) a glacial map of Britain (see Clark et al., 2004);
(b) a fully referenced GIS database of glacial features (see Clark et al., 2004 and website http://www.shef.ac.uk/geography/staff/clark_chris/ britice.html);
(c) a review of the available evidence as presented in this paper.

A smaller version of the Glacial Map of Britain is reproduced here as Fig. 1a. This paper reports on the range of evidence that has been scrutinized for inclusion on the glacial map and highlights the variability of its quality and quantity and existing controversies in ice sheet reconstructions. Hopefully this synthesis will stimulate research in areas that hitherto have received little attention, thus filling gaps in our knowledge and perhaps resurrecting useful evidence that has become “lost” in the expansive literature. In the longer term we hope that this review will stimulate further research on the evidence for the British Ice Sheet and allow a reconstruction of ice sheet behaviour based on glacial geology and geomorphology.

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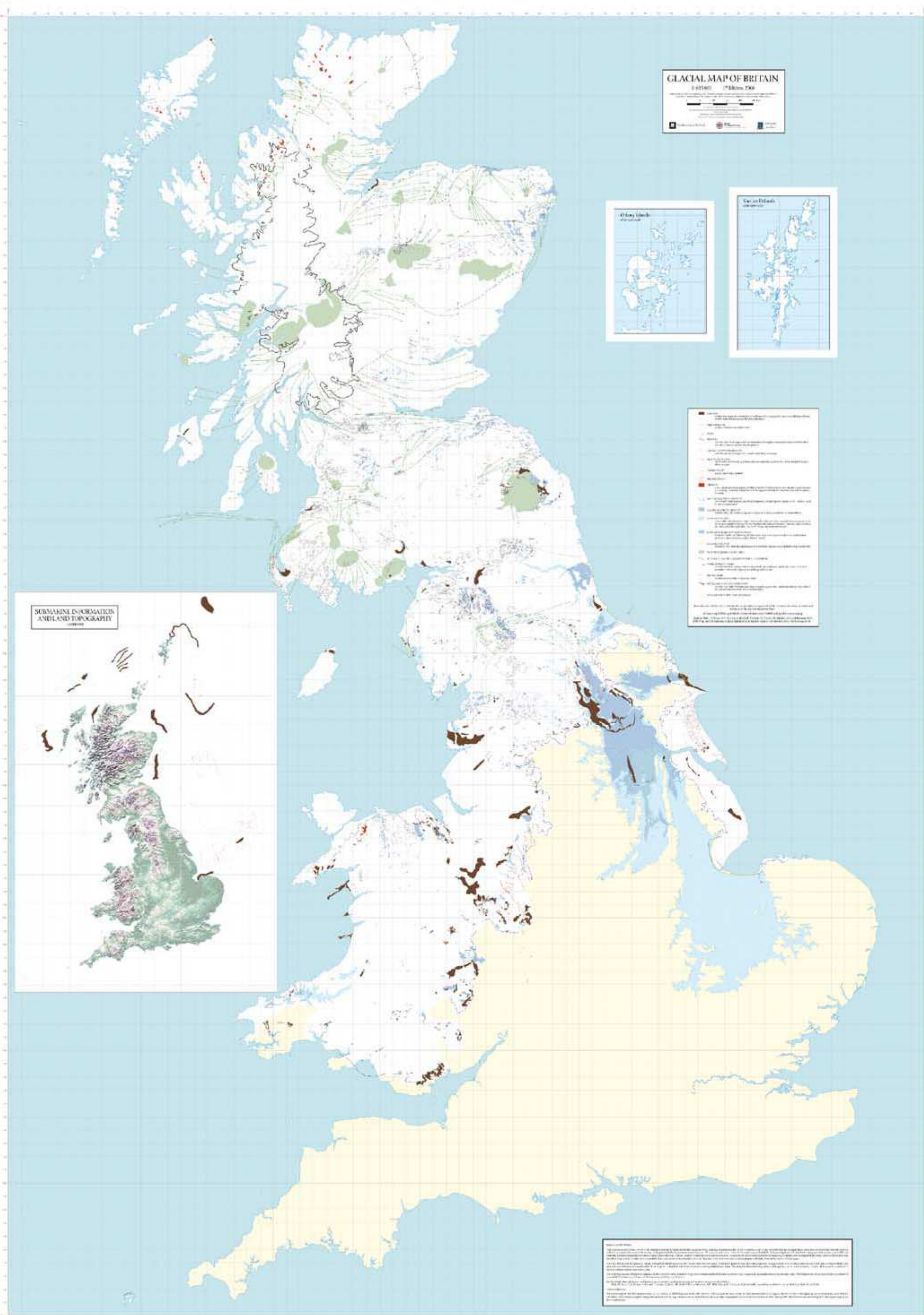

Fig. 1. (a) Reproduction of the Glacial Map of Britain (Clark et al. ,2004) ,which gives an overview of the twenty thematic layers displayed within the BRITICE GIS including terrestrial and offshore features. The database comprises over 20,000 individual features divided into layer sof moraines, eskers, drumlins meltwater channels ,tunnel valleys, shelf-edge fans, trimlines, key drift limits, glaciolacustrine deposit , ice-dammed lakes, erratic dispersal and the Loch Lomond Readvance limit as it pertains to the West Highland glacier complex. A larger format version of this map is available in Clark et al. (2004) and the GIS data base is available at the website http://www.shef.ac.uk/geography/staff/clark_chris/britice.html.

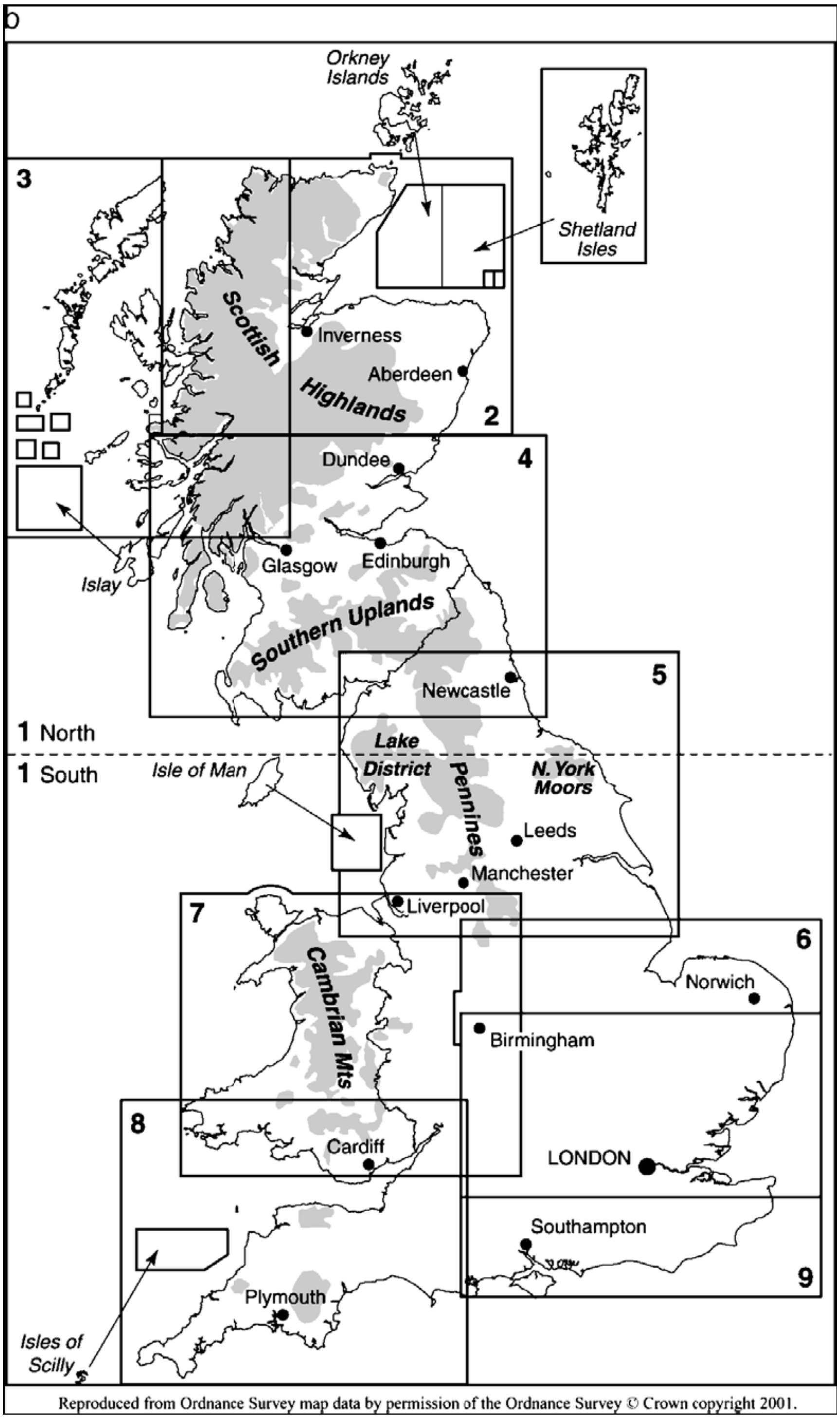

Fig. 1(b) Map of Britain and outlines of major regions used in the presentation of landform evidence. The major regions are defined by the areas covered by the Ordnance Survey 1:250,000 Travelmaster Series maps, which are recommended as a companion to this paper.

The literature available up to December 2003 on the glacial geomorphology, glacial stratigraphy and chronostratigraphy of the British Isles has been thoroughly searched in order to compile an overview of the landforms and sediments suitable for a reconstruction of the limits and dynamics of the last (Dimlington Stadial) British Ice Sheet. Most of this information has been compiled on the Glacial Map of Britain (Clark et al., 2004), which also contains information captured from published and unpublished, BGS ‘drift’ sheets. This information is not reviewed here as much of it is not accompanied by published written explanation, but it is listed in the GIS data base. The review is organized into three main sections pertaining to: (a) glacier flow directions; (b) ice marginal landforms and ice-dammed lakes; and (c) stratigraphic/dating controls on glaciation chronology. The section on ice marginal landforms and ice-dammed lakes is further subdivided according to regions as outlined on Fig. 1b. Most of the evidence reported here has been compiled on the Glacial Map of Britain (Clark et al., 2004) and it is recommended that this is used as a visual guide to the disposition of cited features along with 1:250,000 maps and associated place name indexes (e.g. Ordnance Survey Travelmaster series). In the final section we identify the main areas of weakness in our knowledge and those research directions that may benefit most from further attention.

2. The nature of the published evidence

2.1 Glacier flow directions

2.1.1. Erratics and till lithology

General glacier flow directions can be reconstructed based upon erratic distributions and till lithological properties, although it is important to note that erratic lithologies may have been transported by more than one glacier flow phase prior to their deposition. A large number of erratic source outcrops have been identified in Scotland (see Sissons, 1967a; Sutherland, 1984 for overview) and some appear to have produced Dubawnt type dispersal trains (Dyke and Morris, 1988) beneath individual ice streams. For example, the essexite erratic train was dispersed from its source outcrop near Lennox town in to the Firth of Forth by an easterly flowing ice stream (Peach, 1909; Shakesby, 1978). The Glen Orchy kentallenite and Glen Strae augite/diorite also occur in a very narrow band in a dispersal train stretching westwards towards Oban (Kynaston and Hill, 1908; Thorp1987). The widespread extent of Rannoch granodiorite erratics has been important in the reconstruction of radial flow from an ice centre in the western Grampians (Hinxman et al., 1923; Thorp,1987). Similarly, in Caithness the northwestwards dispersal of an arrow train of conglomerate erratics from Loch Sarclet on the south coast documents the passage of an ice stream fed by an ice dispersal centre over the outer Moray Firth (Peach and Horne, 1881a; see below). In the Assynt area of northwest Scotland Lawson (1995) documents the westerly flow of ice over outcrops of Torridonians and stone, identifiable in distinct down-ice tails of sandstone erratics.

The distribution of shelly tills in northern Scotland provides evidence of onshore flow of glacier ice during the last glaciation. Good examples are the shelly tills of Orkney and Caithness (Peach and Horne, 1880, 1881a; Hall et al., 2002), northern Buchan/Spey Bay (Jamieson, 1906) and northwest Lewis (Geikie, 1873, 1878; Baden-Powell, 1938). These deposits record the flow of glacier ice from an ice divide located over the outer Moray Firth (Hall and Whittington, 1989) when the British and Scandinavian ice sheets were coalescent (Sissons, 1965, 1967a). The easternmost limit of local tills in Caithness, largely represented by a line drawn between Reay and Berriedale, documents the flow of ice from the uplands of southern Caithness and Sutherland. Although this limit has been assigned previously to either the Aberdeen–Lammermuir Readvance (Sissons, 1967a) or the maximum limit of the last glaciation (Sutherland, 1984), there is no firm evidence to suggest that the local tills and shelly tills are not coeval and therefore record the junction of Caithness/Sutherland and Moray Firth ice at the LGM (Peach and Horne, 1881b; Hall and Whittington, 1989; Hall et al., 2002).

The erratic content of the tills in eastern and northeastern England has been used to distinguish different ice source areas but without reference to any one specific glaciation (cf. Harmer, 1928; Raistrick, 1931a,b). Scandinavian erratics found in Devensian tills are thought to have been reworked from older till units (Catt, 1991a). Within Devensian tills along the east coast, erratics have been recorded from two distinct geographic areas. In the lower till (Skipsea Till), erratics indicate a northern source from the Southern Uplands and the Cheviots, whilst the upper (Withernsea Till) has erratics from the Lake District and Pennines suggesting changing dominance of ice source areas during the course of the last glaciation (Catt, 1991b).

A prominent marker erratic in northern England is Shap granite which has a small outcrop in the eastern Lake District, but has a widespread distribution having been reported both around Kendal to the south and in the Vale of Eden to the north (Howarth, 1908a,b,c,d,e; Letzer, 1978). However, it is the far travelled easterly distribution through Stainmore to the east coast and down the Vale of York that is most significant, revealing a breaching of the Pennines by ice from Scotland and the Lake District (Raistrick, 1931b; Letzer, 1978; Catt, 1991a,b). Another distinctive erratic train is found at Norber, near Settle where conspicuous blocks of Silurian gritstone have been emplaced on Carboniferous limestone by upward glacial transportation in a southwesterly direction (cf. Huddart, 2002a and references therein).

2.1.2 Erosional features

Roches moutonnées and striae may provide clear indicators of local and regional ice flow (e.g. Gemmell et al., 1986; Sharp et al., 1989; Sugden et al., 1992; Lawson, 1996; Viellette et al., 1999), although multiple ice-flow directions driven by complexities in bedrock topography need to be taken into account (e.g. Rea et al., 2000). On the Lleyn Peninsula, North Wales, striae measured by McCarroll (1991) and Gibbons and McCarroll (1993) reveal ice flow from NE to SW on the northern shore of the west coast, veering to north–south along the hills in the westernmost tip of the peninsula. Welsh ice produced SW to WSW orientated striae on the inland part of St Tudwals with striae aligned east–west at the east end of Porth Neigwl. Striae record an onshore flow owards the west to northwest on the south tip of St Tudwals.

Reconstructions of an ice divide for an independent Outer Hebrides Ice Cap lying immediately over the west coast of the Uists at the LGM are based upon the transport of erratics and erosional ice-directional indicators (von Weymarn, 1979; Coward, 1977; Flinn, 1978a,b; Peacock and Ross, 1978; Peacock, 1984,

1991). This ice cap coalesced with mainland ice to the east of the Uists and deflected it southwestwards across the Sea of the Hebrides (Davies et al., 1984; Selby, 1989; Peacock et al., 1992). Some controversy surrounds the reconstruction of ice flow directions within the independent ice cap that formed over the Shetland Islands (Peach and Horne, 1879, 1881b,c;
Hoppe, 1974; Flinn, 1977, 1978a,b), although striae and erratic distributions appear to support reconstructions of an ice divide centred on the long axis of the islands (Chapelhowe, 1965; Mykura and Phemister, 1976; Flinn, 1978a,b, 1983). In Caithness and Orkney, striae and ice-moulded bedrock (Peach and Horne, 1881a; Wilson et al., 1935) support the northwesterly
flowing ice suggested by the distribution of shelly tills and indicator erratics (Peach and Horne, 1880, 1881a; Rae, 1976; see above and Gordon, 1993a).

2.1.3. Subglacially streamlined land forms

The most characteristic landforms associated with the last glaciation in northern England are the widespread streamlined landforms, particularly drumlins. As in many other areas of former ice sheet cover, drumlins occur as fields with thousands of individual bedforms and occur in distinct areas of the former ice sheet beds either associated with the existence of fast ice streams (Clark, 1997; Clark and Stokes, 2003) or glacier lobes towards the ice margin (cf. Benn and Evans, 1998). Their origin is associated with subglacial bed deformation and/or erosion, with drumlin long axes orientated parallel to ice flow direction. Various drumlin morphologies have been identified with the more complex forms being associated with superimposed flow patterns (cf. Rose and Letzer, 1977a,b; Benn and Evans, 1998). Records of drumlins are found in the early literature for the northern England, particularly in the Geological Survey memoirs (e.g. Aveline and Hughes, 1888; Dakyns et al., 1890, 1891). However, no attempt was made to reconstruct ice flow patterns from the landforms at this time because attention was focused on striae and the distribution of erratics that identified a local ice centre over Baugh Fell. Detailed field mapping has allowed the reconstruction of a much more extensive ice divide across the western Pennines and Lake District that defined flow convergence southwest down the Lune, north down the Vale of Eden and eastwards towards Stainmore (Letzer, 1978, 1987; Mitchell, 1991b, 1994; Mitchell and Clark, 1994). One of the major problems in reconstructions of the palaeoflow patterns is that in many areas drumlins are associated with conflicting flow directions, indicating that the drumlins were not formed at one time but diachronously in association with different ice flow phases. There has also been no attempt until the production of the Glacial Map of Britain (Clark et al., 2004) to compile an overall map of these landforms so that the quality of information available for drumlins varies from written observations to detailed field mapping. For some areas high quality data are available based on field mapping at scales of 1:25,000 and 10,000, which allows the recording of all drumlins and areas of superimposed drumlins. Published maps of these areas, however, necessarily lose resolution because of reduced scale (cf Mitchell, 1991a,b). In other areas drumlins have been reported on maps at a more general level. Some research papers only record the presence of drumlins by a symbol or by a grid reference within the text and there are also large areas of northern England for which there are no available maps, such as the Tyne valley in Northumberland.

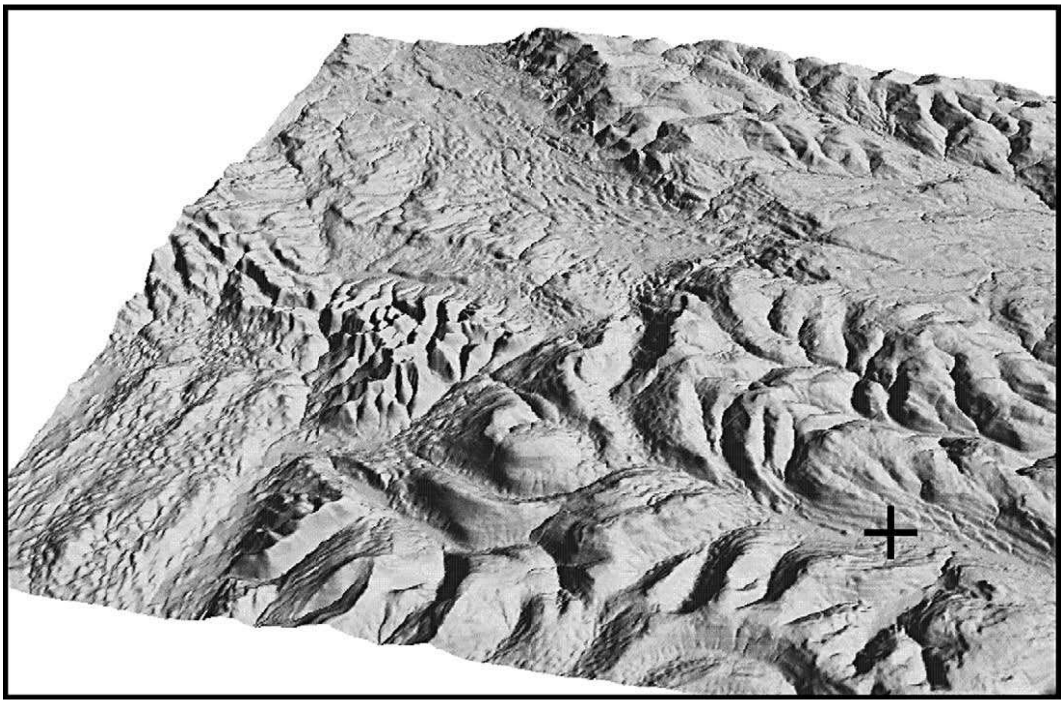

Fig. 2. Perspective view (looking NNW) of Wensleydale (marked with a cross), the Howgill Fells (left of centre) and upper Eden valley illustrating the wealth of drumlins in the area. Note the prominent streamlining and drift tails in the Wensleydale valley, and that whilst most drumlins occupy the lowlands, they can also be seen on high ground. Image is ca. 75 km across in the foreground and is derived from a 50 m grid-sized DEM. © Crown Copyright Ordnance Survey. An EDINA Digimap/JISC supplied service.

In the north of England, major drumlin fields are found in the north west counties of Lancashire and Cumbria, particularly the lowland areas of the Eden valley and Solway Firth (Hollingworth, 1931) and the Lancashire Lowlands around the Forest of Bowland (Raistrick, 1933). Drumlins are also found in the upper parts of the Yorkshire Dales, particularly in Ribblesdale and the area of upper Wensleydale (Fig. 2). Although they are generally found in lowland areas, where they record the former flow patterns as ice moved around topographic obstacles, drumlins can be found also at higher altitudes. For example, they occur at up to 600 m OD across interfluve areas in upper Wensleydale, demonstrating that the ice sheet was capable of flow directions that were independent of relief (Mitchell, 1991a,b, 1994). In the Yorkshire Dales, drumlinoid drift tails have been reported at the confluence of tributary valleys within the main dales (King, 1976; Mitchell, 1991a,b).

Detailed field mapping has been carried out in the upper Eden and upper Lune valleys, covering the ground from Shap Fells eastwards to the Pennine Escarpment, and southwards, from around Appley to the end of the drumlin field at the edge of the western Pennines (Letzer, 1978, 1987). Here the drumlins occur on the lower ground, indicating northward flow along the western edge of the Pennine escarpment. They were formed by ice that had a source area in the western Pennines, Howgill Fells and the eastern part of the Cumbria Fells. Former ice flow was also directed eastwards through Stainmore, a major through valley in the Pennines, which acted as a major conduit of ice into the Vale of York and eastern England (Catt, 1991b). Some detailed field mapping of drumlins has also been completed around Keswick indicating ice flowing northeast from the central Cumbrian Fells towards the Eden valley (Boardman, 1991) and north of Penrith where ice flow was again northwards towards Carlisle (Arthurton and Wadge, 1981). Drumlins also occur to the east of Carlisle over much of the border area towards Haltwhistle and the Tyne Gap. Drumlin forms have been noted around Bewcastle (Day, 1970) and Bellingham (Frost and Holliday, 1980), although the maps only record general locations. These confirm the general west to east trend of the drumlin long axes indicating a former ice flow eastwards from the Southern Uplands towards the Tyne Gap, as established by Hollingworth (1931) whose map unfortunately stops at this point.

Evidence of erratics in the Vale of Eden and Stainmore associated with the last ice sheet indicates that ice flowed southwards at an early stage, up valley from a Scottish source. This is confirmed by the drumlin pattern in upper Eden where there is clear convergence of forms towards Stainmore (Letzer, 1978, 1987). However, the overall pattern of the drumlins records a later stage when ice flowed down the valley from a source in the Cumbrian Mountains and western Pennines (Trotter and Hollingworth, 1932; King, 1976; Mitchell, 1991a,b; Mitchell and Clark, 1994; Huddart and Glasser, 2002). This reconstruction is not clear from the only detailed published map of this drumlin field Hollingworth, 1931), which is clearly a composite of the different diachronous landforms. The most obvious flow direction in this map is the flow northwards towards Carlisle and the Solway where it then met ice flowing in the opposite direction. These drumlins are based on field mapping of the area but although this was done at a large scale (1:10,560) the individual drumlins were not always recorded (unpublished data held by BGS) and the integrity of many of the drumlins may be suspect. This is particularly the case in the lower part of the Eden around Carlisle where the complex interaction of the different ice flows still remains to be resolved.

Drumlin fields are also well developed in many of the upper Yorkshire Dales and are associated with divergent flow from a major ice centre over this upland area (Dakyns et al., 1891; Raistrick, 1926; Mitchell, 1991a, 1991b). Drumlins have been mapped in upper Wensleydale and associated tributary dales and show a pronounced convergent flow pattern and changes in drumlin morphology, particularly length, that suggest the former existence of an ice stream in this valley (Mitchell, 1991b, 1994). Superimposed drumlins in key areas on either side of the ice divide indicate migration of the divide northwards with time such that Wensleydale ice eventually captured ice which had previously drained northwards into the Eden (Mitchell, 1994). In the Yorkshire Dales, drumlinoid drift tails have been reported at the confluence of tributary valleys with the main dales (King, 1976; Mitchell, 1991a).

Drumlins are also recorded in the upper Lune valley to the west of the Howgill Fells and the Rawthey valley. These converge to indicate the location of a major southwest ice flow towards the Irish Sea from an ice divide over the Howgill Fells/Baugh Fell area that extended towards the Lake District (Mitchell and Clark, 1994). There is also a large drumlin field in Ribblesdale in the southern part of the Yorkshire Dales. Although this is extremely well known and often cited as a "textbook" drumlin field, no detailed mapping has been completed in this area except in the northern part to allow definition of an ice divide across the southern part of Wensleydale (Mitchell, 1991b). A summary map of the drumlins was included in Raistrick (1931a,b, 1933) and showed a former ice flow direction to the south, but with topographic divergence of flow at Ribbleshead. It also showed convergence of flow from the eastern tributary valleys indicating a southwestward flow and suggestive of a further ice centre in this area in upper Wharfedale which is not reported in the literature. In the southern part of the valley towards the Craven Lowlands, drumlins have been reported around Settle (Arthurton et al., 1988). As in the upper Ribble, these indicate a southern palaeoflow direction with a divergence of flow into the Craven Lowlands and with flow both southeast towards Skipton and Leeds and southwest towards the lower Ribble. However, there has been no research directed to this area since Raistrick (1933).

In the Alston area of the northern Pennines, specifically the area lying to the north of Stainmore and the Tyne valley, drumlins have been reported in the upper Tees valley on the eastern side of Cross Fell. Dwerryhouse (1902) noted the occurrence of drumlins in this area, but their significance was not considered and emphasis was placed on interpreting many of the ridges as lateral moraines of a former Tees glacier. In fact there was some doubt about their origin probably because of their high altitude. Field mapping of the drumlins has confirmed that drumlins occur up to 609 m OD and are associated with local ice centred over Cross Fell with an ice divide that can be traced eastwards towards Tyne Head and southwards along the Pennine Escarpment, giving rise to a former ice flow southeast and transverse to the upper Tees. Superimposed drumlins in this area indicate migration of the ice divide with time (Mitchell, unpublished). Drumlins are also recorded further down the Tees valley, particularly around Barnard Castle (Mills and Hull, 1976; Frost, 1998). A small number of drumlins have also been reported from the South Tyne valley north of Alston indicating ice flow northwards down valley.

In the Irish Sea Lowlands, defined as the drainage area of the Southern Lake District and the valleys of the Lune and Ribble around Morecambe Bay, there are no detailed records but the presence of drumlins has been reported around Barrow-in-Furness (Greswell, 1962) and around Kendal (Vincent, 1985). Drumlins become much more widely distributed towards Lancaster where they form an extensive area of ground associated with former ice flow down the Ribble and flow convergence with the Irish Sea ice (Brandon et al., 1998). All the drumlins are associated with ice advancing from an ice divide across the Cumbrian Fells to the Western Pennines, rather than Irish Sea ice which was presumably held offshore by local ice.

Flutings and drumlins are also common in the lowlands of southern Scotland where they document the draw-down of ice streams into offshore basins. The channelling of ice down the Tweed valley is documented by a large drumlin field stretching from Tweedmouth inland to Hawick (Raistrick, 1931a,b; Rhind, 1969; Clapperton, 1971a,b). At Tweedmouth this ice stream flowed southward parallel to the Northumberland coast as recorded by striae. Glacier flow directions over Ayrshire appear to have been more complex. For example (Monro, 1999) reports two prominent sets of drumlin and crag-and-tail orientations in the Irvine District, an earlier east–west trend being overprinted by a later north–south orientation. This is similar to drumlin and crag-and tail trends in central Ayrshire where the later flow deviates from north–south to northeast–southwest (Richey et al., 1930; Eyles et al., 1949; Holden, 1977; Monro, 1999). The deflection from a southerly to a southwesterly flow is thought to represent a deflection of Highland ice by the radially draining Southern Upland ice. The northern part of the Solway Firth around The Machars of Galloway is also known to have a large number of drumlins (Cutler, 1979). Complex overprinting of flow-sets of drumlins and flutings has been documented by Salt (2001) from this area of southwest Scotland. He identifies twenty-eight flow sets, seven of which record topographically constrained, late stage glacier flow directions. Only the most recent flow phase (surge) in the Loch Ryan basin is associated with unequivocal ice-marginal deposits and landforms, although an ice-contact grounding line fan/delta at 40 m OD on the south coast of the Machars documents glacilacustrine deposition after early uncoupling of Southern Uplands and Solway/Lake District ice (see below).

By far the most impressive and dense clustering of drumlins in central Scotland occurs within the Clyde Valley around Glasgow and the lowlands to the east, documenting vigorous eastward flow of Highland ice across the region (Menzies, 1976, 1981, 1996; Rose, in Jardine, 1980; Rose, 1981, 1987; Rose, in Price, 1983; Paterson et al., 1990, 1998; Forsythe et al., 1996; Cameron et al., 1998; Hall et al., 1998). The superimposition of smaller drumlins on larger "megadrumlins" has been identified by Rose and Letzer (1977a,b) and Rose (1987), suggesting that late stage flow directional changes took place.

2.1.4. Summary

A significant problem in the mapping of former glacier flow directions in Britain is the lack of systematic assessments of till lithology and subglacial bedform patterns. Whereas till provenance and bedform overprinting have been employed successfully to decipher complex ice flow patterns over large areas of the beds of the Laurentide and Fennoscandinavian ice sheets during the last glaciation (e.g. Gwyn and Dreimanis, 1979; Shetsen, 1984; Nordkalott Project, 1986/87; Dyke and Prest, 1987; Dyke and Morris, 1988; Klassen and Thompson, 1989; Boulton and Clark, 1990a,b; Kujansuu and Saarnisto, 1990; Shilts, 1993; Kleman and Borgstrom, 1996), such approaches have been piecemeal in Britain even though their potential has been demonstrated in some regions (e.g. Catt and Penny, 1966; Rose and Letzer, 1977a,b; Burek and Cubitt, 1991; Salt, 2001; Salt and Evans, 2004). The availability of satellite imagery that clearly depicts subglacial bedforms and their complex relationships should ensure that future mapping can be undertaken on a regional basis by a small team, thereby avoiding the pitfalls of variable data quality between map sheets. A systematic sampling strategy is required for the assessment of till properties, which, although expensive and logistically problematic, will deliver valuable information on the provenance of glacigenic sediments and former glacier flow histories.

2.2. Ice-marginal landforms and ice-dammed lakes

This section considers the wide range of landform evidence employed in the demarcation of former glacier margins of various ages. The evidence includes trimlines, drift limits, moraines, ice-contact glacifluvial forms, meltwater channels and icedammed lake deposits and is reviewed according to the regions outlined in Fig. 1b.

2.2.1. Northern and central Scotland

The existence of unglaciated enclaves and nunataks in northwestern Scotland and Hebrides during the LGM has been a subject of considerable debate in the literature. Extensive mapping in the mountain terrain of northern Scotland has allowed the upper limits of glaciation to be determined in a number of areas using periglacial trimlines and palaeonunataks dated to the LGM by cosmogenic isotope techniques (Geikie, 1878; Ballantyne, 1990, 1997, 1999a,b; Ballantyne and McCarroll, 1995, 1997; McCarroll et al., 1995; Dahl et al., 1996; Ballantyne et al., 1997; 1998a,b; Stone et al., 1998). Upper limits of the northwest margin of the British Ice Sheet are recorded by such trimlines on South Uist in the Outer Hebrides at just below 500 m (Ballantyne and Hallam, 2001). This evidence supports reconstructions of an ice divide of an independent Outer Hebrides ice cap lying immediately over the west coast of the Uists at the LGM based upon erratic transport and erosional ice-directional indicators (see above). Submarine morainal banks on the shelf south of St Kilda (Selby, 1989; Stoker et al., 1993) mark the maximum limit of the last glaciation in the area, although St Kilda was not overrun by Outer Hebrides ice. Small valley glaciers formed on Hirta, the largest island of the St Kilda archipelago (Selby, 1989; Peacock et al., 1992; Sutherland et al., 1984). However, the possible presence of a periglacial trimline at much lower elevation on Stac Pollaidh is difficult to reconcile with the overall regional ice gradient (Bradwell and Krabbendam, 2003).

Early research indicated that a large area of Buchan in northeast Scotland had also escaped Devensian glaciation (“moraineless Buchan”; Charlesworth, 1956; Synge, 1956). This was disputed by Clapperton and Sugden (1977) although stratigraphic evidence for non-glacial conditions has been reported from the region in the form of soliflucted till overlying interstadial deposits dating to 26–22 ka BP and the widespread preservation of saprolites (Hall, 1984). This prompted renewed support for unglaciated enclaves in Scotland (Sutherland, 1984). In addition to northeast Buchan, Sutherland (1984) proposed that the tip of Caithness and the Orkney islands escaped Devensian glaciation. Although areas like Buchan, Caithness and Orkney contain only sparse glacial landforms and deposits their Devensian age glacierization has been demonstrated by offshore evidence (see below). Additionally, Hall and Whittington (1989) have provided stratigraphic evidence that the shelly tills of Caithness date to the last glaciation. Moreover, the interstadial peats of Buchan have been re-assessed as early Devensian in age and overlain by glacifluvial gravels (Hall, 1997; Hall et al., 2002; Merritt et al., 2003a,b). An integrated network of meltwater channels in the area form the most prominent record of Late Devensian deglaciation (Clapperton and Sugden, 1975, 1977), typical of recession by cold-based ice (e.g. Dyke, 1993). Other deposits documenting the deglaciation of Buchan include eskers and kames that parallel the coast north of Aberdeen (e.g. the Kippet Hills; Gemmell, 1975; Merritt, 1981; Cambridge, 1982; Smith, 1984). Of particular interest in the Kippet Hills are fossil shells similar to the Red Crag of East Anglia, indicative of onshore glacier flow by an ice stream emanating from Strathmore to the south (Hall and Connell, 1991; Merritt et al., 2003a,b). Recession of this ice led to the damming of lakes inland (Hall, 1984; Thomas, 1984c; Thomas and Connell, 1985; Connell and Hall, 1987; Aitken, 1990, 1991). Similar lakes were produced in northern Buchan at the southern margin of the Moray Firth ice (Connell and Hall, 1987). Smaller lakes in the Don Valley and Glen Nochty were dammed by local valley ice and possible ice-cored moraine dams at valley constrictions (Aitken, 1990, 1991).

It has been proposed that a small area of the northwest coast of Lewis remained unglaciated during the LGM. Central to this interpretation are the complex stratigraphic exposures along the coast in addition to local till cover and moraines (see Gordon, 1993b for review). The concept of an unglaciated enclave was proposed by von Weymarn (1974) and subsequently confirmed by Peacock (1984) and Sutherland and Walker (1984). Specifically, a raised beach (the Galson Beach) overlain by periglacial slope deposits defines a small area of coastline not overrun by Outer Hebridean ice. A drift ridge composed of glacifluvial sands and gravels and interpreted as an end moraine by Sutherland and Walker (1984) marks the limit of glacier ice moving northwards along the northeast coast of Lewis and covering the Butt of Lewis. Ice also moved from the central dispersal area on Lewis towards the north coast where its limit is marked by the edge of the till cover near Breivig. Between here and Dell Sands the lack of till over the Galson Beach demarcates the unglaciated LGM enclave. The nature of the evidence for an unglaciated enclave has been challenged by Hall (1995), who suggests that the Galson Beach is overlain by till and that there is a lack of ice-marginal features in the area. This would relegate the drift ridge on the tip of the island to a deglacial feature.

Evidence for extensive glacier ice during the LGM has been uncovered during surveys of the offshore areas of northern Scotland. Submarine moraines have been located at the edge of the continental shelf to the north of Lewis and west of Orkney and Shetland (Stoker, 1988). Additionally, a large submarine moraine is located off the northeast coast of Shetland (Stoker et al., 1993; see below for dating evidence). A series of prominent submarine end moraine ridges (Wee Bankie Moraine) mark the eastern edge of the glacial sediments of the Wee Bankie Formation off the east coast of Scotland (Thomson and Eden, 1977; Stoker et al., 1985). Beyond the Wee Bankie moraine the Marr Bank Formation contains glacimarine sediments. These features and sediments can be traced to the outer Moray Firth where tills extend up to the Bosies Bank moraine (Hall and Bent, 1990). The apparent lack of till beyond the Wee Bankie and Bosies Bank moraines has been used previously to support their interpretation as the Late Devensian ice limit but recent research on cores from the North Sea indicates a more extensive ice coverage (see below). In the central North Sea approximately between latitudes 588 and 598N, Jansen (1976) reported ice marginal deposits associated with the Fennoscandinavian Ice Sheet, named the Hills Deposits. They include morainic terrain separated from coeval glacimarine sediments by a 30 m high, single end moraine ridge. The glacimarine sediments are the lateral equivalent of the Marr Bank Formation.

In addition to the prominent deglacial moraines offshore like the Wee Bankie and Bosies Bank moraines, ice-marginal landforms and sediment accumulations are widespread on land. Glacier recession is also clearly marked by meltwater channels in many drainage basins. Several readvances of Scottishcentred ice have been proposed based upon extensive and substantial moraine systems. The Aberdeen–Lammermuir Readvance (Synge, 1956; Sissons,1967a, 1981) was proposed to account for almost continuous assemblages of glacifluvial sands and gravels. In the Aberdeen area this is typified by a 3 km wide belt of ice-marginal deposits stretching across the lower Dee valley (Brown, 1993). Linear accumulations of glacifluvial deposits were mapped in the Aberdeen area by Synge (1963b) and associated with glacier margins during readvances (e.g. Dinnet Readvance and Aberdeen Readvance). Such features were previously assigned to a Dinnet Readvance by Bremner (1931a) and Synge (1956) and then reinterpreted as stagnation deposits by Clapperton and Sugden (1972, 1977), Sugden and Clapperton (1975) and Murdoch (1975). The most recent assessment of the ice-marginal deposits in the Dee Valley by Brown (1993) proposes a marginal supraglacial origin similar to the evolution of moraines on sub-polar glaciers. This explains their linearity and diverse sedimentology and indicates a series of glacier recessional stages. Additionally, the meltwater channels of the area are subdivided by Brown into discordant (subglacial) and concordant (ice-marginal) types.

The marginal recession of the Forth glacier is well documented by a series of spectacular meltwater channels and glacilacustrine sediments on the southern margin of the Ochil Hills (Russell, 1995). Two lake levels at 269 m and 233 m are recorded for an ice-dammed lake in Glen Devon. Further meltwater channels on the north side of the Ochil Hills and on the hillsides to the west of Auchterarder document recession by a glacier lobe in Strathallan (Sissons, 1961a; BGS Drift Sheet 39, Stirling). The concordant, inset nature of these channels suggests that they were cut by lateral meltwater draining along the cold-based margins of a receding glacier (cf. Dyke, 1993). The Perth Readvance was proposed by Simpson (1933) and supported by Sissons (1963a, 1964, 1967b) based upon the distribution of ice-marginal depositional complexes, particularly glacifluvial landform assemblages, and their association with raised shorelines. Although the Perth Readvance was later rejected by Paterson (1974) and even by Sissons (1974a,b, 1976a, 1981) himself, account needs to be taken of the glacifluvial landform assemblages originally used in its identification. A glacier margin associated with the “Main Perth Shoreline” has been confirmed in the Stirling area by Francis et al. (1970) but no evidence for a readvance has been forthcoming. Another ice margin further up valley at Doune Lodge is represented by “dead ice terrain” (Smith et al., 1978) and records the recession of the Teith glacier from the Forth valley.

In the northwest Highlands a prominent moraine system has been assigned readvance status by Robinson and Ballantyne (1979) and Ballantyne (1993) and named the Wester Ross Readvance (cf. Sissons and Dawson, 1981). This former ice margin is demarcated by discontinuous moraine ridges, drift limits and boulder belts. Eskers join the moraine on its ice proximal sides in some places. Together the various sections of the moraine provide an outline of glacier lobes that occupied Loch Torridon, Loch Gairloch and Loch Ewe.

A widespread readvance or stillstand, termed the Otter Ferry Stage was proposed by Sutherland (1981, 1984) to account for abrupt drops in sea level in the sea lochs at the head of the Firth of Clyde. This period of ice-marginal stabilization was explained by Sutherland (1984) as a response to glacier recession into areas of shallow water over localized fjord sills. In the Oban area Synge (1966) and Synge and Stephens (1966) proposed a readvance or stillstand base upon landforms they called the Oban–Ford moraine. Gray and Sutherland (1977) later questioned the evidence of synchronicity for the Oban–Ford moraine, proposing instead brief and diachronous halts in overall recession. However, they did highlight a possible stillstand in the Ford–Kilmartin area and at Loch Scamadale based upon a significant sea level fall across ice-marginal deposits and equated it with their Otter Ferry Stage.

McCann (1963) documents two re-advance moraines at the head of Loch Carron. The innermost of these, the Strath Carron End Moraine, is now equated with the Loch Lomond Readvance but the outermost Tullich Moraine is presumably from an earlier period. McCann depicts a regional ice margin that associates the Tullich Moraine with parts of the moraine system now related to the Wester Ross readvance.

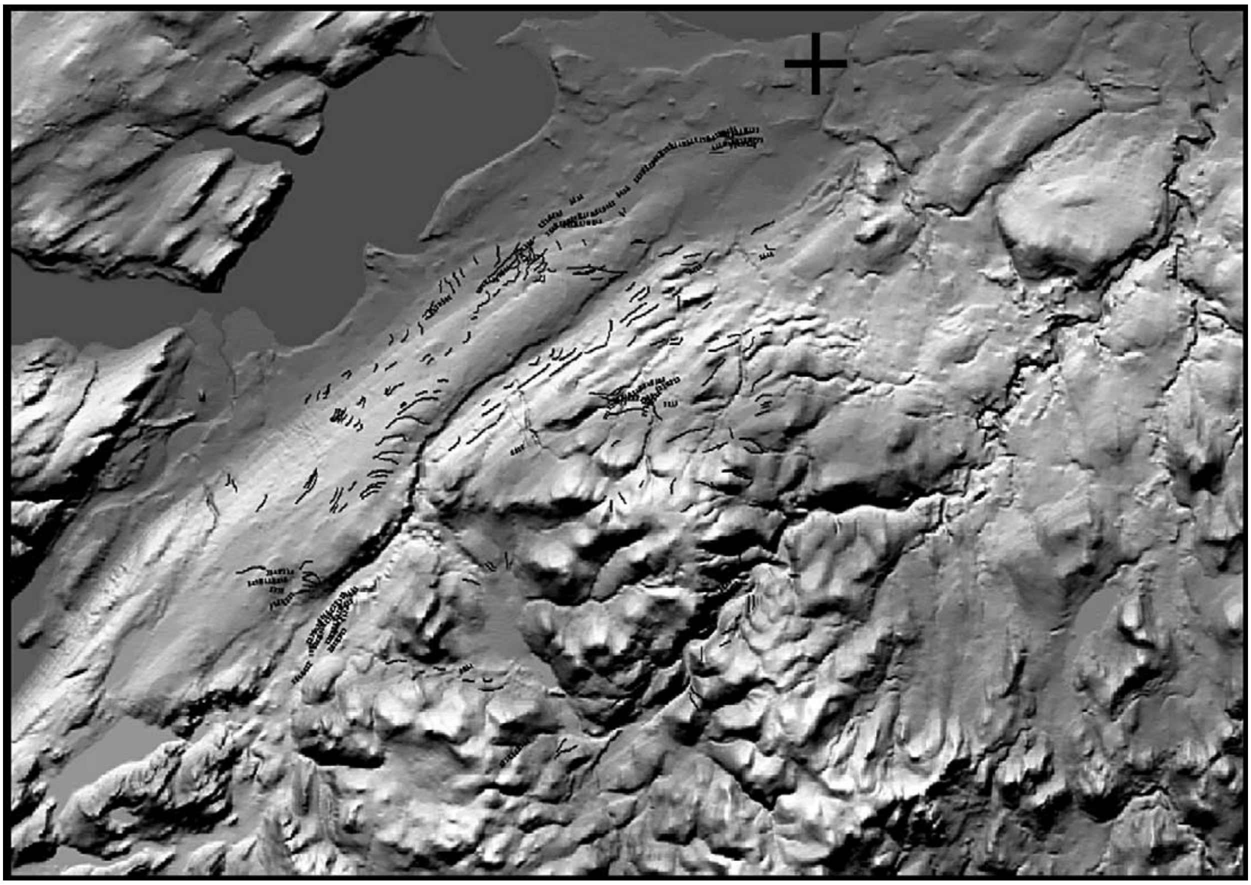

Fig. 3. Nadir view of a solar-shaded DEM of the terrain containing the Flemington eskers and associated kame and kettle topography to the south of the Moray Firth in Scotland. Location of Nairn is marked with a cross. Esker ridges (marked as "railway tracks") and meltwater channels (black lines), as depicted by Merritt et al. (1995) and on the Glacial Map of Britain (Clark et al., 2004) are highlighted. Note the large lateral meltwater channels slightly NE of the image centre. These are outside the area mapped by previous researchers and therefore not included on the Glacial Map of Britain. Similarly, the fluted terrain visible in the NW sector of the image and documenting the flow direction of the former Moray Firth Ice Stream has never been mapped. Image is 45 km across and is derived from a 50-m grid-sized DEM. © Crown Copyright Ordnance Survey. An EDINA Digimap/JISC supplied service.

Following on from the identification of recessional stages/readvances by Kirk et al. (1966), Smith (1968, 1977), Synge (1977a,b) and Synge and Smith (1980) in the Moray Firth, Merritt et al. (1995) and Fletcher et al. (1996) provide a comprehensive geomorphic record of the former Moray Firth ice stream. Recessional positions are elucidated by the decline in marine limit altitudes in the Moray Firth (Firth, 1989a; Firth and Haggart, 1989). During initial downwasting large numbers of meltwater channels were excavated on the highland to the south of the Moray Firth and considerable quantities of glacifluvial sediment were deposited in the Flemington Eskers and associated kame and kettle topography (Fig. 3; Auton, 1992; Gordon and Auton, 1993). The Flemington Eskers (also known as the Kildrummie Kames) are probably genetically linked to the equally impressive Littlemills eskers located at the mouth of the Great Glen (Gordon, 1993c). Both esker systems are associated with large expanses of flat-topped and kettled kames deposited in wide ice-walled channels that were produced by ice tunnel collapse during the more advanced stages of deglaciation. Transverse crevasse-fill ridges were also deposited on the higher land between the Nairn Valley and the Moray Firth. Merritt et al. (1995) and Fletcher et al. (1996) also provide corroborative evidence for earlier proposals of a readvance, the Ardersier Oscillation at approximately 13 ka BP. Although this readvance was questioned by Firth (1989b), clear evidence for an ice-marginal oscillation is manifest in glacitectonized marine sediments (Ardersier Silts Formation, the local Errol Beds) in a large thrust or push moraine that stretches along the south shore of the Inverness Firth from Ardersier to Alturlie Point. A later ice margin, related to either a stillstand or readvance, was located at Alturlie Point as recorded by a grounding-line subaqueous fan (Merritt et al., 1995).

Glacilacustrine deposits in lower Strathspey document the earlier recession of Moray Firth ice from the south coast and the production of ice-dammed lakes at 189 m (Glacial Lake Knockando), 122 m (Glacial Lake Rothes) and finally at 73 m (Glacial Lake Fochabers; Hinxman and Wilson, 1902; Bremner, 1931b). The occurrence of till and glacifluvial sediments overlying glacilacustrine sediments suggests that Glacial Lake Fochabers was the product of a minor glacier readvance (Peacock et al., 1968; Aitken et al., 1979).

A complex network of meltwater channels (Sugden and Clapperton, 1975) documents early recession of glacier ice in the northern Cairngorms. Mapping of moraines and ice-contact deltas on the northern flanks of the Cairngorm Mountains by Brazier et al. (1996a,b; 1998) has elucidated the nature of deglaciation in the area. The landform–sediment assemblages demonstrate that ice-dammed lakes existed between the Cairngorm outlet glaciers and the Glenmore lobe of the Scottish ice sheet as they became uncoupled and that minor readvances characterized glacier recession. More recently, Everest (2002) has reported 15 ka BP dates for the moraines that document the damming of the mouth of Gleann Einich by the Glen More lobe. He reports similar ages for lateral moraines deposited by Cairngorm ice on the western side of Gleann Einich and on the eastern side of Glen Dee. The subsequent deglaciation of the Glenmore Basin and upper Strathspey is recorded by meltwater channels, eskers and kames as mapped by Young (1974, 1978). The density and distribution of glacifluvial ridges interpreted as eskers in research papers of this period have more recently been re-interpreted as supraglacial channels fills or dissected ice-contact outwash, excellent examples of the former occurring in the Nith valley, Dumfriesshire where they terminate at the ice-contact face of a proglacial outwash plain (cf. Stone, 1959; Huddart, 1999).

Pre-Loch Lomond Stadial ice sheet moraines have been identified on southern Skye by Ballantyne (1988), Ballantyne and Benn (1991) and Benn (1991). The Strollamus boulder moraine on the southshore of Loch na Cairidh is cross-cut by Loch Lomond Readvance moraines and was originally interpreted by Ballantyne (1988) as a lateral moraine deposited by an ice lobe of possible Wester Ross Readvance age. Benn (1990, 1991) later re-interpreted the feature as a medial moraine deposited in the lee of Beinn na Caillich during ice sheet deglaciation. A further pre-Loch Lomond Stadial moraine occurs in Glas Choire in the Kyleakin Hills, interpreted by Ballantyne (1988) and Ballantyne and Benn (1991) as a recessional moraine deposited by a lobe of the Devensian ice sheet as it retreated across the col to the south. A remarkable medial moraine deposited by the Devensian ice sheet occurs on Jura where it documents former northwesterly ice flow over the island (Dawson, 1979). Evidence of moraines and a possible minor readvance by Devensian ice on Islay occurs at Loch Indaal (Synge and Stephens, 1966; Peacock and Merritt, 1997). Areas of hummocky moraine have been mapped on Islay and Jura by McCann (1964), Dawson (1982) and Peacock (1984) and related to ice sheet recession from the islands. The innermost of these moraines is composed of ice-contact deltaic foresets that have been glacitectonized and overprinted by till.

Moraines deposited during ice sheet recession occur in other parts of northern Scotland where their implications for glacier dynamics and age remain to be fully elucidated. Numerous transverse moraine ridges document recession of glacier ice on Harris and are interpreted by Peacock (1984, 1991) as the products of active glacier margins. Mykura and Phemister (1976) report a moraine belt on Papa Stour, Shetland Islands that may represent ice marginal stabilization after recession from the continental shelf. Moraines are rare in nearby Caithness and comprise ridges and mounds of gravel lying on both the local and shelly tills (Peach and Horne, 1881b; Crampton and Carruthers, 1914). They are thought to relate to local ice that became dominant after the recession of the northerly flowing Moray Firth ice.

It is evident from a number of studies reviewed above that meltwater channels are ubiquitous features in the glacial landform legacy of Scotland. Many areas of meltwater channels were mapped in the 1960s and 1970s following the impact of the seminal work of the Scandinavian geomorphologists on glacial drainage networks (e.g. Mannerfelt, 1945, 1949). Particularly significant was a paper by Sissons (1958a) in which the influential ideas of Kendall (1902) on overspill channels were finally overthrown (see Ballantyne and Gray, 1984 and Evans, in press for review). There followed a large number of papers on the distribution and interpretation of glacial meltwater channels in Scotland (e.g. Sissons, 1958a,b,c, 1960, 1961a,b, 1963b; Price, 1960, 1963, 1983; Derbyshire, 1961), demonstrating that glacial meltwater drained subglacially, marginally and proglacially and that lake overspill channels were likely to occur only where other unequivocal evidence for ice-dammed lakes existed (e.g. Russell, 1995).

2.2.2. Southern Scotland and Northumberland

Considerable accumulations of glacifluvial sediment have been often used as evidence for readvances by the receding Highland and Southern Upland ice masses. Two eminent researchers of Scottish glacial geomorphology, J.K. Charlesworth and J.B. Sissons were responsible for championing major regional readvances, including the Lammermuir–Stranraer, Aberdeen–Lammermuir and Perth Readvances (detailed briefly above). The Lammermuir–Stranraer Readvance of Charlesworth (1926a,b) and Sissons (1961c), although later rejected by Sissons (1974a,b, 1976b), was based upon the occurrence of a prominent belt of glacifluvial landforms in the Stranraer/ Galloway coast area and in central Scotland, specifically around the Pentland and Lammermuir Hills. The lack of end and lateral moraines at the margins of the proposed Aberdeen–Lammermuir and Perth readvances was initially explained by Sissons (1967a) as the result of rapid ice stagnation before he rejected their readvance status. Charlesworth's (1926a,b) "kame moraines" of his Lammermuir–Stranraer Readvance included some substantial accumulations of glacifluvial sediment such as the "Carstairs kames". The landforms at Carstairs have been the subject of considerable debate in the glacial literature (e.g. Gregory, 1915, 1926; Goodlet, 1964; Sissons, 1967a; McLellan, 1969; Laxton and Nickless, 1980; Jenkins, 1991; Gordon, 1993d; Huddart and Bennett, 1997), but a concensus on their origin is starting to emerge that involves an esker origin in the coalescence zone of Highland and Southern Upland ice. Specifically, Thomas and Montague (1997) propose that the esker was deposited on the floor of a lake that developed between the two ice masses as they uncoupled. Some supraglacial deposition resulted from the eventual emergence of the subglacial conduit onto the wasting glacier surface. Like the extensive glacifluvial sediments and landforms further north, Charlesworth's "kame moraine" records wastage of the ice sheet in valley settings during standstills rather than a regional readvance (e.g. Cutler, 1979; Huddart, 1999). This physiographic control on glacifluvial landform distribution is particularly striking in compilation maps of glacifluvial sediment in Scotland (e.g. Sutherland, 1991; Gordon and Sutherland, 1993) and explains a large number of esker complexes and associated ice-marginal kames (e.g. Gregory, 1913; McLellan, 1969; Insch, 1976; Terwindt and Augustinus, 1985; Salt, 2001).

After the separation of the Highland and Southern Uplands ice and the production of an interlobate lake and esker over Carstairs, a series of ice-dammed lakes occupied the Clyde valley during recession of the Clyde glacier westwards. Deltas and glacilacustrine sediments document lake levels between 60 and 90 m OD (Price et al., 1980; Clough et al., 1920, 1925). The sediments of this Lake Clydesdale (Bellshill and Ross Members; Browne and McMillan, 1989; Hall et al., 1998) lie upstream of an arcuate moraine at Blantyreferme in southeast Glasgow, which documents a stillstand during glacier recession (Clough et al., 1925; Hall et al., 1998). A further lake (Lake Kelvin) was dammed in the Kelvin valley.

The recession of ice into the mountains of the Southern Uplands is documented by sequences of inset latero-frontal moraines according to Charlesworth (1926a), although many if not all of these "moraines" are most likely ice-contact glacifluvial landform assemblages (e.g. Huddart, 1999). Firm evidence for a readvance of Highland ice in southern Scotland has been provided, however, by Salt (2001), who has identified ice-marginal landforms at the outer limit of long flutings aligned parallel with Loch Ryan. These ice-marginal landforms were originally identified by Charlesworth (1926a,b) and associated with his Stranraer–Lammermuir Readvance. Salt (2001) interprets the landforms as a complex of glacitectonically thrust ridges and depressions, a lateral moraine and kame and kettle topography grading northwards into large drift lineations (flutings). This assemblage of landforms, being similar to the surging glacier landsystem of Evans and Rea (1999), strongly suggests that fast flowing ice moved down the Loch Ryan basin during a surge. The earliest evidence of ice-marginal deposition in southwest Scotland is provided by an ice-contact grounding line fan/delta at 40 m OD on the south Machars (Salt, 2001). This feature was originally interpreted by Charlesworth (1926a) as part of his "kame moraine" but has more recently been interpreted as a delta formed in a glacilacustrine depo-centre situated between Southern Uplands and Solway/Lake District ice after their initial uncoupling (Salt, 2001). Further local evidence for such a lake is an extensive area of large P-forms that disappear under till above approximately 40 m OD, thereby suggesting that till has been widely removed below that altitude by wave-washing.

A dense concentration of meltwater channels occurs in southeast Scotland and neighbouring Northumberland, where many notable examples record flow around the eastern flank of the Cheviots and southward along the coastal plain (Raistrick, 1931b; Clapperton, 1971a,b). Many of these were originally interpreted as glacial lake overflow channels (Kendall and Muff, 1901, 1903) but have since been explained as primarily of subglacial origin (Sissons, 1958a; Derbyshire, 1961; Clapperton, 1968). Classic examples include the Humbleton Hill and The Trows channels (Huddart, 2002b).

Sand and gravel landforms around Coldstream were interpreted by Charlesworth (1957) as moraine ridges and used by Sissons (1967a) as part of his Aberdeen–Lammermuir Readvance. Carruthers et al. (1930, 1932) had previously mapped an arcuate belt of sands and gravels east of the Cheviot massif and north of the River Aln, and suggested they marked a possible margin of the receding Tweed glacier. Included within this kame belt was the Bradford Kaims or Bradford-Charlton Gravels which mark the northern limit of a linear belt of sand and gravel on the coastal plain near Bamburgh. There has been much debate in the literature as to the origin of these landforms and they are generally thought to be eskers associated with a northward retreating ice margin (cf. Gregory, 1922; Parsons, 1966; Huddart, 2002c and references therein). Also in the vicinity is the "Cornhill kettle moraine", a large expanse of ice contact forms that continue towards Wooler on the margin of the Cheviots (Douglas, 1991). The large accumulations of glacifluvial sediment in the River Aln/lower River Beamish/Wooler area document deglaciation of the eastern Cheviots (Clapperton, 1971a,b). As the Tweed ice stream downwasted and uncoupled from Tyne ice flowing round the south of the Cheviot Massif glacial lakes occupied the lower River Beamish and River Aln. Meltwater channels, eskers, kame terraces, deltas and laminated lake sediments record the recession of the Tweed ice stream during which ice-dammed lakes were temporarily ponded in the northerly draining valleys of the River Till and its tributaries. The last lake was at 45 m and occupied the Milfield basin.

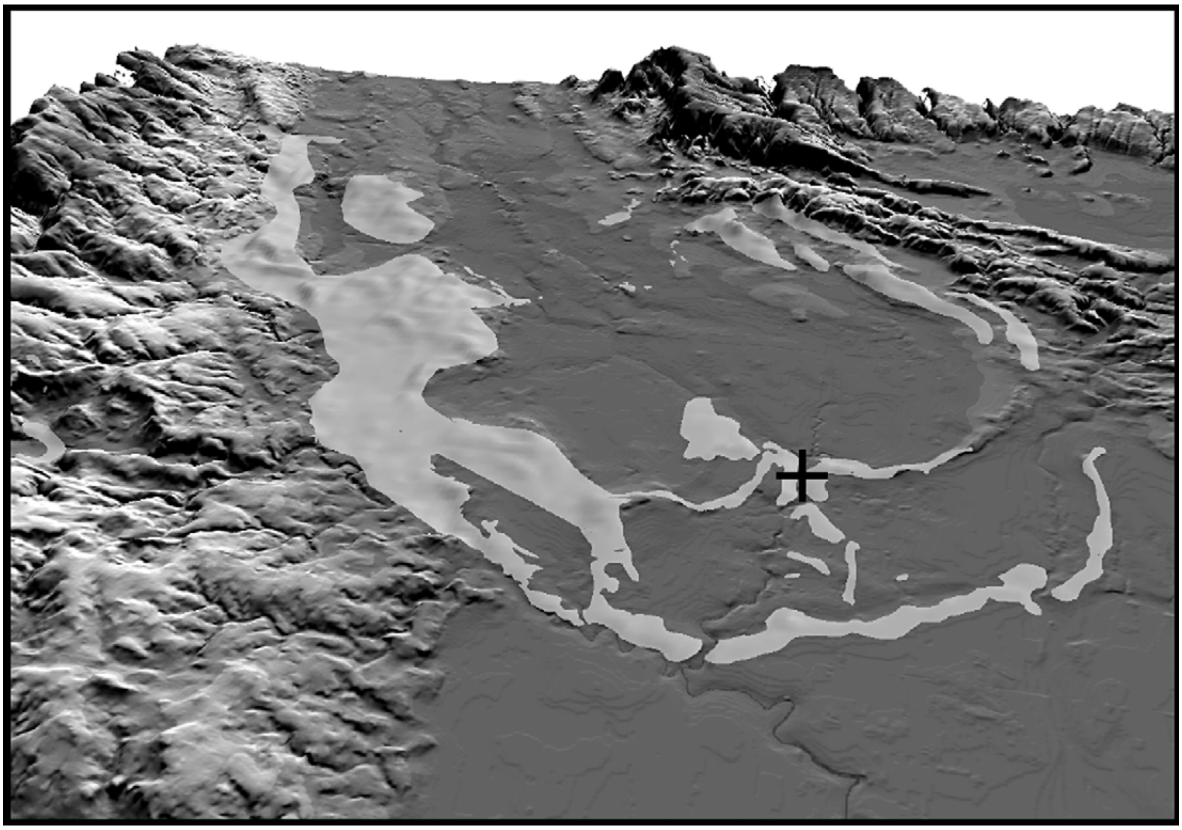

Fig. 4. Perspective view (looking north) of the topography of the Vale of York. Location of York is marked by a cross. Overlaid on the DEM (in light grey) are the York and Escrick moraines and the Linton-Stutton gravels, as mapped by various researchers and depicted on the Glacial Map of Britain (Clark et al., 2004). Note that the DEM picks out sections of the moraine ridges not mapped previously. Image is ca. 60 km across in the foreground and is derived from a 50-m grid-sized DEM. © Crown Copyright Ordnance Survey. An EDINA Digimap/JISC supplied service.

2.2.3. North east and eastern England

In the southern part of the Vale of York three moraines/ice-marginal glacifluvial landform assemblages are recognized, specifically the York moraine, the Escrick moraine and the Linton-Stutton gravels (Fig. 4). The York and Escrick moraines (Dakyns et al., 1886; Lewis, 1887, 1894; Kendall and Wroot, 1924; Edwards et al., 1950) form well defined ridges across the vale and continue eastwards onto the flanks of the Howardian Hills and northwards to the Coxwold Gap where the Ampleforth (=Escrick) moraine blocks the west end of the Vale of Pickering. To the west the moraine is less clear but can be identified by the pattern of marginal meltwater channels at Tadcaster and the southeasterly diversion of the River Wharfe between Wetherby and Boston Spa which now follows a route on the distal side of the Escrick moraine (Kendall and Wroot, 1924; Melmore and Harrison, 1934; Cooper and Gibson, 2003). Bulmer Beck occupies a lateral meltwater channel cut along the former ice margin when it stood at the Escrick moraine. North of Wetherby the York and Escrick moraines merge to form one wide belt of hummocky topography that is mapped along the western margin of the Vale of York (Cooper and Burgess, 1993) where the Vale of York ice was confluent with ice flowing out of Wensleydale (Raistrick, 1932).

The Linton-Stutton gravels occur outside the Escrick moraine between Tadcaster and Wetherby and form an extensive but discontinuous spread of ice-contact gravel mounds with subdued relief. Sections show extensive deformation and they are thought to have formed at an earlier stage of the last glaciation than the York–Escrick moraines (Edwards et al., 1950). These deposits were interpreted as a kame by Edwards et al. (1950) but as an esker north of Bramham by Straw (1979). Straw (1979) suggested a submarginal (or interlobate?) origin for this "esker" associated with the Nidderdale Glacier, which was deflected over Harrogate by the Vale of York lobe during a pre-Escrick phase [Early Main Dales Glaciation of Edwards (1938)]. Lateral meltwater channels along the former western margin of the Vale of York lobe were mapped by Kendall and Wroot (1924) and Melmore and Harrison (1934). Where these lay outside the York–Escrick limits and in association with the Linton-Stutton gravels, they were regarded as pre-LGM. The diversion of the River Wharfe was associated with the construction of the Escrick moraine because it now occupies a course that follows the distal side of the moraine and does not follow its preglacial course. Gaunt (1981) suggested that the Vale of York lobe initially advanced as a surge into Glacial Lake Humber all the way to Wroot in Lincolnshire based on one main line of evidence, the "Wroot Thorne Gravels". The margin later stabilized at the York and Escrick moraines when the level of Glacial Lake Humber dropped. Straw (1979) had previously proposed that the till to the south of the York and Escrick moraines was "older", probably early Devensian in age. There has however been continued debate regarding the existence of this surge and in a recent review of the possible evidence for such an event, Straw (2002) rejects this sequence of events and argued that the stratigraphic record south of the Escrick moraine is pre-Devensian and that the last ice sheet limit in the Vale of York is at Escrick. A lower stage Lake Humber is recorded by interdigitation of lake sediments (the "25 foot drift") and the till and gravels of the Escrick Moraine, demonstrating that the moraine was deposited at an ice-contact lake margin (Gaunt, 1970).

A prominent hummocky moraine on Flamborough Head was identified by Farrington and Mitchell (1951) who regarded it as an end moraine. This moraine assemblage widens and stretches northwards to Speeton and then to Cayton where it has been called the Speeton moraine by Valentin (1957), who allocated it to a "second readvance" on Holderness. Specifically, the prominent morainic relief was the result of the compression of ridges against the chalk scarp and the overriding of "first readvance" deposits. This was later termed the "Cayton-Speeton Stage" by Straw (1979). The Flamborough Moraine to the south was allocated to the "Seamer-Flamborough Stage" by Penny and Rawson (1969) who referred to the linearity in the moraine at Reighton as possible drumlins. Further south on Holderness, Valentin (1957) traced a line connecting the outcrops of the Kelsey Hill Gravels at Burton Agnes, Brandesburton, Bilton and Patrington out to the North Sea, which he regarded as a product of the "second readvance"; this line also encloses the extent of the Withernsea (Purple) Till (see Section 2.3). Outside this limit the margins of the "first readvance" are marked by subdued morainic ridges at Routh, Wawne and Sutton (Straw, 1979). The high land at

Dimlington has been interpreted by Catt and Penny (1966) as a push moraine formed by the thickening of tills and the disturbance of Basement Till. The various readvances proposed for this area are based upon ridges identified within the moraine belt but they are not supported by chronostratigraphic or morphostratigraphic evidence, with the exception of the deposition of the Withernsea Till (see Section 2.3). Consequently, the readvances were not recognized by Catt (1987) or more recent workers because the limits were based largely upon freshness of hummocky topography (e.g. hummocks on east Holderness) and were not verified by stratigraphy or by dating evidence. More recently, Evans et al. (2001) and Thomson and Evans (2001) have highlighted the linearity of the glacifluvial landforms of interior Holderness and proposed that the numerous parallel sand and gravel ridges in the area are former ice-contact, coalescent subaqueous fans representing recessional positions of the North Sea lobe as it receded from the interior (Fig. 5). At the east end of the Vale of Pickering the preglacial drainage of the ancestral River Derwent was reversed by hummocky drift that plugs the valley mouth south of Scarborough. A drift limit at 183 m just west of Scarborough (Straw, 1979) provides an estimate of glacier thickness at the southern flanks of the North Yorkshire Moors/Tabular Hills. During glacier occupancy of the coast, glacial lake Pickering (first suggested by Kendall, 1902) filled the Vale of Pickering and produced shorelines at 70 m and 45 m (Straw, 1979). The 70 m level is associated with a kame terrace between West Ayton and Wykeham, referred to as the Wykeham moraine and representative of the ice margin during the "Wykeham Stage" (Penny and Rawson, 1969). An outwash fan at Seamer, fed by the Mere Valley, graded to the lower Lake Pickering level of 45 m when ice stood at the Cayton-Speeton Stage limit (Penny and Rawson, 1969; Straw, 1979). A more extensive western ice limit was suggested for the Vale of Pickering by Foster (1985, 1987) based on the distribution of the Sherburn Sands, which represent an outwash train stretching from Flotmanby to Hovingham at the western end of the Vale.

a

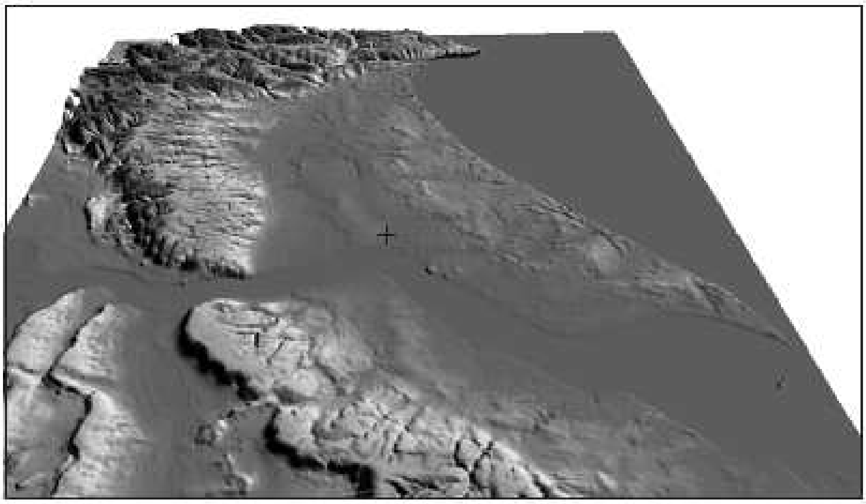

Fig. 5. (a) Perspective view (looking north) of the topography of Holderness. River Humber and Spurn Head are visible in the foreground. Cross marks location of Kingston upon Hull. To the north, on the low ground of Holderness, a subtle hummocky terrain belt is evident, showing the linearity of the coalescent subaqueous fans deposited at the ice sheet margin when it occupied a recessional position east of the Skipsea Till limit. Image is ca. 60 km across in the foreground and is derived from a 50-m grid-sized DEM. © Crown Copyright Ordnance Survey. An EDINA Digimap/JISC supplied service.

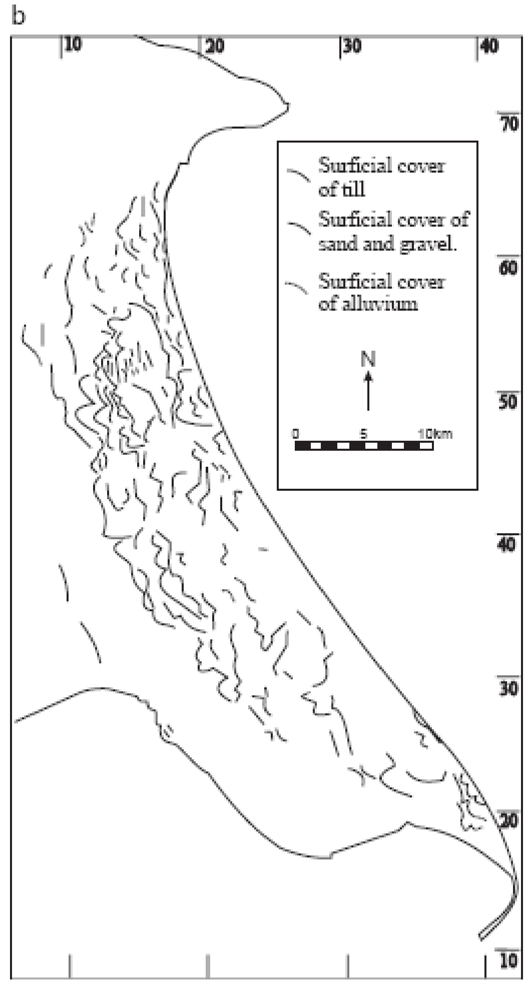

Fig. 5 (b) Map of linear ridges based upon DEM in (a) (Thomson, 2003). Ridges are coded according to the type of sediment cover

The westernmost penetration by North Sea glacier ice up the Humber estuary is marked by the Horkstow Moraine which extends from the Horkstow area across the estuary to Brough. This limit corresponds with the westernmost limits of the Skipsea (Hessle) Till (e.g. Suggate and West, 1959; Gaunt et al., 1992) and has been associated with the Stickney Moraine and the Lower Marsh Till of Lincolnshire by Straw (1979). Straw also suggested that glacial lakes Humber and Fenland were dammed by this ice margin. Gaunt (1981) referred to the Horkstow Moraine as a hummocky moraine of Skipsea type till lying in the area between Brough, Winteringham, Winterton and Horkstow. He further suggested that another moraine ridge between North and South Ferriby, which is associated with interbedded till and lacustrine deposits, was a stabilization point for the ice as it receded back out of the Humber estuary. The furthest extent of the Vale of York lobe has been mapped at Wroot in the Isle of Axholme based upon the southern limit of the elongate chain of “Younger Glacial Sand and Gravel” by Gaunt (1976) and between Doncaster and Gainsborough by Palmer (1966). This was regarded as the limit of a surge into Lake Humber by Gaunt (1981).

Glacial Lake Humber was linked to Glacial Lake Fenland at the time of the surge to Wroot (Gaunt, 1981) at an elevation of 30–33 m through the Lincoln gap (Lewis, 1887, 1894; Melmore, 1935; Straw, 1963; Palmer, 1966; Gaunt, 1981; Gaunt et al., 1971), indicating that ice blocked both the Humber and The Wash (Lake Humber I; Straw, 1979). Ice contact and meltwater-fed deltas, associated with the Horkstow/ Stickney Moraine, mark the eastern shoreline of Lake Fenland (Straw, 1979). The ca. 30 m western shoreline of Lake Humber is demarcated by a series of lake sediments stretching from a delta at the southern end of the Linton-Stutton kame chain (Wingate Hill at Towton) all the way to Doncaster and called the "Older Littoral Sands and Gravels" by Edwards (1937) and Gaunt et al. (1971). Laminated sediments in the Humberhead area ("25 foot drift") record a lower Lake Humber level of 15 m (Lake Humber II; Straw, 1979), at which time no Lake Fenland existed, indicating that The Wash was not closed off by ice. In north Lincolnshire Lake Humber II was associated with the extensive Killingholme-Hogsthorpe Moraine and the Upper Marsh Till (extending up to 80 m in the central Lincolnshire Wolds) by Straw (1961, 1969, 1979), who suggested that this assemblage of sediments and landforms represented a late Devensian "readvance" (essentially the Late Devensian maximum). Gravels west of the Killingholme Moraine are therefore the same age as Lake Humber II (Straw, 1979). To the south, Dimlington Stadial glacier ice impinged upon the area of south Lincolnshire, The Wash and northwest Norfolk as recorded by the deposition of the Hunstanton Till (Holkham Member, Lewis, 1999) which overlies Ipswichian raised beaches (Straw, 1960, 1979; Gallois, 1978). The till forms a drift limit that nowhere exceeds 37 m OD between Heacham and Morston and locally forms the Heacham Moraine/ kame. The till is associated with glacifluvial sediments (Ringstead Member, Lewis, 1999) that locally thicken to form depositional features like the Hunstanton Esker. Straw (1960, 1979) correlated these deposits with the Lower Marsh Till and the Stickney Moraine to the north of The Wash.

On the northern flank of the North Yorkshire Moors the maximum limit of Late Devensian glaciation is marked by a drift limit at 245 m in Eskdale (Straw, 1979), suggesting that the highest summits remained ice free. Considerable emphasis was placed on a series of supposed ice-dammed lakes along the northern margins of the moors as extra evidence of former ice extent after the influential work of Kendall (1902). However, the existence of these lakes was questioned after Gregory (1965) re-interpreted the important spillway evidence as subglacial in origin. Similar criticism must be levelled at the ice margins mapped by Best (1956) around the Cleveland and Hambleton Hills using "overflow channels". In Eskdale Kendall (1902) identified the Lealholm and Kildale moraines which he suggested, not unreasonably, dammed the natural drainage. He also mapped a large moraine further downstream in Glaisdale which was deposited presumably by the same glacier lobe. The existence of former ice-dammed lakes in the Tees estuary was proposed by Agar (1954), who reported laminated clays draping the glacial landforms of the area. Additionally, Glacial Lake Edder Acres and Glacial Lake Wear have been mapped by Smith and Francis (1967) and Smith (1981, 1994).

A large number of moraines were identified during early research in the Pennine Dales. In the Airedale/Bradford area, Jowett and Muff (1904) referred to, but did not map, a number of major moraines. For example, they describe the Tong Park Moraine on the valley floor on the northwest side of the Aire, the Nab Wood Moraine immediately west of Saltaire, and the Bingley Moraine in addition to lateral moraines on Hallas Rough Park immediately SSW of Cullingworth Station, and at Lanshaw Delves at the NE corner of Rumbles Moor. The latter are aligned NW–SE and extend from Rumbles Moor to near Hawksworth, descending from 1175 ft to 575 ft and marking the upper limit of drift on the south side of Wharfedale. Penny (1974) mentions the Arthington Moraines in Wharfedale and the Apperley Bridge Moraine in Airdedale but the relationships to Jowett and Muff's moraines in the area are uncertain. A map of moraines for the region was provided by Edwards (1938) who also mapped recessional moraines in the Nidd valley which were thought to be associated with the York–Escrick stage in the Vale of York. Raistrick (1926) mapped details of moraines and lateral meltwater channels in Wensleydale, Swaledale and the lowland to the east. He followed this with a map of inset lateral meltwater channels for Wharfedale and maps of moraines in Swaledale, Wensleydale, Wharfedale, Airedale and Ribbledale (Raistrick, 1927, 1931a, 1933), which appear to document sequential glacial recession up those valleys and include some of the moraines documented by Jowett and Muff (1904). Correlation of moraines between valleys was proposed by Raistrick (1927, 1932) although this was not based upon any firm dating evidence. Specifically, Raistrick recognized the following moraines in order of up valley succession: (a) the Airedale Moraines at Tong Park, Hirst Wood, Bingley, Utley and Cononley; (b) the Wharfedale moraines at Pool, Burley, Middleton, Drebley, Kilnsey and Skirfare Bridge; (c) the Nidderdale moraines at Darley, Glasshouses, Pateley and Gowthwait; (d) the Wensleydale moraines at Mickley, Thirn, Redmire, Aysgarth and Hawes; and (e) the Swaledale moraines at Ellerton, Grinton, Fleetham and Gunnerside. Behind these moraines Raistrick documented lake flats thought to be the locations of moraine-dammed lakes after ice recession. These moraines were summarized by Palmer (1967) who accepted Raistrick's maps and correlation of major end moraines in the dales converging on the Leeds region. Edwards (1938) and Edwards et al. (1950) proposed a Main Dales Glaciation based upon the dales moraines and subdivided it into an Early Main Dales Stage and a York–Escrick Stage, implying that features outside the York–Escrick moraine were older than the late Devensian (Dimlington Stadial). In the Barnard Castle/Stainmore area, Mills and Hull (1976) reported scattered morainic drift comprising "ablation moraine", "ice-margin moraine" and kame moraine (mapped as morainic drift) which may represent glacier margins receding into the dales. Additionally they identified a large "belt of hummocky topography" north of Darlington. It is generally accepted that the LGM glaciers occupying the dales south of Swaledale coalesced with the Vale of York lobe. This is with the exception of Airedale where a moraine near Rawdon has been used as the maximum limit (Catt 1977, 1987), although no explanation of this choice is apparent in the literature.

Within the northern Pennines there is little recorded in the literature regarding recessional moraines in contrast to the Yorkshire Dales (cf. Raistrick, 1931b; Douglas, 1991). In Teesdale a supposed lateral moraine of the former Teesdale Glacier stretches from Cronkley Scar to Holwick (Dwerryhouse, 1902; Francis, 1970). In County Durham, emphasis has been placed on the sequence of large glacial lakes that formed during deglaciation (Teasdale and Hughes, 1999). Much of the lowlands north of the Tees are underlain by lacustrine sequences. As the different ice masses divided during deglaciation, a large lake formed between them. Glacial Lake Wear (Raistrick, 1931b) is demarcated by the “Tyne–Wear Complex” of Smith (1994), which extends up to 132 m OD although the most extensive phase of the lake has been mapped at 43 m when the valley of Tunstall Hope was cut as a spillway. The Lake Edder Acres sediments are bounded to the east by the Easington– Elwick moraine, north–northwesterly aligned ridges that grade eastwards into hummocky moraine and kame and kettle topography (Francis et al., 1963; Smith and Francis, 1967; Evans, 1999).

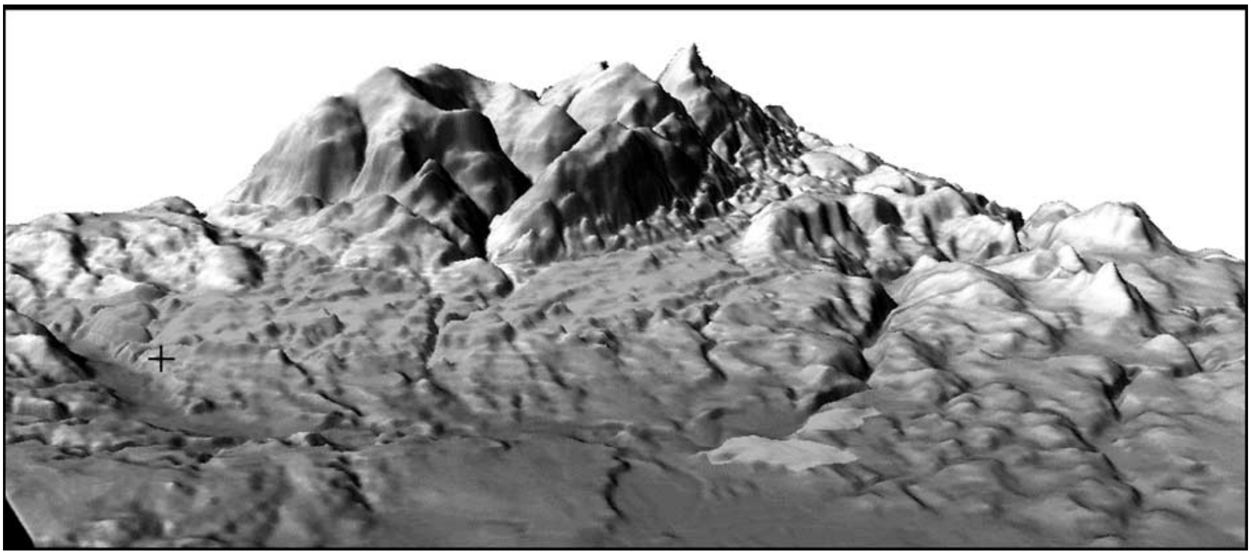

Fig. 6. Perspective view (looking NE) of the topography to the east of Carlisle looking towards Cold Fell, Cumbria. The extensive ice-marginal kame complex around the Brampton area (Brampton marked with a cross) can be seen as the hummocky ground in image centre at the base of the hill. Image is ca. 20 km across in the foreground and is derived from a 50-m grid-sized DEM. n Crown Copyright Ordnance Survey. An EDINA Digimap/JISC supplied service.

2.2.4. Northwest England

In the Edenside/Solway area an extensive marginal kame complex located south of Brampton (Fig. 6) was mapped by Trotter (1929), who interpreted the feature as the marginal deposit of the downwasting Eden Valley glacier. Similarly, Burgess and Holliday (1979) identified morainic drift that separated the till of the Vale of Eden from the bedrock of the Pennine escarpment, which they suggested represents the lateral moraine remnants of the Vale of Eden Glacier. Trotter (1929) also mapped “outwash moraines” or large ice marginal deposits west of Annan and north of Gretna. Smaller accumulations mark recessional positions to the south around Carlisle but Trotter’s interpreted recession pattern seems unlikely as the ice is more likely to have receded up the Solway, almost opposite to his direction. Marginal meltwater channels mapped in the area south of Wigton by Eastwood et al. (1968) document the downwasting of the Solway glacier on the northern edge of the Lake District. Arthurton and Wadge (1981) have mapped deeply incised and densely spaced meltwater channels in the Penrith area. They are best developed on the western slopes of the Pennines in the vicinity of Melmerby Fell and stretch out across the Vale of Eden where they are associated with long discontinuous eskers. Hollingworth (1931) mapped a series of elongate sand and gravel accumulations, which he interpreted as icemarginal positions in the Eden valley (west side of the river), although these could easily be subglacial. He further mapped lobes of ice receding from the east Lakeland fells and some lobes receding into the Lakes. On the northern margins of the Pennine massif, Trotter (1929) identified small recessional moraines in East and West Allendale documenting the retreat of Pennine upland ice.

Early mapping by Jowett and Charlesworth (1929) depicted ice margin recessional positions in the west Pennines based upon marginal meltwater channels. They also mapped lakes but these are not everywhere verified by deposits. Their easternmost glacier margin is marked by the upper limit of "extraneous drift". To the north, Jowett (1914) had earlier mapped ice recession again using meltwater channels although most of these were referred to as lake spillways rather than lateral features. Local depositional forms identified by Jowett and Charlesworth (1929) included kame–moraine belts, specifically an example at Whaley Bridge where an ice lobe occupied the Goyt Valley. This feature was also mentioned but not mapped by Stevenson and Gaunt (1971), who additionally document an ice-contact delta along the same ice margin at Buxworth. They go on to suggest that the Rowarth valley was blocked to produce a lake with shorelines during ice recession from the area, but again no maps were produced. Around Stockport, the eastern margin of former glacier ice was partially mapped by Taylor et al. (1963) using a drift limit at heights of at least 1400 ft on the southern Pennine foot slopes. They also mapped recessional margins based upon pockets of glacilacustrine sediments and meltwater channels. The largest lake produced at this time was Lake Tytherington, documented by locally widespread current-bedded sand (classified as "Middle Sands"). Taylor et al. (1963) refer to an Irish Sea Readvance but there is no morphological evidence of this, only ice marginal glacifluvial deposits documenting ice retreat northwards. Johnson (1965a,b) mapped the upper limit of glacial erratics on the west Pennine slopes in addition to a lower altitude limit of "boulder clay" (ages unknown but presumably last glaciation). Johnson's mapping also includes Lake Tytherington and numerous meltwater channels (presumably mostly lateral), which document the recession of ice in this region and in the Rossendales.

In their recent summary, Aitkenhead et al. (1991) mapped some major moraine systems in Lancashire, namely the Eagland Hill Ridge, the Skitham Trashy Hill Ridge and the Kirkham Ridge. The latter feature was referred to as the "Kirkham End Moraine" by Gresswell (1967), who noted that the feature merged westward into a wide belt of hummocky moraine (Fig. 7). Longworth (1985) reported extensive exposures in the Kirkham Moraine to be complex interdigitated tills and stratified sediments. He linked the sedimentary structures to supraglacial processes and stagnation. A lateral moraine was depicted by Gresswell (1967) along the western flank of the Bowland Fells, a feature that merges into the hummocky moraine at the east end of the Kirkham Moraine. This presumably demarcates the margins of a lobe sitting over the Fylde. A former ice-marginal assemblage of “moulin kames” mapped by Brandon et al. (1998) in the River Wyre valley possibly documents the recession of this lobe.

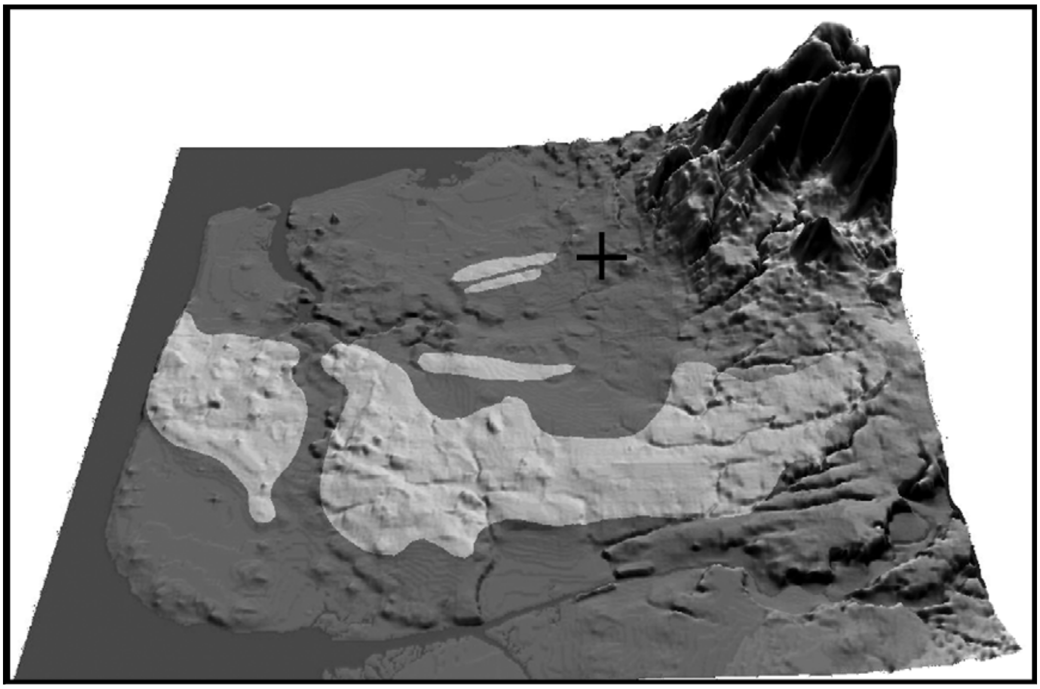

Fig. 7. Perspective view (looking north) of the topography of the Garstang area (Garstang is marked with a cross), Lancashire showing the "Kirkham End Moraine" and associated wide belt of hummocky moraine as mapped by various researchers and depicted on the Glacial Map of Britain (Clark et al., 2004). Arcuate ridges lying outside the area mapped as moraine could also be glacial depositional features. Image is ca. 30 km across in the foreground and is derived from a 50-m grid-sized DEM. n Crown Copyright Ordnance Survey. An EDINA Digimap/JISC supplied service.

Early work in West Cumbria and the Lake District by Trotter (1929) identified a series of readvances. Glacier margins were mapped in the area by Smith (1932) using marginal meltwater channels. It has long been recognized that ice-dammed lakes occupied the coastal lowlands between the Scottish/Irish Sea ice and the Cumbrian highlands, producing lake sediments and deltas (Eastwood et al., 1931; NIREX, 1997b; Akhurst et al., 1997; Merritt and Auton, 2000). These lakes were originally identified by Kendall (1924) and Dixon (1922) but the sediments were also interpreted as marine deposits by Kendall (1881), De Rance (1871) and Reade (1900). The upper limit of glaciation during the LGM is marked by palaeonunataks (Lamb and Ballantyne, 1998) in the central Lake District. During the deglaciation of west Cumbria and the Lake District, glacifluvial outwash fans were deposited in the valley bottoms around Whitehaven and St Bees. These were later glacitectonized and overlaid by red diamictons or tills related to the readvance of Irish Sea ice. This “Gosforth Oscillation” has been dated to 17 ka BP by Merritt and Auton (2000). The glacitectonized outwash forms a belt that extends up to 5–10 km inland from the coast (Trotter et al., 1937; Huddart, 1991, 1994; NIREX, 1995), referred to as “glacially overridden terrain” (NIREX, 1993, 1995, 1997a,b), produced by a readvance of Scottish ice. A second readvance by Scottish ice, termed the “Scottish Readvance” and dated to 14 ka by (Merritt and Auton, 2000), produced the glacitectonic structures of the St Bees moraine (Trotter et al., 1937; Williams et al., 2001). This moraine is part of a narrow coastal strip referred to as "glacitectonic thrust terrain" by NIREX (1995). The very prominent Bride Moraine on the Isle of Man (Thomas, 1984b) has often been linked with the St Bees Moraine in previous reconstructions of a "Scottish Readvance" (e.g. Charlesworth, 1926a,b; Penny, 1964). Indeed the intense glacitectonic disturbance of the Shellag Formation on the Isle of Man (Slater, 1931) is interpreted as the product of a significant readvance by Irish Sea ice by Thomas (1976; Orrisdale Readvance). Ice contact glacilacustrine deltas and proglacial outwash mark the margin of the Scottish/Irish Sea ice along the Cumbrian coast during the Scottish Readvance (e.g. at Holme St Cuthbert at 30 m and at Harrington from 30 to 152 m) and document the large proglacial lakes Wasdale and Eskdale (Huddart, 1970, 1991, 1994; Huddart and Tooley, 1972). Moraines recording the margins of Lake District glaciers have also been documented. For example, at the western end of Wastwater a large moraine was mapped by Smith (1932) and later also recognised by Huddart (1991). Gresswell (1952) mapped end moraines in the SE Lake District that probably relate to ice recession back into the valleys. McDougall (1998) has mapped the recession of plateau outlet glaciers dating to Loch Lomond Stadial but has also identified more extensive ice in some valleys presumably dating to ice sheet deglaciation. Similarly, moraines at Rosthwaite and Thornythwaite mark the limits of valley glaciers in Borrowdale (Marr, 1916; Raistrick, 1925). These were regarded as evidence for readvances postdating the LGM by Hay (1944) and Wilson (1977), but Clark and Wilson (1994) suggest a more extensive Loch Lomond Stadial ice coverage to explain the moraines. The former margins of the receding Lonsdale Glacier were mapped by Moseley and Walker (1952) using lateral meltwater channels. This glacier originally flowed down the Lune Valley and around the northern edge of the Forest of Bowland.

2.2.5. North Wales

Glacial depositional landforms document the nature of glaciation of the Lleyn Peninsula, North Wales. A considerable thickness of glacial sediment contained in hummocky drift along the north coast, extending from Pen y Groes to Trevor and then from the steep cliffs of the highlands westwards (Lleyn Peninsula moraine belt or Clynnog Fawr/Bryncir moraine), was interpreted initially by Mitchell (1960) and Synge (1963a, 1964) as the southern limit of Irish Sea ice during the last glaciation. However work by Saunders (1968a), Simpkins (1968) and Whittow and Ball (1970) firmly established that Late Devensian tills exist outside this limit. This led to proposals for two glacier advances on the Lleyn (Whittow and Ball, 1970), the later advance being classified as a readvance by Saunders (1968b) and Bowen et al. (1986), a concept proposed earlier by Pocock et al. (1938) as their Scottish Readvance. Saunders (1968b) suggested that the upper till on north Lleyn coast recorded a readvance and the moraine deposited at the maximum of this readvance was at Bryncir, rejecting his earlier opinion that this moraine marked the maximum limit of Late Devensian glaciation. Whittow and Ball (1970) recognized an early phase when Irish Sea ice moved over the northern and western Lleyn and Welsh ice (Criccieth Advance) moved in from the NE and impinged on St

Tudwal's Peninsula. Their later phase was characterized by Irish ice (Main Anglesey Advance) moving from the NE to encroach upon the north coast up to the area of Cors Geirch, producing the Clynnog Fawr moraine. At the same time Welsh ice (Arvon Advance) moved from Snowdonia along the flanks of Yr Eifl towards Bodfean and from the Vale of Ffestiniog across Cardigan Bay to St Tudwals Peninsula and the east end of Porth Neigwl (much of the Lleyn was ice free during this later phase) and up to the Sarn Badrig moraine (see below). Bowen et al. (1986) later retained the Main Anglesey Advance as their Gwynedd Readvance.

McCarroll (1991) and Gibbons and McCarroll (1993) provide a summary map of the Clynnog Fawr "moraine" and suggest that it is a topographically induced stillstand position with active ice lying along the north coast and stagnant ice lying on the south Lleyn and so they question the status of the Gywnedd Readvance. Firm evidence for a readvance exists, however, at Dinas Dinlle (see below). Further depositional features that have initiated debate are the Cors Geirch sand and gravel terraces, interpreted as the deposits of glacial lake Bodfean by Matley (1936). They were later studied by Saunders (1968c) who identified three stages of development during deposition between Irish Sea ice on the north coast andWelsh ice on the south coast of the Lleyn. Terraces from 86 m downwards through 50 m, 45 m to 20 m and less were mapped by McCarroll (1995) and Young et al. (1997), who interpreted them as kame terraces deposited in water ponded between the Welsh and Irish ice in the area, thereby verifying the conclusions of Saunders (1968b) and Bowen et al. (1986).

Recent mapping of Quaternary geology and geomorphology on the eastern Lleyn by Thomas et al. (1990, 1998) has highlighted complex glacifluvial sequences and moraines with glacitectonic structures, suggestive of glacier halts or minor readvances. Specifically, they report a large outwash tract (the Rhoslan sandur) that was deposited between the Irish and Welsh ice as they decoupled over eastern Lleyn and morainic ridges that document ice marginal recession. During the later stages of this uncoupling at around 16 ka BP (just before 14,468 BP based on a radiocarbon date at Glanllynnau) Thomas et al. (1998) depict the ice lobes in Cardigan Bay with their margins located at the Sarn Badrig and Sarn-y-Bwlch moraines (see below). Prominent moraines in the Pant Glas area are composed of glacitectonized glacifluvial sediment (Pont y Felin, Bryncir/Llecheiddior, Derwyn Fawr and Graianog moraines) recording sequential halts/readvances by the ice as it retreated towards the watershed of the Peninsula. Summarily, recent research indicates that the glacial landforms of the Lleyn Peninsula represent features typical of the uncoupling of the Welsh and Irish ice (Young et al., 1997), interrupted by possible readvances or at least minor oscillations of Irish ice on the north coast; these were certainly necessary to produce the glacitectonic features reported by Thomas et al. (1990, 1998) and the thrust block of Dinas Dinlle. A readvance by Welsh ice appears to be recorded at the east end of Porth Neigwl where Irish Sea till has been thrust in an east–west direction over the "intermediate waterlain sequence" (Young et al., 1997). The latter records stagnation of the Irish Sea ice and ponding of water on its surface. Once the Irish Sea ice had retreated it allowed the Welsh ice to advance over the sediments.

On the west Wales coast, substantial moraines lie offshore in Cardigan Bay. Foster (1970b) and Garrard and Dobson (1974) identified the Sarn Badrig, Sarn-y- Bwlch and Sarn Cynfelin offshore moraines documenting Welsh glacier flow into Cardigan Bay as a series of lobes. Garrard and Dobson (1974) further identified the limit of Welsh ice based upon erratics that extend as far offshore as the westernmost limits of the moraines. Based upon its erratic content, Foster (1970b) suggested that Sarn Badrig was a medial moraine located between Irish Sea and Welsh ice but the glaciation of the Lleyn Peninsula east of St Tudwal's by Irish Sea ice has never been verified by till provenance in terrestrial outcrops. A terrestrial continuation of the Sarn Badrig moraine is reported by Allen and Jackson (1985), marking the coalescence zone of glaciers formerly occupying the Dwyryd and Ysgethin valleys. The upper limit of the last glaciation in nearby Snowdonia has been mapped by McCarroll and Ballantyne (2000) based upon a clear weathering trimline. They also propose radial flow from the ice located over the Snowdonia massif, which was deflected southwestwards in the north by Irish Sea ice flowing over Anglesey but which flowed westwards across St Tudwal's Peninsula, supporting Foster's (1970b) reconstruction and the striae measurements of Gibbons and McCarroll (1993). The later glacial history of the Snowdonia massif prior to the Loch Lomond Stadial/Younger Dryas is poorly known, many cirque moraines being equated with the Younger Dryas despite a general lack of morphostratigraphic control. One exception is in the Cwm Dwythwch cirque (north Snowdon massif) which contains a large moraine that is demonstrably older than the Younger Dryas (Seddon, 1957, 1962; Unwin, 1970, 1975).

On the north coast of Wales, Fishwick (1977) documents Irish Sea and Welsh tills in the lower Conway basin, indicating that the two ice masses were coalescent and the coalescence zone fluctuated during glaciation. He also maps the meltwater channels of the basin, suggesting that many are subglacial rather than marginal as was suggested by Embleton (1961) in his reconstruction of a small glacier in the valley. The maximum extent of Irish Sea ice is marked along the North Wales coast by the upper limit of Irish Sea till. Irish Sea ice penetrated up the Vale of Clwyd (e.g. Strahan, 1886) as recorded by Irish Sea drift up to 300 m on Halkyn Mountain and up to 300 m on Gloppa Hill, Oswestry (Wedd et al., 1929) and up to 426 m at Moel Tryfan in Snowdonia (compilation map produced by Embleton, 1964a). Irish Sea ice also deposited till on Ruabon Mountain (Wedd et al., 1928) and Selattyn Hill (Wedd et al., 1929). Ice moved to the limit marked by the Bodfari–Trefnant Moraine (Rowlands, 1955) after 18 ka (see Tremeirchion cave date below). This moraine and the Tremeirchion site have been used by Bowen (1974) to suggest a readvance coeval with the Lleyn Peninsula (Gwynedd Readvance) event. Deltas at 67 m record the existence of glacial Lake Clwyd when glacier ice stood at the Bodfari–Trefnant Moraine (Rowlands, 1955). Glacial depositional landforms record Irish Sea ice recession from the Clwydian Range (e.g. Brown and Cooke, 1977; Embleton, 1957, 1961, 1964 a,b,c). The glacial features of the Wheeler Valley have received attention from various researchers, early reconstructions indicating the presence of a glacial lake (Embleton, 1957; Peake, 1961). This was rejected by Derbyshire (1962) and Brown and Cooke (1977), who suggested that the drift mounds and ridges of the valley represented crevasse fills and kames, although the features that Brown and Cooke (1977) record as "drift ridges" are morainic in appearance in that they are isolated and occupy valley sides or trend diagonally across valley bottoms. A glacilacustrine origin of sediments and landforms at Rhosesmor has been proposed more recently by Thomas (1984a,b,c, 1985) who suggests that ice receding from the east end of the Wheeler Valley dammed the drainage to produce a small proglacial lake (Rhydymwyn Lake) into which two overlapping deltas were prograded at water levels at altitudes of 180 m and 192 m.

2.2.6. South Wales

The first regional map of glacier coverage in south and central Wales was that of Charlesworth (1929), whose South Wales end moraine (SWEM) was demarcated by what he interpreted as the limit of the "newer drift". Numerous revisions of the SWEM have followed on from his work (see Bowen, 1981; Campbell and Bowen, 1989 for a review) but large parts of the former glacier margin represented in these revisions are poorly constrained by any glacial geomorphology. An ice lobe in Swansea Bay was fed by ice moving down the Nedd, Tawe and Afan valleys. A readvance moraine associated with the lobe occurs in the Tawe valley at Glais (Strahan, 1907; Trueman, 1924; Jones, 1942; Anderson, 1977). The Glais Moraine possesses a hummocky surface and contains glacitectonic structures. The latter may explain the description by Strahan (1907) of transverse ridges running across the moraine, products of proglacial glacitectonic compression. Downstream from the Glais Moraine lies the Glandwr Moraine (Bowen, 1970), recording early recession from the Swansea area. Immediately to the west, located on the upper shoulders of the lower Tawe valley, lie the Tirdonkin and Pant-lasau moraines (Bowen, 1970). To the northwest of the Gower/Swansea region moraines document glacier recession/readvance in a number of valley systems, for example in the Loughor valley near Pontarddulais (Waun Gron Moraine), on the north shore of the Loughor estuary near Llanelli (Machynys Moraine), in the Gwendraeth Fawr valley (Pont-newydd and Ponthenry moraines), in the Gwendraeth Fach valley (Llandyfaelog and Pontantwn moraines) and in the Afon Cywyn valley (Sarnau Moraine) near St Clears (Bowen, 1970). The Nedd, Tawe and Afan valleys all contain evidence of glacier readvances or stillstands during overall recession of the ice sheet in the form of lacustrine sediments, kame terraces and moraines (Charlesworth, 1929; Groom, 1971; Hughes, 1974; Anderson and Owen, 1979; Culver and Bull, 1979), the largest ice marginal features in the Nedd being the Tonna and Clyne moraines of Anderson and Owen (1979; see also Aberdulais moraine of Bowen, 1970). Substantial valley moraines are reported by Bowen (1970) in the Ely River valley (Talbot Green Moraine), in Taff Vale (Treforest Moraine) and the Rhymney Valley (Bedwas Moraine). In the Vale of Glamorgan the limit of the last glaciation is marked by hummocky moraine/kame and kettle topography around Pyle, Margam (Margam- Pyle Moraine) and Llanilid (Bowen, 1970). A large area of kame and kettle topography located between Pencoed and Newport marks the ice sheet margin fed by glaciers from the South Wales Coalfield dValleysT ice (Charlesworth's, 1929 "Glamorgan Piedmont Glacier"). This feature was later confirmed by Bowen (1970) as the ice sheet margin between Pencoed and Cardiff, but regarded by him as a recessional position between Cardiff and Newport based upon reconstructions of an ice lobe flowing offshore in that area, fed by the coalesced dValleyT and eastern coalfield glaciers (Welch and Trotter, 1961).

In Pembrokeshire Charlesworth's SWEM limit was drawn along the north Pembroke coast, based upon the Gwaun–Jordanston marginal meltwater and glacial lake overflow channels and the glacifluvial/glacilacustrine sediment bodies at Banc-y-Warren and Tregaron. The Gwaun–Jordanston channels were later re-interpreted as subglacial in origin by Bowen and Gregory (1965) and Bowen (1967), thereby suggesting an ice limit further to the south and questioning the existence of lakes in the area. However, the sediments and landforms at Banc-y-Warren and Tregaron, together with evidence of former lacustrine environments at other sites, provide clear documentation of glacifluvial and glacilacustrine environments during glacier recession, as would be expected if glacier ice uncovered such a large system of deep subglacial channels. Watson (1970) used pingo distribution to demarcate the margins of the last ice sheet in southwest Wales but this has been questioned via various alternative interpretations (e.g. Handa and Moore, 1976; Bowen, 1973a,b, 1974). A long history of study of the Banc-y-Warren succession has resulted in a thorough documentation of the internal structures and sedimentology. Helm and Roberts (1975) suggested a deltaic origin, explaining the large scale faulting in the sequence as a result of subaqueous mass failure. This supported the contention of Jones (1965) that the lower Teifi Valley had hosted a lake during Late Devensian glaciation. Whilst accepting that some sediments at Banc-y-Warren were deltaic, Allen (1982), and later Worsley (1984), suggested that the majority of the material was glacifluvial outwash and that a small supraglacial lake was formed at the same location some time during deglaciation. Although Banc-y-Warren has been cited as a possible Gilbert-type raised marine delta by Eyles and McCabe (1989), its location on an interfluve high above the Teifi Valley mouth clearly requires Irish Sea glacier ice impinging on the coast. Moreover, the 140 m altitude of the feature is far in excess of predicted marine limit altitudes for the Irish Sea basin (McCarroll, 2001). Another glacifluvial feature to the south of Banc-y-Warren at Trefigin was one of a number of features regarded as eskers by Bowen (1982a,b; Monington esker) but which has recently been interpreted as an erosional remnant of a former outwash plain or kame terrace associated with the Irish Sea glacier by Owen (1997). Crimes et al. (1994) mapped a series of kame terraces, eskers, kames, meltwater channels and ice-contact proglacial outwash (including the Monington Esker), that in some places grade into features and sediments associated with a deglacial Lake Teifi such as fan deltas and lake sediments. A linear suite of these depositional features is used by Crimes et al. (1994) to reconstruct the Irish Sea ice margin at approximately 14,500 years BP. Prior to this date and the development of Lake Teifi, Irish Sea ice penetrated as far inland as Cilgerran where a line of kame mounds records the ice margin (Crimes et al., 1994). Lake Teifi has been resurrected also by Fletcher and Siddle (1998) who suggest an upper level of up to 200 m based upon borehole stratigraphies in the area around Llandudoch. Lake sediments below the Monington "esker" outwash may equate with the laminated sediments observed in boreholes by Fletcher and Siddle and therefore record an early (Irish Sea ice advance phase) lake in the Teifi valley. This suggests that the Banc-y-Warren and Monington outwash prograded into the lake and were dissected when lake levels fell. Further up the Teifi Valley, an extensive area of morainic topography extends for some distance south of Tregaron, prompting Charlesworth (1929) to propose that this feature was part of his South Wales End Moraine. Small quarry exposures located immediately north of Lampeter have subsequently revealed delta foresets severely disturbed by glacitectonic overriding (Davies et al., 1997), indicating an ice margin located south of the Tregaron Moraine. Charlesworth (1929) identified these stratified sediments assuming them to be deltas prograding into Glacial Lake Teifi, but reported no evidence of glacitectonic disturbance.

In the Usk valley Dimlington Stadial tills have been reported to overlie buried valley deposits by Williams (1968a). Thomas (1997), following on from initial work by Williams (1968b), has mapped a complex moraine belt lying between Usk and Crickhowell. Individual "kame moraine" ridges are used by Thomas (1997) to reconstruct at least nine glacier margins associated with a piedmont lobe receding up the Usk valley. He suggests that this glacier lobe was coeval with the glaciers that formed the Wye Valley End Moraine at Hereford (Luckman, 1970) and the Llanfihangel–Crucorney Moraine (Lewis, 1970a). These moraine systems traditionally have been used in reconstructions of the Devensian limit in the region (e.g. Charlesworth, 1929; Bowen, 1981). The Llanfihangel– Crucorney Moraine is a large expanse of hummocky terrain (Strahan and Gibson, 1900; Grindley, 1905; Charlesworth, 1929; Lewis, 1970a) thought to represent the maximum of the Dimlington ice sheet in the area. The moraine was deposited by an offshoot of the Usk piedmont lobe that flowed due east near Crickhowell, spilling into the Vale of Ewyas. Relating to the recession of this offshoot, Lewis (1970a) identified three moraines recording retreat of the ice back into the main Usk valley.

Further moraines in the Usk basin provide evidence of glacier recession into the Welsh mountains. Lewis (1966, 1970b) identified a moraine at Llandetty near Talybont which he regarded as a readvance position, part of his "Mountain Readvance". The Hay Moraine (see below) was regarded as contemporaneous with the Llandetty Moraine by Lewis (1970a), both moraines representing the margin of Wye/Llyfni ice after the recession of the Usk piedmont lobe. Large terminal and lateral moraines at Penoyre/Cradoc and Llanfrynach/Groesffordd were regarded by Lewis (1970a) as the products of Wye/Llynfi ice receding eastwards after penetrating as far west as Aberyscir. The Wye/Llyfni ice came into contact with the receding

Usk glacier at Aberyscir as documented by a large interlobate moraine dissected by the Bran River. Continued recession of the Wye/Llyfni ice from the Usk basin is recorded by four moraine stages, mapped by Lewis (1970a) in the Llyfni basin. The outermost two moraines are associated with an icedammed lake named Glacial Lake Llangors. Once the ice had receded to the confines of the Wye valley it produced another large moraine at Llyswen, a large kame and kettle feature. Further recession was punctuated by the deposition of the Llanelwedd/Alltmawr Moraine, the Newbridge Moraine and a moraine upstream of Llanwrthwl.

Moraines around Rhayader document the splitting of the Wye glacier into three sub lobes, namely the Wye, Elan and Glan-Llyn glaciers. In the same area Lewis (1970b) reports that the Wye glacier poured off the central Wales plateau into the valley below Llangurig, forcing its way up Marteg valley into the St Harmon basin. Moraines document this invasion between St Harmon village and the Wye. Much larger moraines exist in other east–west aligned valleys east of the Wye, a spectacular example being the "kame moraine" near Nantmel in the Dulas valley. In the north Brecon Beacons Ellis-Gruffydd (1972, 1977) used striae and erratics to suggest that a Breconshire Ice Cap submerged the region and debouched ice into the Usk Valley to the north, westwards to join mid Wales ice flowing down the Tywi Valley and south across the South Wales coalfield. Ellis-Gruffydd also mapped a sequence of ten end moraines documenting glacier recession in the upper Usk basin.

2.2.7. Welsh/English borders and Cheshire–Shropshire lowlands

In the Welsh/English Borderlands, reconstructions of glacial history have centred on interpretations of the stratigraphy of the region. Wedd et al. (1928) subdivided the Quaternary sediments in the area into a tripartite sequence including "Lower Boulder Clay", "Middle Sands" and "Upper Boulder Clay" (see below). Central to reconstructions of the ice sheet margins in the area is the Wrexham "delta–terrace", thought to have been deposited by Welsh Ice as it retreated, the meltwater feeding a delta into a lake impounded by Irish Sea ice (e.g. Peake, 1961). Peake argued that the Wrexham terrace was fed by meltwater draining down the Alyn Valley from the north. Poole and Whiteman (1961) suggested that the terrace was not a delta but part of an end moraine system extending from Wrexham to Ellesmere. Specifically, they suggested that the Middle Sands record the recession of the ice responsible for the deposition of the Lower Boulder Clay and that the Upper Boulder Clay recorded a later advance by the Irish Sea ice right up to the base of the Welsh upland massif.

During recession the Wrexham "delta terrace" was deposited as an ice contact sandur plain by meltwater flowing westwards; therefore the outer escarpment was seen as an ice-contact slope rather than a delta front. This reconstruction was supported by Worsley (1970). Francis (1978) suggested that the "terrace" was a subaerial fan deposited on dead ice, explaining the well developed kettle hole topography at Vicarage Moss (supported by Dunkley, 1981 and Wilson et al., 1982). Thomas (1985) suggests a diachronous and complex deposition history, involving an oscillating Irish Sea ice margin characterized by ice-contact outwash fans and short-lived lake basins. Thomas (1984a) supported Peake's (1961) earlier proposal of lakes dammed between the Irish Sea ice and the Welsh massif but in much smaller proportions (seven lake phases were suggested by Peake), thereby verifying that her spillway channels were probably subglacial as reassessed by Derbyshire (1962) and Embleton (1964b). According to Thomas (1985), who bases his reconstructions on a number of quarry sections in the narrow corridor of Irish and Welsh ice coalescence over the Alyn Valley, deposition of the Wrexham terrace was over considerable volumes of stagnant ice thereby explaining the pitted outwash surface to the northeast of Wrexham. The terrace was fed by outwash flowing over a linear sandur plain located at first between receding Irish and Welsh ice and then between Welsh ice and the highlands of the Welsh massif (e.g. Halkyn, Hope and Ruabon mountains). The complex nature of the depositional environment indicates that the terrace is neither an alluvial fan nor a delta, but an accumulation of sediment fed by icefront debris flows and alluvial fans, sandur, and proglacial and ice-contact lakes. To the south and east of Wrexham the terrace sediments are underlain by lake deposits which record the damming of the meltwater during early stages of ice recession. Upward-coarsening in the sedimentary succession above the lake deposits records sandur progradation.

Prominent moraines in the Cheshire–Shropshire lowlands have been the subject of considerable speculation on ice marginal fluctuations in the region. The largest of these moraines forms an arcuate multiridged feature (Fig. 8) variously named the "Wrexham-Ellesmere-Wem-Whitchurch-Woore Moraine" (e.g. Lewis, 1894; Pocock and Wray, 1925; Peake, 1961; Poole and Whiteman, 1961; Yates and Moseley, 1967), the "Middle Sands Moraine" by Poole and Whiteman (1961) and "Bar Hill-Whitchurch-Wrexham Moraine" by Boulton and Worsley (1965). Thomas (1989) mapped the moraines in the area between Whitchurch and Shrewsbury and highlighted a series of ice marginal positions with associated outwash tracts and proglacial lakes. The later stages of uncoupling of Irish Sea and Welsh ice led to the linkage of outwash tracts in the Alyn Valley (Thomas, 1985; see above) with those to the south of Ellesmere. The ice-dammed lakes, unlike the vast Lake Lapworth envisaged by early workers, were small and relatively short lived and included Lake Prees (dammed between the Irish Sea ice and the Triassic escarpment) and Lake Bangor, into which the outer edge of the Wrexham terrace was deposited. Throughout the uncoupling of the two ice masses the outwash was deposited around the margins of the prominent moraine ridges abandoned during earlier recession. Thomas (1989) identified five major ice limits during the recession from the Dimlington Stadial limit. The first is located at a moraine south of Shrewsbury. The second is marked by a moraine ridge and associated linear ice-marginal sandur between Ellesmere and Shrewsbury, documenting the early uncoupling of the Welsh and Irish Sea ice. The third is marked by a moraine lying on the edge of the Triassic escarpment and looping northwestwards to Ellesmere where the margin remained stable since the development of the second moraine. The fourth is the Whitchurch-Wem-Ellesmere-Wrexham moraine loop comprising multiple ridges between Whitchurch and Ellesmere.

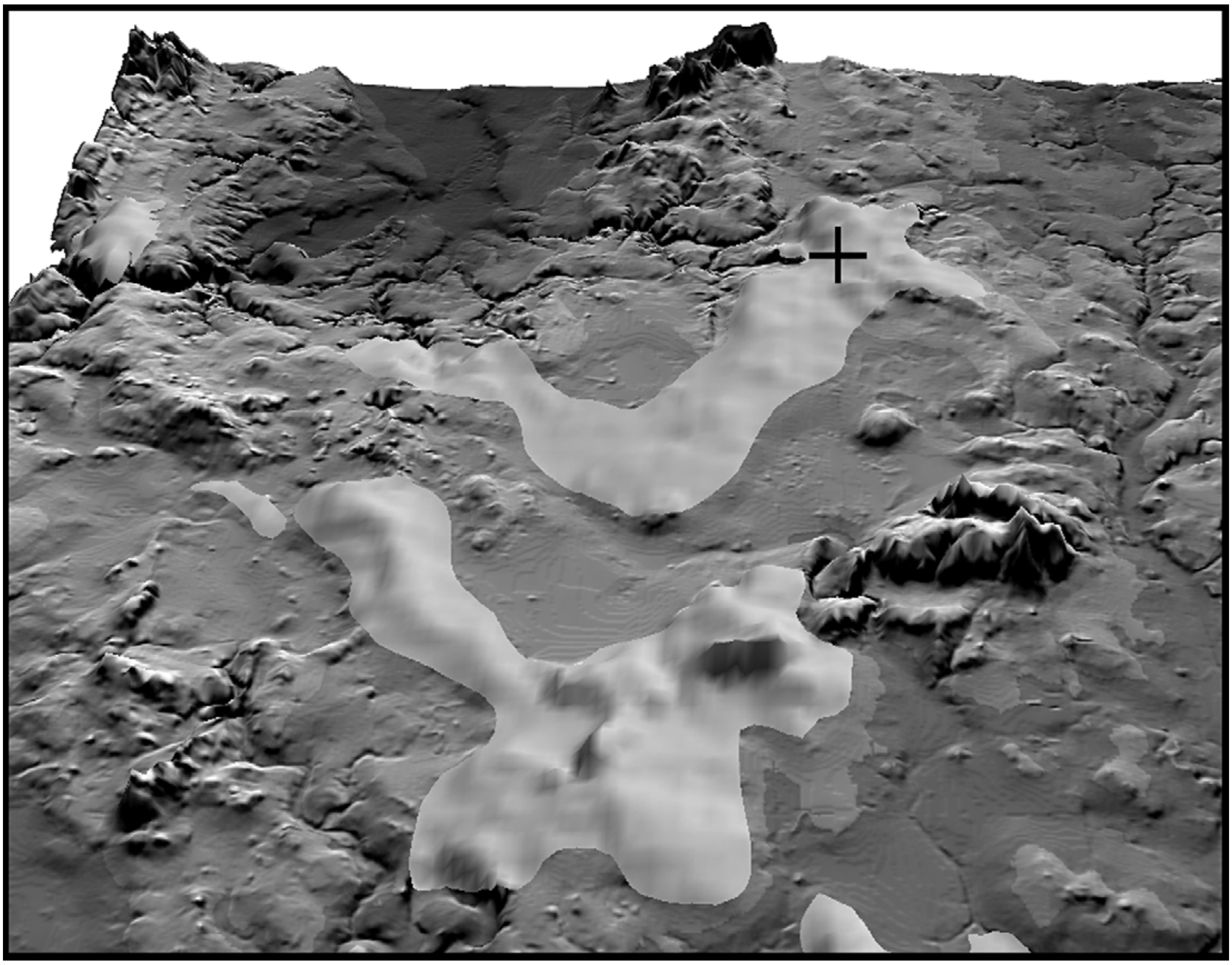

Fig. 8. Perspective view (looking north) of the topography of the Cheshire–Shropshire lowlands illustrating prominent moraines in the area. The largest forms an arcuate multi-ridged feature variously named the "Wrexham-Ellesmere-Wem-Whitchurch-Woore moraine" (moraine marked as mapped by various researchers and depicted on the Glacial Map of Britain; Clark et al., 2004). Cross marks the location of Whitchurch. Image is ca. 30 km across in the foreground and is derived from a 50-m grid-sized DEM. © Crown Copyright Ordnance Survey. An EDINA Digimap/ JISC supplied service.

A large reentrant, centred over Whitchurch, had developed in the ice margin by this time, from which the moraine curved southeastwards as the Woore moraine. As the ice had pulled back from the Triassic escarpment, the proglacial Lake Prees developed in the terrain between ice and upland. At the same time the Wrexham terrace was prograded along the west margin of the Irish Sea ice. The fifth lies to the north of Whitchurch and forms the margin of two lobes that had become separated as the Mid Cheshire Ridge emerged from the receding glacier surface. The reentrant in the ice margin had deepened as a result of the emergence of the ridge. At this time the proglacial Lake Bangor developed along part of the margin of the westernmost lobe, into which the outer edge of the Wrexham terrace was prograded. The presence of further moraines between the fourth and fifth major moraines attests to minor oscillations of the glacier margin in the Ellesmere and Whitchurch areas. Throughout the recession of the Irish Sea ice, proglacial outwash tracts were developed by way of the partial erosion and burying of earlier moraines and outwash fans. Rees and Wilson (1998) provide a map of the major moraines on the east side of the Cheshire Plain including: the Woore moraine (Lewis, 1894; Wedd, in Gibson, 1925; Jowett and Charlesworth, 1929; Poole and Whiteman, 1961; Boulton and Worsley, 1965; Yates and Moseley, 1967; Gemmell and George, 1972), thought by some researchers to be a readvance moraine; the Wrinehill moraine (Yates and Moseley, 1958), a hummocky moraine thought to represent ice stagnation; the Foxley moraine (Wedd, in Gibson, 1925), comprising glacitectonically disturbed stratified sediments and therefore interpreted as a readvance limit; and the Sheepwash moraine near Caverswall.

Three stages of ice recession have been mapped in the Madeley area by Yates and Moseley (1967), comprising the Tern Lake, Woore moraine/Lake Madeley and Wrinehill moraine/Lake Craddocks phases. The development of lakes dammed between the ice sheet margin and the west Pennine slopes is a subject of long standing debate centred specifically on the evolution of supposed spillway channels (cf. Kendall, 1902; Sissons, 1960, 1961a,b), in this case the “Stoke Series” of channels (Jowett and Charlesworth, 1929; Yates, 1955; Yates and Moseley, 1958; Walton, 1964; Johnson, 1965a,b; Knowles, 1985). More recently the Stoke Series of channels were considered by Knowles (1985) to be polygenetic, whereby Tertiary drainage systems were incised first by proglacial meltwater during ice advance against the west Pennine slopes and then by subglacial meltwater during the later stages of glaciation. Any proglacial lakes produced during ice advance and recession would have been small, shallow and short-lived, and so earlier models of spillway cutting by proglacial lake water spilling over the Pennine watershed northwest of Stoke have been largely superceded. Earp and Taylor (1986) report terminal moraines in the Burwardsley– Peckforton–Beeston area, marked on geology maps as an area of morainic drift. The moraines (originally identified by Poole and Whiteman, 1961) are named the Burwardsley Hill, the Burwardsley- Peckforton Castle, the Peckforton Hall and the Beeston Castle moraines. Earp and Taylor (1986) also refer to a major stillstand limit from Kingsley through Norley to Cuddington, although the landforms/ sediments used in the identification of this line are not specified.

Further south, outlet glaciers from the Welsh uplands formed piedmont lobes that partially covered hill masses such as the Stiperstones in Shropshire, which were thought to be a nunatak (Rowlands and Shotton, 1971; Goudie and Piggott, 1981). Large moraine spreads were reported by Charlesworth (1929) in the valleys immediately to the west of Craven Arms and to the east of Lydham. The outermost limit of glaciation was traced in this area and to the north by Dwerryhouse and Miller (1927). In the Church Stretton area Greig et al. (1968) identify two moraines, the outermost one at Brockhurst and a recessional moraine at Botvyle. Additionally the outermost limit of Devensian glacier ice in the Cardington area is mapped as an area of morainic drift on the BGS Drift Geology Sheet 166. Greig et al. (1968) proposed that the Welsh and Irish Sea ice was coalescent in the area south of Church Stretton but Welsh ice moraines are mapped only at Myndmill

Bridge and Clunbury (Oaker Moraine). Following on from the report of a moraine in Ludlow by Pocock (1925), Cross (1966, 1968), Hodgson (1972) and Cross and Hodgson (1975) documented evidence for glacier ice up to 243 m thick and deglacial lakes in the Wigmore and Presteigne basins. They also mapped a more extensive limit for the last Wye glacier in the Orleton area than that proposed by Luckman (1970; see below) based upon a more recent estimate of the age of till patches located outside the Orleton moraine. Additionally, they failed to locate any Wye glacier till east of Bromyard, thereby questioning Luckman's (1970) acceptance of Grindley's (1937) identification of till at Ankerdine Hill. The distribution of till is employed in addition to major moraines to demarcate the last ice sheet margins in the Teme basin (Cross and Hodgson, 1975).

Ice of Welsh origin advanced from the northwest and extended to Ludlow and Stanton Lacy. Two tills in the area, a lower reddish brown local till overlain by a greyish brown Welsh till, record ice advance from two different source areas. The Welsh ice retreated from the Teme basin before the Wye glacier receded from the Orleton area, based upon the observation that meltwater from Irish Sea ice to the north carried distinctive Irish Sea erratics from the Church Stretton lobe through the Onny Gap and the Whitcliffe gorge to be deposited as deltas in glacial Lake Woofferton. The highest stand of Lake Woofferton is marked by deltas at 91 m. The Herefordshire end moraine of Luckman (1970) marks the limit of the Wye valley glacier around the Hereford lowlands. The Kington-Orleton kettle moraine (Dwerryhouse and Miller, 1930; Luckman, 1970; Cross and Hodgson, 1975) lies at the foot of the Silurian dipslope along the northern margin of the lowlands. The moraine continues southward from Orleton to Dinmore Hill as a linear series of gravel mounds and benches. Both Charlesworth (1929) and Luckman (1970) suggest that the hill masses in the Dinmore Hill area forced the glacier margin to split into two lobes, the southernmost of which deposited the continuation of the Herefordshire end moraine in the Hereford area. Charlesworth (1929) maps a moraine loop that encloses the village of Wellington but this is not reproduced by Luckman (1970). According to Luckman (1970), the moraine continues southeastwards from Yarsop to Burghill and then as a wide arcuate belt from Swainshill/Credenhill across the Wye valley to Thruxton, but Brandon (1989) indicates that the ice margin was less lobate and the hill masses in the area were small nunataks. The moraine belt continues as a series of kames located in valleys at altitudes of up to 210 m along the Old Red Sandstone scarp face that constitutes the south side of the Wye valley. The continuation of the Wye glacier is marked from that point westwards by a drift limit on the lower slopes of the Black Mountains up to 360–390 m (Dwerryhouse and Miller, 1930; Grindley, 1954) and by "kame-like benches" on the north side of the valley (up to 274 m at Cwm-yr-Eithin; Luckman, 1970).

The Wye glacier was also confluent with ice in the Arrow valley. Ice pushed into the Dore Valley (Golden Valley) by flowing over the valley head col, as evidenced by considerable volumes of drift as far south as Ewyas Harold (Burnham, 1964). The most recent survey of Devensian till cover by the BGS (Brandon, 1989) agrees most closely with the mapping of Charlesworth (1929) but does not differ significantly from the map of Luckman (1970), especially in the Leominster/Orleton area. Furthermore, the BGS mapping did not continue into the high land on the south side of the Wye Valley and so Luckman's limits in the area are accepted by Brandon (1989). Lying inside the Herefordshire end moraine is the Staunton moraine (Aldis, 1905; Pocock, 1940; Luckman, 1970; Palmer, 1972), a continuous single ridge that crosses the Wye valley to join an area of well developed kame and kettle topography below the upper drift limit of the Wye glacier at Bredwardine. A further "large irregular drift mass" (Luckman, 1970) occurs at Hay where it has been named the Hay moraine since its description by Pocock (1925, 1940) and Dwerryhouse and Miller (1930). A narrow moraine belt wrapping around Burton Hill was considered to represent an embayment in the terminal moraine of the Wye Glacier but it is regarded as a recessional moraine by Brandon (1989) who links it to a similar moraine to the northwest of Stretton Sugwas.

The Smestow–Wolverhampton Line, demarcated by the limit of the Dimlington Stadial till (Wills, 1924, 1938; Whitehead et al., 1928; Mitchell, 1960; see below), has been traditionally regarded as the southernmost limit of the last glaciation in the West Midlands/Cheshire–Shropshire lowlands. North of this line, the Newport esker chain and the Penkridge esker were interpreted by Wills (1924, 1948) and Whitehead et al. (1928) as the deposits of two subglacial channels documenting recession of the last ice sheet margin north of Wolverhampton. Large individual mounds along the chain, particularly in the vicinity of Boscobel were interpreted as "esker fans" deposited at the glacier margin as it receded. Wills interpreted the larger accumulations as ice-contact deltas deposited into his Glacial Lake Newport once the ice had receded north of the Triassic escarpment. A subglacial channel occurs in association with the Penkridge esker and is interpreted as the product of the same stream that deposited the esker. Two glacier margins are recognized based upon (a) an outwash fan at the end of one section of the subglacial channel located southeast of Penkridge; and (b) a "kame" with a prominent ice contact face (presumably an icecontact fan) associated with the esker in Penkridge. Similar chains of glacifluvial mounds, crossing the cols in the higher terrain of the Shrewsbury district, were mapped by Pocock et al. (1938) and compared to the esker chains of the Newport area. They also identified large pockets of "terrace-like gravels" which they associated with Glacial Lake Lapworth. The name Glacial Lake Lapworth was proposed for the expansive area thought to have been submerged beneath ice-dammed lake waters during the recession of the Irish Sea ice from the Cheshire–Shropshire lowlands (Wills, 1924). The earliest phases were characterized by lake damming in the Severn basin (Lake Buildwas) and around Newport (Lake Newport), the spillways being the Ironbridge Gorge and Gnosall channel respectively (Wills, 1924, 1948; Poole and Whiteman, 1961; Gemmell and George, 1972). Although small, short-lived ice-dammed lakes probably occupied the freshly deglaciated terrain of the Shropshire lowlands, no sedimentological evidence exists for a lake as extensive as Lake

Lapworth. More recently the evolution of the Ironbridge Gorge has been linked, at least in its initial stages of downcutting, to subglacial meltwater erosion. This was followed by proglacial meltwater erosion and eventually the capture of the upper Severn drainage (Worsley, 1991). Nonetheless, Wills (1924) mapped small morainic ridges that document the early recessional margins of glacier ice in the Bridgnorth area.

2.3. Ice-marginal till stratigraphies and associated chronostratigraphic control

A reasonable but nonetheless modest number of radiometric and luminescence dates are available relating to the inception and decay of the British Ice Sheet. As these have been reported in overviews elsewhere (e.g. Jones and Keen, 1993; Bowen et al., 2002) we concentrate here on the chronological controls on former glacier margins and ice-marginal stratigraphies crucial to glacier reconstruction. Ages are given in uncalibrated radiocarbon years except where they were obtained by other dating methods.

2.3.1. North Sea offshore evidence

The offshore limit of the Dimlington Stadial glacier ice in the North Sea has been a contentious issue since seismic surveys and borehole information was first made available. Various models have been proposed based upon different combinations of morphological and stratigraphic evidence. The distribution of large incisions, interpreted as subglacial tunnel valleys, has been used by some researchers to demarcate the last ice limit (Valentin, 1957; Flinn, 1967; Jansen et al., 1979; Ehlers and Wingfield, 1991). However, both the age and the origin of the North Sea "tunnel valleys" pose considerable problems for ice sheet reconstructions. The subglacial interpretation of all incisions on the North Sea bed has been questioned by Long and Stoker (1986) who provide evidence for the subaerial erosion of some valleys. This questions the use of the distribution of North Sea incisions as an indicator of glacier extent. Given the problems associated with the North Sea "tunnel valleys", other workers have concentrated more on the glacial sediments and stratigraphy.

Chronostratigraphic control on the depositional sequence in the southern North Sea is provided by a seismic unconformity in middle Devensian sediments above which lies the Dogger Bank Formation (Cameron et al., 1992). This formation is of glacilacustrine and glacimarine origin and thought to record the advance of the British Ice Sheet during the late Devensian (Cameron et al., 1987; Long et al., 1988). Glacial deposition during the LGM is thought to be represented by the Cape Shore, Coal Pit and Bolders Bank Formations. Balson and Jeffrey (1991) and Cameron et al. (1992) report that the Bolders Bank Formation is lithologically similar to the onshore Skipsea Till of Holderness and therefore regard both as the same deposit. To the north, the Bolders Bank Formation continues as the Wee Bankie Formation (Gatliff et al., 1994). Towards the southern limit of the Bolders Bank Formation a system of "tunnel valleys" have been cut in a pattern that is radial towards the former glacier margin and are therefore interpreted as subglacial in origin. Stratigraphically and morphologically these channels can be distinguished from earlier tunnel valleys that lie to the south of the Bolders Bank Formation. More recently Carr (1999) and Carr et al. (2000) have proposed a more extensive offshore Dimlington Stadial limit based upon micromorphological assessments of core material from the North Sea. They suggest that the Dogger Bank and Bolders Bank Formations are at least in part subglacial in origin and therefore the easternmost limit of the British Ice Sheet was considerably more extensive than the limit of the Bolders Bank Formation. This supports the reconstruction of Ehlers and Wingfield (1991) for the southernmost part of the North Sea based upon the distribution of Dimlington Stadial tunnel valleys but questions their restricted ice limits proposed further north based on the Wee Bankie Formation (cf. Long et al., 1988).

The status of “moraineless Buchan” in northeast Scotland is significant in this respect (see above). A more extensive ice cover, at least in the northern part of North Sea basin, was proposed also by Sejrup et al. (1994) based upon the discovery of a Late Weichselian till dated to the period 22 ka–29 ka BP. At this time (the Cape Shore Episode) the British and Fennoscandinavian ice was coalescent and deposited the Dogger Bank/Jutland Peninsular “mega-moraine” of ice-marginal fans and glacitectonites. The two ice masses had separated by 20 ka BP but this was followed by a less extensive readvance (the Bolders Bank Episode) between 19 ka and 15 ka BP and correlated with the Dimlington Stadial by Sejrup et al. (1994). At this time the eastern margin of the British Ice Sheet advanced along the east Yorkshire coast and pierced the western edge of the Dogger Bank–Jutland mega-moraine, perhaps while undergoing a surge (S.J. Carr, personal communication). In the northern part of the North Sea basin the prominent Wee Bankie and Bosies Bank moraines mark former glacier margins beyond which the Marr Bank Formation was deposited (Holmes, 1977; Stoker et al., 1985). The Marr Bank Formation is a glacimarine deposit dated at 17.7 and 21.7 ka BP. Although the Wee Bankie and Bosies Bank moraines have been interpreted as the Late Devensian ice limits due to a lack of till on their eastern flanks (Holmes, 1977; Thomson and Eden, 1977; Stoker et al., 1985; Hall and Bent, 1990), recent discoveries of till in the North Sea basin have prompted their association with the Bolders Bank Episode (Carr, 1999). An independent ice cap flowed radially out from the Shetland Isles during the Bolders Bank Episode as recorded by the Otter Bank sequence. This followed on from the deposition of the Cape Shore Formation, deposited during the early part of the LGM when the British and Fennoscandinavian ice sheets were coalescent over the North Sea (see Holmes, 1997 for review). The diamictons of the Otter Bank sequence occur as submarine morainal banks on the shelf and are correlated seismostratigraphically with the Oxygen Isotope Stage 2 Tampen and Sperus formations to the east of Shetland (Johnson et al., 1993).

2.3.2. North Atlantic margin

At the northwest margins of the ice sheet, a maximum age for the LGM at Tolsta Head on the Isle of Lewis, Outer Hebrides is provided by a radiocarbon date of 27,333±240 years BP on interstadial organic lake detritus lying beneath Late Devensian till (von Weymarn and Edwards, 1973). A radiocarbon date of

22,480±300 years BP from glacimarine sediments deposited beyond the morainal banks on the shelf south of St Kilda provides a Late Devensian age for the glacier margin. Further dates from cores in the St Kilda basin indicate that it was deglaciated sometime after 15.2 ka BP even though glacimarine sedimentation continued until after 13.5 ka BP (Peacock et al., 1992). The bracketing of other major landform assemblages in Scotland has proved difficult in the absence of good chronostratigraphic control. An age bracket for the ice-marginal landforms of the lower Dee Valley (Brown, 1993) is provided by a radiocarbon date of 12.6 ka BP on basal peat near Braemar and a minimum age of 15,370 years BP for onshore recession of ice on the coast of Aberdeenshire (Hall and Jarvis, 1989). The maximum age of the Ardersier Readvance in the Moray Firth (Merritt et al., 1995) is provided by the dating of the Errol Beds that were glacitectonized during ice overriding. Gordon and Sutherland (1993) have bracketed the Errol Beds between 17 ka and 13 ka BP. Although an age bracket of 18 ka–13 ka BP was originally proposed for the Wester Ross Readvance by Robinson and Ballantyne (1979), more recent speculation has placed the age of the event at 13.5 ka–13 ka BP (Ballantyne et al., 1987; Ballantyne, 1993) at the time when the oceanic polar front migrated north of the Scottish west coast (Ruddiman and McIntyre, 1973). A basal peat date of 12,810±155 years BP provides a minimum age for the event (Kirk and Godwin, 1963).

2.3.3. South central Scotland

A deglacial date of 12,750±120 BP has been reported from the basal peat of a kettle hole in glacifluvial deposits that lie between Doune Lodge and the Loch Lomond Readvance at Callander (Lowe, 1978). This provides a minimum date for the recession of the Teith glacier from the area. The deglaciation of the Paisley-Renfrew area of the Glasgow region is reported to have been around 12.8 ka or slightly earlier according to radiocarbon dates on marine shells in the local Clyde Beds (Peacock, 1971; Browne et al., 1977; Sutherland, 1986).

2.3.4. Eastern England

A variety of evidence, poorly constrained by absolute dating, has been used to propose limits of an early Devensian glacial advance and two later readvances on Holderness (see also section 2.2.3), where the North Sea lobe of the British Ice Sheet moved onshore. The westernmost limit of the drift on Holderness has traditionally been used as the limit of a Devensian glaciation and an early Devensian age has been supported for the relatively more subdued drifts west of the River Hull (e.g. Straw, 1979). As outlined above, the readvances of Holderness and the east end of the Vale of Pickering have not been recognized by more recent researchers, because the limits were based largely upon the nebulous criteria of moraine freshness and lacked chronostratigraphic control. Moreover, the smoother topography of the tills in west Holderness is a result of Holocene flooding and draping of lower altitude hummocks rather than an age difference to the slightly more hummocky drift east of the River Hull. The westernmost limits of the North Sea ice in Holderness during the

LGM are presently drawn at the feather edge of the Skipsea Till (previously mapped as Hessle Till; e.g. Suggate and West, 1959). The oldest glacial deposit recognized in this area is the Basement Till, which is overlain by the Skipsea ("Drab") and Withernsea ("Purple") tills along an extensive stretch of the Holderness coastline (Madgett, 1975; Madgett and Catt, 1978). Debate continues as to whether the Basement Till is Dimlington or pre-Devensian in age (e.g. Lamplugh, 1890; Catt and Penny, 1966; Eyles et al., 1994). However, radiocarbon dates of 18,500±400 years BP and 18,240±250 years BP obtained by Penny et al. (1969) on organics between the Basement and Skipsea Tills provide a maximum age for the onset of the Dimlington Stadial in the region (Rose, 1985). An additional date for the onset of the stadial of 17,500±1600 (16,600±1700 BP corrected and in calendar years) was obtained by thermoluminescence techniques from beneath the Skipsea Till on the Wolds dip slope (Wintle and Catt, 1985). Peacock (1997) has argued that the Skipsea till represents a readvance by North Sea ice as late as 15–14 ka radiocarbon years BP. A radiocarbon date of 13,045±270 on organics from a kettle hole at Roos provides a minimum for deglaciation (Beckett, 1981). A radiocarbon date of 16,713±340 from basal peat in a kettle hole on glacial outwash in Kildale provides a minimum age on deglaciation of the northern slopes of the Cleveland Hills (Jones, 1977).

On the central Lincolnshire Wolds, a drift limit, marked by the edge of the "Lower Marsh Till", occurs at 114 m and has been equated with an Early Devensian glaciation by Straw (1957, 1958, 1961, 1979). However, this Early Devensian age is refuted by Madgett and Catt (1978) based upon intensive studies on all the tills of Holderness and Lincolnshire. Lake Humber I has been dated to 21,835±1600 years BP by Gaunt (1974, 1976, 1981) using a bone fragment found at the base of lacustrine sediments near Brantingham. This date has been questioned by Straw (1979) as it refutes his Early Devensian age for Lake Humber 1 and associated tills and landforms, but it is generally regarded as a maximum age for lake initiation. A minimum age of 11,110±200 years BP for lake emptying has been obtained from a buried soil at the top of the lacustrine sediment (Gaunt et al., 1971). Gaunt (1981) suggests that the lower lake was dammed by the moraine in the Humber after ice had receded and it actually silted up rather than drained through this moraine. A minimum age of 11,205±120 years BP on the deposition of the Hogsthorpe moraine was reported by Suggate and West (1959) based upon basal peat in a kettle hole.

2.3.5. Northwest England and Irish sea basin

In northwest England, the glacial deposits of the Blackpool area were reviewed by Wilson and Evans (1990), who emphasized a stratigraphy of upper and lower "boulder clays" separated by a thick sequence of stratified sediments that could record glacier recession followed by a readvance. This sequence has now been classified as the "Kirkham Formation" by Thomas (1999) and is correlated with the "Stockport Formation" of the Cheshire–Shropshire Lowlands (Maddy, 1999). Early deglacial dates in west Cumbria include one of 12,560±170 radiocarbon years BP from peat on the St Bees Moraine (Coope and Joachim, 1980) and another of 14,623±360 radiocarbon years BP from Windermere (Coope and Pennington, 1977). A basal moss layer lying over glacial outwash at Glen Ballyre on the northwest coast of the Isle of Man yielded a radiocarbon age range of between 18,900 and 18,400 ^{14}C years BP, providing an early deglacial date for the area (Shotton and Williams, 1971, 1973; Dackombe and Thomas, 1985).

The late Quaternary deposits of the Isle of Man have been traditionally subdivided into two suites, an upland suite of locally derived periglacial materials and a lowland suite of foreign glacigenic sediments (Thomas, 1976, 1977; Dackombe and Thomas, 1991). The upland suite has been explained as a product of cold climate slope processes either during Devensian glaciation, implying that the Manx uplands were nunataks, or paraglacial reworking of glacigenic materials during ice sheet wastage. Dackombe and Thomas (1991) support the latter interpretation based upon the ice sheet model of Boulton et al. (1977). The lowland suite has been interpreted as the product of terrestrial glacial deposition by Thomas (1976, 1977) and Dackombe and Thomas (1985, 1991) and as glacimarine sediment by Eyles and Eyles (1984) and Eyles and McCabe (1991). Although Thomas (1976, 1977) and Pantin (1978) have acknowledged that the receding Irish Sea glacier margin on the Isle of Man may have been partially marine based, the 200 m water depths required by the Eyles and Eyles (1984) and Eyles and McCabe (1991) model cannot be reconciled with the sea level history of the Irish Sea Basin at the close of the Dimlington Stadial. The last glaciation of the island by Irish Sea ice is recorded predominantly in the Shellag Formation and Orrisdale Formation of the lowland suite of deposits in the north (Thomas, 1999). These overlie Ipswichian raised beach and locally derived slope deposits. The Shellag Formation records the initial advance and recession of glacier ice (Shellag Advance, Thomas, 1976) and is unconformably overlain by the Orrisdale Formation. The latter was deposited during the Orrisdale Readvance, during which the Bride Moraine was constructed. Subsequent short-lived oscillatory readvances are recorded by the northward offlapping Jurby Formation (Thomas, 1976, 1999). New dates and a revision of the lithostratigraphy suggest a more complex sequence of events with an extensive glacial phase followed by rapid oscillations of the ice margin during deglaciation and then a substantial ice advance (Thomas et al., 2004).

2.3.6. North Wales

Considerable emphasis has been placed on the till sequences of the Lleyn Peninsula and their role in reconstructing the glacial advances in North Wales. Early work on glacial sediments revealed a tripartite sequence of two tills separated by stratified sediments (e.g. Jehu, 1909; Nicholas, 1915). This stratigraphy was used in conjunction with the drift morphology to propose multiple glaciations (e.g. Synge, 1963a, 1964), but more specifically it revealed that the Lleyn Peninsula was an area of coalescence between Irish Sea and Welsh ice. Sections on the north coast of the Lleyn record southerly flow of Irish Sea ice. Moreover, sites on the west of the peninsula contain evidence of Irish Sea till only, whereas sites between Gwydir Bay and Dinas

Dinlle in the east record Irish Sea and Welsh ice movement. Sections on the south coast of the peninsula are divided into those that contain only Welsh tills (e.g. sites east of Porth Neigwl) and those that are characterized by Irish Sea till. Intermediate sites like Porth Neigwl contain evidence of Irish and Welsh ice convergence. Complex till sequences, originally thought to be separated by weathering horizons (e.g. Synge, 1963a, 1964; Saunders, 1968b; Simpkins, 1968), are now accepted as the products of competing ice flows from Welsh and Irish Sea origins and/or the complex deposits of single ice advances (Boulton, 1977). Evidence for a readvance does exist at Dinas Dinlle, one of the north Wales coastal sites containing two tills and therefore central to proposals of multiple glacial events. The stratigraphy at Dinas Dinlle includes a lower, Irish Sea (Trevor) till, overlain by glacifluvial sediments (Aberafon Formation) and then Clynnog till. These have all been glacitectonically thrust by proglacial compression in front of an advancing glacier flowing towards 1708 (Harris et al., 1995, 1997), indicating that a readvance of Irish Sea ice can be supported by stratigraphic and geomorphic evidence regardless of the interpretation of the dual till sequence. Dating control on late Devensian events in North Wales is restricted to a few sites. A date of 29,000±1200 BP from shells in Irish Sea till at Porth Neigwl provides a maximum age on glaciation of the Lleyn Peninsula by Irish Sea ice. At Tremeirchion Caves, Vale of Clwyd, a radiocarbon date on mammoth bone of 18,000±1200 BP underlying Irish Sea till provides either a maximum age for glaciation of the area (Rowlands, 1971) or a maximum age of a readvance to the Bodfari–Trefnant moraine (Bowen, 1974). Organics in the Bryncir moraine have been dated $16{,}830^{+970}/_{-860}$ BP (Foster, 1968, 1970a). At Glanllynnau a radiocarbon date of 14,468±300 BP from peat over a complex till sequence in a kettle hole provides a date on the melting of stagnant ice on the south Lleyn Peninsula (Coope and Brophy, 1972). A deglacial date of 13,670±280 was reported by Ince (1981) from an infilled lake basin at Clogwynygarreg, west of Snowdon massif.

2.3.7. South Wales

In southernmost Wales considerable emphasis has been placed on Quaternary stratigraphy in order to demarcate the last ice limit in the absence of extensive glacial depositional landforms. Campbell and Bowen (1989) report that in southeast Wales the limit is marked by prominent moraines and hummocky moraine, part of the drift mapped by the Geological Survey and used by Charlesworth (1929). After early work by Charlesworth (1929) and George (1932, 1933) defining "older" and "newer" drifts, the ice limit on Gower has traditionally been drawn based upon stratigraphy. Specifically, the head deposits between Rotherslade in the east and Worm's Head in the west have been used as evidence of non glacial conditions during the last glaciation (e.g. Bowen, 1969, 1970). The exact position of the ice limit on the Gower was changed throughout the 1970s and 1980s by Bowen (e.g. Bowen, 1970, 1981; Bowen et al., 1985; Campbell and Bowen, 1989, Chap. 2) based upon interpretations of the local tills. A southerly excursion in the ice limit was proposed for the southwest Gower based upon the position of the "Paviland moraine" but this was revoked after "drilling and geophysical work". Firm stratigraphic evidence of Late Devensian ice on the west side of the Gower is provided by glacitectonically folded shelly till overlying Ipswichian raised beach deposits at Broughton Bay (Campbell et al., 1982; Campbell, 1984). Shells from the till have been dated to 17 ka BP (Bowen et al., 1986; Bowen and Sykes, 1988). This site, together with a large accumulation of ice proximal glacifluvial outwash at Rhosili Bay, constitutes the evidence for the western edge of the Carmarthen Bay ice lobe (Campbell and Shakesby, 1982; Stephens and Shakesby, 1982; Campbell, 1984). The terminus of the Swansea Bay ice lobe, fed by ice moving down the Nedd, Tawe and Afan valleys, was located in what is now an offshore position but glacial sediments occur on the Gower at Rotherslade (Langland Bay) as a sequence of ice-contact glacifluvial deposits (Rijsdijk, 2000) originally interpreted as till (George, 1933; Bowen, 1970; Campbell, 1984). The stratigraphic evidence for this piedmont lobe was provided by Al-Saadi and Brooks (1973), Culver (1976), Anderson and Owen (1979) and Culver and Bull (1979). The stratigraphy and sedimentology of the tills at Llanilid were used by Donnelly (1988) and Harris and Donnelly (1991) to verify their Late Devensian age compared to apparently older tills in the Vale of Glamorgan. A similar study proposed a Late Devensian age for tills at Pontypridd (Harris and Wright, 1980).

In southwest Wales John (1970a,b) and Bowen (1970, 1973a,b) identified stratigraphic evidence of Irish Sea ice on the Pembrokshire coast in the form of Irish Sea till often overlying Ipswichian beaches. John also suggested that the Preseli Mountains lay above the ice sheet margin during the deposition of the Irish Sea till. John (1965) reported outwash from this ice containing marine molluscs dating 37 ka BP. The southernmost limit of Irish Sea ice in SW Wales is interpreted differently by John (1968, 1970b, 1971) and Bowen (1971, 1974, 1977) based upon their views on the origin of diamictons of Irish Sea origin; John suggests that they are tills and therefore ice at least covered Milford Haven and West Angle Bay, whereas Bowen interprets them as colluvially redeposited materials and consequently draws the ice limit considerably further north, wrapping around the Preseli Mountains. Bowen (1977) and John (1970a) extend the ice limit as far south as the south shore of St Brides Bay based largely on an exposure through till overlying a supposed Ipswichian beach remnant at Druidston Haven. A kame terrace at Mullock Bridge near Milford Haven was used by John (1970a) as an Irish Sea marginal feature. Exposures through Irish Sea till at locations like Poppit Sands, Traeth-y-Mwnt and Abermawr on the north coast of Pembrokeshire document Irish Sea incursion during the Dimlington Stadial (Williams, 1927; Jones, 1965; John, 1968, 1970a, 1971; Bowen, 1977; Bowen and Lear, 1982).

Chronological control on glacial events in south Wales is restricted to a few sites. Shells in interglacial shorelines beneath glacigenic sediments and periglacial slope deposits on Gower have been equated with oxygen isotope stages 5e and 3, indicating that the glacial material dates to stage 2 (Campbell et al., 1982; Bowen, 1984; Bowen and Sykes, 1988). The most recent maximum age for LGM ice advance is that of 17 ka BP from glacitectonically folded till overlying Ipswichian raised beach deposits at Broughton Bay (see above). In Paviland Cave, remains of Homo sapiens sapiens have been radiocarbon dated to 18,460±340 BP years BP (Oakley, 1968), suggesting that the cave was occupied during the last glacial maximum (Campbell, 1977). At Banc-y-Warren, John (1967) reported wood dating to 31,800+2500/_1900 years BP and Brown et al. (1967) reported organic mud dating to $31{,}800^{+1400}/_{-1200}$ years BP, both from the sands and gravels. The organics contain *Pinus* pollen that John (1970a) suggests date to an earlier interstadial. Bowen (1984) more recently demonstrated that the sequence was Late Devensian and that all previous radiocarbon dates merely recorded reworking of pre-existing materials.

2.3.8. Welsh/English borders

A complex glacial history in the northernmost Welsh/English borderlands has long been appreciated based upon the stratigraphy of the region. Wedd et al. (1928) subdivided the Quaternary sediments in the area into a tripartite sequence including “Lower Boulder Clay”, “Middle Sands” and “Upper Boulder Clay”. The Lower Boulder Clay is of largely Irish Sea origin but includes intercalations of Welsh till. It is overlain in the Wrexham/Mold area by extensive spreads of outwash of local and northern provenance. The Upper Boulder Clay was regarded as coeval with the outwash and thought to be a lacustrine deposit recording lakes at the margin of the receding Irish Sea ice. The sequence was regarded as the product of a single ice advance during which the Middle Sands were deposited by receding Welsh ice, forming the Wrexham “delta–terrace” in a lake that covered much of Cheshire and Shropshire. Stratigraphic and geomorphic evidence has been employed over the years to propose readvances. Peake (1961, 1979, 1981) identified an Irish Sea till that was distinct from the Upper Boulder Clay and overlay the Wrexham “terrace” at Llay, proposing a Llay Readvance based upon that evidence. Francis (1978) suggested that the deposit was the product of a debris flow and therefore did not support the readvance theory. However, Dunkley (1981), Wilson et al. (1982) and Thomas (1985) have proposed a glacial landform evolution model for the Wrexham area that involves sequential wastage interrupted by readvances. The stratigraphy of the area has been clarified and linked to glacial geomorphic events by Thomas (1984a). Specifically, ice coalescence in the Alyn valley area is documented by the deposition of the Ruabon Till Member (Welsh till) over bedrock and under the Irish Sea Dee Till Member (previously named the Lower Boulder Clay). Overlying and partly intercalated with the Dee Till Member and the Ruabon Till Member is the Singret Member composed of the waterlain sediments of the Wrexham terrace and the Alyn valley. On the east side of the Alyn valley the Llay Till Member (equivalent of the Upper Boulder Clay) overlies the Singret Member and verifies the Llay Readvance of Peake (1961) and Dunkley (1981). The latter is thought to have been a minor ice marginal oscillation rather than a major readvance (e.g. Ellesmere Readvance of Peake, 1961, 1981). The Ruabon, Dee, Llay and Singret Members are grouped together as the Wrexham Formation by Thomas (1985).

The Smestow–Wolverhampton Line (Wills, 1924, 1938; Whitehead et al., 1928; Mitchell, 1960) has been traditionally regarded as the southernmost limit of the last glaciation in the West Midlands/Cheshire–Shropshire lowlands, although Boulton and Worsley (1965) have pointed out that tills to the south of the Woore moraine can be differentiated from those to the north based upon the depth of carbonate leaching, which is considerably greater to the south. The stratigraphic equivalent of the Stockport Formation, a complex lithostratigraphic unit recording glaciation, extends and thins up to this moraine. At Chelford the glacial sediments recording the last glaciation include till, outwash and lacustrine deposits (Stockport Formation; Worsley et al., 1983; Worsley, 1985) and overlie the Chelford Sands, which are non glacial alluvial fan sediments deposited during the early and middle Devensian. The Stockport Formation is the equivalent of what was originally called the Lower Boulder Clay–Middle Sands–Upper Boulder Clay sequence (see above). At Four Ashes the glacial sediments overlie Devensian organics dating 30 ka BP and at Stafford they are overlain by organics dating to 13.5 ka BP (Morgan, 1973). These dates thereby provide an age bracket for the deposition of the Stockport Formation. The surface till of the Stockport Formation, devoid of substantial morainic landforms, stretches from Church Stretton, Shropshire (see above) to Wolverhampton and then northwards where it abuts the south Pennine slopes (e.g. Greig et al., 1968; Hamblin, 1986; Hamblin and Coppack, 1995). The extent of the Devensian till sheet in the Birmingham area is described by Powell et al. (2000) as a line from north Dudley, through Walsall to Sutton Coldfield, based upon the earlier work of Martin (1891) and Eastwood et al. (1925), although they emphasize that the exact location of the margin is imprecise due to the difficulty in differentiating between pre- Devensian and Devensian tills.

2.3.9. Southwest England

The very restricted distribution and nature of the glacial deposits of the north coast of southwest England have fuelled considerable controversy about the exact nature of their deposition and their age. Synge (1977a,b, 1979, 1981, 1985) proposed that a floating ice shelf covered most of the Celtic Sea during the Devensian, emplacing glacigenic deposits at Fremington in Devon, Treberthick in Cornwall and on the Scilly Isles (see below). However, the Fremington till of Irish Sea origin (Mitchell, 1960; Stephens, 1966, 1970; Wood, 1974; Kidson and Wood, 1974) has been equated with earlier glaciations (originally Wolstonian but more recently Anglian) based upon its stratigraphic association with other dated deposits (Hawkins and Kellaway, 1971; Gilbertson and Hawkins, 1978; Andrews et al., 1984; Bowen et al., 1985; Bowen and Sykes, 1988). Although shell fragments from the Fremington deposits have yielded amino acid ratios indicative of a range of ages from early Pleistocene to late Devensian (Bowen, 1994), an Anglian age is preferred by Croot et al. (1996). Although the age of the glacigenic deposits in southwest England is now firmly placed in pre-Ipswichian times, extensive Dimlington Stadial ice in the southern Celtic Sea has been proposed by Scourse et al. (1991) based upon till-like deposits recovered from offshore sediment cores.

The maximum southerly extent of this Celtic Sea ice lobe has been mapped on the Scilly Isles using stratigraphic evidence for a Devensian glaciation (Scourse, 1986, 1987, 1991a,b; Scourse and Furze, 2001). Specifically, solifluction breccia overlying and enclosing organic layers has been observed to overlie a raised beach deposit. Radiocarbon dates and palynological evidence from the organics record a tundra grassland between 34 ka and 21 ka BP. The breccia is overlain by till (Scilly Till) in the northern Scillies and by loess on the southern isles, the loess having been dated by thermoluminescence at 18.6 ka BP (Wintle, 1981). Earlier work by Mitchell and Orme (1967) identified a southern limit for the glacial deposits that impinged upon the north coasts of the northern Scillies. They proposed a Gipping (Wolstonian) age for the till based upon comparisons with stratigraphic sequences in southwest England. Bowen (1969, 1973b) later argued that the till had been soliflucted after having been deposited originally in the Wolstonian. These age estimates were based upon "the interpretation of the number and age of the stratigraphically juxtaposed raised beach units" (Scourse, 1991b) on the isles and therefore are less secure than the age derived from large numbers of absolute dates on the sub-till breccia. Characteristics of the Scilly Till are summarized by Scourse (1991a,b; Hiemstra et al., in press). It contains abundant siliceous sponges and Miocene glauconite-bearing micritic limestone derived from the Jones Formation lying offshore to the north (Pantin and Evans, 1984), indicating that it is derived from the southern Celtic Sea. The continuation of the till offshore has been assessed by Scourse (1986) and Scourse et al. (1990, 1991, 2000) based upon samples collected and analysed previously by Pantin and Evans (1984). The characteristics of "till-like" materials in the Devensian Melville Formation (Cameron and Holmes, 1999) lying in the southern Celtic Sea have been interpreted as the deposits of an ice stream emanating from the Celtic Deep to the south of St George's Channel and flowing southwestwards towards the shelf edge. A change in depositional processes at approximately 49°30′ is thought to represent either a grounding line position or a change from proximal to distal glacimarine deposition (Scourse, 1986; Scourse et al., 1991). The "till-like" material in the Melville Formation is correlated with the Scilly Till by Scourse et al. (1991), implying that the Dimlington Stadial ice in the Celtic Sea was considerably more extensive than the previously accepted limits. This more extensive coverage has gained support from more recent work in southern Ireland (see below). In the Irish Sea, St George's Channel and north Celtic Sea contain two widespread deposits linked to the whole Devensian Stage but also documenting Dimlington Stadial glaciation (Cameron and Holmes, 1999). The Western Irish Sea Formation and Cardigan Bay Formation include glacigenic sediments relating to the last glaciation. Specifically, the Upper Till Member of the Cardigan Bay Formation is interpreted as a basal till and correlated with the Irish Sea Drifts of southern Ireland by Wingfield (1994, 1995). The Western Irish Sea Formation includes waterlain sediments and is sub-divided into Formation B, the uppermost of which is coeval with the Upper Till Member of the Cardigan Bay Formation, and Formation A, predominantly deglacial glacimarine sediments and the infills of deep glacial incisions (Wingfield, 1994, 1995). The very patchy nature of the Upper Till Member in the Celtic Sea south of 51820V prompted Wingfield (1994) to suggest that the Scilly Till probably represents a short-lived advance by the Dimlington Stadial ice stream.

2.3.10. Irish coast

On the southern Irish coast the distribution, origin and age of the Irish Sea Drifts, particularly the enclosed Irish Sea Till have been at the centre of controversies over the extent of the last glaciation for almost a century (e.g. Wright and Muff, 1904; Synge, 1981; Warren, 1985). Although the Irish Sea Drifts have been traditionally interpreted as terrestrial glacigenic deposits (e.g. Wright and Muff, 1904; Synge, 1978, 1981; van der Meer et al., 1994), an alternative glacimarine origin has been proposed by Eyles and McCabe (1989). Moreover, proposed ages for the deposits range from Midlandian/Devensian (Warren, 1985; Gallagher and Thorp, 1997; McCabe, 1999) to pre-Devensian (Mitchell et al., 1973; Synge, 1981; McCabe, 1987). Recent sedimentological analysis of the Irish Sea Till by O´ Cofaigh and Evans (2002a,b) and Evans and O´ Cofaigh (2003) has confirmed a terrestrial depositional origin, implying the onshore movement of an Irish Sea glacier lobe between Cork Harbour and Kilmore Quay during the last glaciation. The raised beach that stratigraphically underlies the Irish Sea Till has been dated 162–129 ka BP by infrared luminescence (Gallagher and Thorp, 1997), placing it firmly in the age range of the last interglacial. Previous interpretations depicted an ice margin moving offshore at Kilmore Quay and then trending eastwards across the north Celtic Sea. Recession by the Irish Sea glacier resulted in the damming of the local drainage along the south coast and the deposition of glacilacustrine sediments in ice-dammed lakes. Glacitectonic disturbance of the lake sediments by ice of inland origin documents the Midlandian glaciation of inland areas of southern Ireland previously thought to be covered by pre-Midlandian tills. The bouldery ridge of St Patricks Bridge and other arcuate ridges located to the east are ice marginal accumulations deposited by the Irish Sea glacier as it retreated from the southeast Irish coast. O´ Cofaigh and Evans (2002a,b) and Evans and O´ Cofaigh (2003) suggest that the Irish Sea Till on the southern Irish coast is the onshore record of the short-lived (rapid or surge) advance of the Irish Sea lobe recorded in the Celtic Sea and on the Scilly Isles.

Scottish Highland ice advance into Northern Ireland is recorded by a lower till sheet in County Down that contains Ailsa Craig microgranite erratics and shells radiocarbon dated at 24,050±650 BP (I-3268). If the date is correct it provides a maximum age on the advance of glacier ice onto Northern Ireland. An upper till of local derivation records the later expansion of Irish ice to exclude Scottish Highland ice from County Down and the coast of County Antrim (Hill and Prior, 1968).

3. Summary and discussion of the evidence for the last (Dimlington Stadial) British Ice Sheet

3.1. The vertical limits of the ice sheet

Recent developments in the cosmogenic isotope dating of mountain summit blockfields in upland Britain (e.g. Ballantyne et al., 1998a,b) has given Quaternary researchers confidence in identifying LGM nunataks and thereby demarcating the upper limits of the last ice sheet (e.g. Ballantyne, 1990; Ballantyne and McCarroll, 1995; McCarroll et al., 1995; Ballantyne, 1997; Ballantyne et al., 1997; McCarroll and Ballantyne, 2000). Although alternative interpretations of periglacial trimlines as the former boundaries of ice sheet thermal regimes persist (e.g. Kleman and Borgstrom, 1990), the increasing number of pre-LGM dates on blockfields (e.g. Stone et al., 1998; Bowen et al., 2002) together with independent evidence of prolonged periods of weathering, for example the occurrence of gibbsite (Ballantyne, 1994; Dahl et al., 1996), appear to be confirming the existence of full glacial unglaciated enclaves. In the absence of large numbers of sites with pre-LGM/sub-till dates, the continued mapping and dating of palaeo-nunataks will provide the most comprehensively dated upper limits of former glaciation in the British Isles.

3.2. The British Ice Sheet maximum limits

The greatest paucity of data regarding the LGM British Ice Sheet lies in the realm of dating maximum limits in lowlands and offshore areas. The most significant dates on the LGM are from the following regions:

(a) the northern North Sea where Sejrup et al. (1994) propose coalescent Scandinavian and British ice based upon a till dated 22 ka–29 ka BP. This was followed by a readvance, correlated with the Dimlington Stadial, between 19 ka and 15 ka BP;
(b) the NW continental shelf where a 27.3 ka BP maximum age for the LGM comes from the Isle of Lewis and glacimarine sediments distal to morainal banks south of St Kilda date to 22.5 ka BP;
(c) on Holderness where maximum ages for ice advance are 18.5 ka BP (Penny et al., 1969) and 17.5 ka BP (Wintle and Catt, 1985) but an age as recent as 14 ka has been proposed (Peacock, 1997);
(d) on the Lincolnshire Wolds, Lake Humber I is dated at 21.8 ka BP (Gaunt, 1974, 1976, 1981);
(e) in North Wales, the glaciation of the Lleyn Peninsula occurred sometime after 29 ka BP and a sub-till age of 18 ka BP in the Vale of Clwyd predates either the LGM or a later readvance (Rowlands, 1971; Bowen, 1974);
(f) in south Wales a maximum age on LGM glaciation of the west side of the Gower is 17 ka BP (Bowen et al., 1986; Bowen and Sykes, 1988), and a 36Cl age of 23.2 ka was obtained by Bowen et al. (2002) on the Arthur's Stone erratic boulder on the southcentral Gower;
(g) in the west Midlands, glacial sediments at the Four Ashes type site post date 30 ka BP (Morgan, 1973);
(h) on the Scilly Isles a maximum age for glacial advance is 21 ka BP (Scourse, 1986, 1987, 1991a,b);
(i) in Northern Ireland a maximum age for the advance of Scottish Highland ice is 24.1 ka BP. It is clear that significantly more geochronological control is required in order to elucidate the timing of LGM ice sheet advance, oscillations and retreat history in Britain. Although cosmogenic radionuclide dating is providing more success in demarcating the upper limits of the ice sheet, considerable problems remain in the lowland and offshore locations of ice sheet termini due to the lack of appropriate datable materials. This geochronological problem applies also to sedimentlandform associations deposited during the

Lateglacial when sub-Milankovitch scale climate oscillations are thought to have been influential in the readvances/ stadials of other ice sheet systems (see below).

3.3. Major ice-marginal depo-centres and possible readvances

Inside the maximum limits of the last British Ice Sheet, significant but to date under-utilized evidence exists for glacier marginal stillstands and/or readvances. Although radiocarbon dates of 14.7–14 ka BP on glacimarine sediments on the Irish coast provide some chronological control on the "drumlin readvance" or Killard Point Stadial (McCabe et al., 1998), very few of the many other moraine systems in the British Isles are temporally tightly constrained by radiometric dates. Because of this paucity in chronological control, regional correlations are speculative. For example, the two-phase LGM advance recognized in the North Sea by Sejrup et al. (1994) could correlate with events in the Vale of York where ice initially advanced all the way to Wroot in Lincolnshire and then later stabilized at the York and Escrick moraines. On Holderness, a similar sequence of ice marginal activity is recorded by the feather-edge of the Skipsea Till and the more recent ridge composed of coalescent ice-contact subaqueous fans that are approximately coeval with the Withernsea Till. The lack of a major moraine at the most extensive limit recognized in these areas compares with evidence on the Scilly Isles, the west Midlands and the south coast of Ireland, locations associated with the possible impacts of short-lived incursions by glacier ice. Prominent moraines like the Kirkham Moraine in Lancashire, the Whitchurch-Wem-Ellesmere-Wrexham moraine, the St Bees-Bride moraine, Brampton "kame belt" in Edenside and the Clynnog-Fawr and Dinas Dinlle moraines in North Wales are examples of ice sheet responses to either climatic forcing, internal dynamics or physiographic controls. The St Bees-Bride moraine has been tentatively dated to c.14 ka BP. Its formation was preceded by an episode of glacial overriding and glaciotectonism which extended beyond the moraine and was tentatively dated at 17 ka BP (Merritt and Auton, 2000). The concerted effort that was expended on the dating of moraines and other readvance features in this area should be replicated in other regions of lowland Britain in order to secure chronological control on ice dynamics, thereby facilitating inter-regional correlation and allowing assessment of climate-driven synchroneity.

Later glacier margins are demarcated by sequences of large cross-valley moraines, for example in the Usk drainage basin and the Pennine Dales. Large moraine systems in Scotland appear to record glacier margins that pre-date the Younger Dryas/Loch Lomond Readvance. For example, moraines demarcating valley glaciers in the Cairngorms are dated by cosmogenic isotopes to around 15 ka BP (Everest, 2002). The Wester Ross moraine is dated to around 13 ka BP (Ballantyne et al., 1987; Ballantyne, 1993) and may correlate with the Ardersier Oscillation, marked by a thrust moraine dated at approximately 13 ka BP in the Moray Firth. Any such correlations are extremely tentative until a more systematic dating programme is undertaken on the mountain valley moraines of Britain.

3.4. Palaeo-icestreams

Compared to other glaciated regions, very little research has been undertaken on the identification of ice stream imprints in the British glacial record, despite the availability of geomorphic criteria for the identification of palaeo-ice streams (e.g. Stokes and Clark, 1999, 2001; Clark and Stokes, 2003). Based upon the evidence compiled in the Glacial Map of Britain we suggest that the most compelling evidence available so far for palaeo-ice streams occurs in the Vale of York, Vale of Eden, the Yorkshire Dales and the Tweed Valley, where subglacial lineations document the impact of former trunk flow. Elsewhere, such as the Irish Sea basin and Moray Firth, ice stream activity has been proposed based upon topographic constraints and sedimentological evidence (e.g. Merritt et al., 1995; Evans and O´ Cofaigh, 2003). Although it is becoming clear that several generations of ice flow lineations exist and demonstrably cross-cut each other (e.g. Salt, 2001; Smith, 2003), reconstructions of British Ice Sheet palaeo-ice flow dynamics are in their infancy compared to other regions (e.g. Dyke and Morris, 1988; Dyke et al., 1992; Boulton and Clark, 1990a,b). We suggest that a systematic subglacial lineation mapping programme is urgently required for Britain in order to assess the dynamics of the former British Ice Sheet through the last glacial cycle.

3.5. Surging activity

Attainment of the maximum limits of the British Ice Sheet has been variously ascribed to surging activity. Eyles et al. (1994) employed the evidence of the Holderness hummocky moraine and the feather edge of Withernsea Till in conjunction with Kelsey Hill gravels to support their reconstruction of a surging ice sheet margin in East Yorkshire. The surging concept was first entertained by Lamplugh (1911) and later supported by Boulton et al. (1977). Additionally, the low ice profile of 1:750, as represented by the east coast drift limit from North Norfolk to the North Yorkshire Moors, was regarded as possible evidence of a surging lobe by Straw (1979). Some corroboration for surging in the North Sea comes from Carr's (personal communication) suggestion that the piercing of the western edge of the Dogger Bank-Jutland mega-moraine during the Dimlington Stadial was due to a surge. The Irish Sea Till on the southern Irish coast and the Scilly Till on the north coast of St Martins in the Scilly Isles (Scourse, 1986, 1987, 1991a,b; O´ Cofaigh and Evans, 2002a,b; Evans and O´ Cofaigh, 2003; Hiemstra et al., in press) have been equated with a short-lived onshore flow of the Irish Sea ice stream/ lobe, although the possibility of palaeo-surging is restricted to evidence of glacier marginal thrusting and hydrofracture fills produced by elevated groundwater pressures.

Suggestions have also been made that surges characterized some margins of the ice sheet during its recession. Evans and O´ Cofaigh (2003) cite evidence of thrust moraines on the eastern Irish and north Welsh coast as possible surge landforms. Long flutings terminating in glacitectonically thrust masses in Loch Ryan, SW Scotland, are cited by Salt (2001) as evidence of a late stage surge by Scottish ice.

Despite these proposals from various regions, no unequivocal geomorphic and sedimentary evidence for palaeo-surging has been reported from the area covered by the British Ice Sheet. Only future systematic, regional assessments of sediment-landform associations can facilitate comparisons with contemporary surging glacier landsystems (Evans and Rea, 1999, 2003; Evans et al., 1999) and therefore verify or refute palaeo-surging activity.

3.6. Ice sheet dynamics and amphi-Atlantic climatic events

Considerable advances have been made in recent years with respect to the correlation of millennial scale events around the margins of the North Atlantic, specifically in the identification of Dansgaard- Oeschger events and Bond Cycles in the Greenland ice cores (e.g. Johnsen et al., 1992; Bond et al., 1993; Dansgaard et al., 1993) and the massive calving episodes that affected the margins of the Laurentide and European ice sheets, which launched iceberg armadas into the neighbouring oceans and deposited the Heinrich layers (e.g. Heinrich, 1988; Bond and Lotti, 1995) that are now equated with short pulses of rapid cooling. Although these amphi-Atlantic events have been identified in the unusually long palaeoecological records at locations such as Gransmoor in East Yorkshire, Llanilid in south Wales, Whitrig Bog in southern Scotland and Borrobol in northern Scotland (Mayle et al., 1999), evidence of British Ice Sheet responses to the events is largely circumstantial (McCabe et al., 1998) and major moraine systems that could date to the same period covered by the Bond Cycles and Heinrich events remain largely undated (cf. Wester Ross moraine of ca.13 ka BP; Ballantyne et al., 1987; Ballantyne, 1993). Recent advances in cosmogenic isotope dating have allowed Everest (2002) to date valley moraines in the Cairngorms to 15–15.8 ka. This approach clearly needs to be extended to the major moraine systems of Britain in order to date ice sheet dynamics more precisely and allow comparisons with the high resolution palaeoclimate record available in the Greenland ice core stratigraphy.

4. Conclusion

This paper has reviewed the nature of the glacial geomorphological evidence for the last British Ice Sheet in Scotland, England, Wales, the Isle of Man and surrounding continental shelves. It is based upon the published literature and unpublished PhD theses and is compiled on a Glacial Map of Britain (Clark et al., 2004). The map also contains further information captured from unpublished BGS drift sheets not accompanied by memoirs, which are not reviewed here but are listed in the GIS data base. Although the Irish sector of the ice sheet has been excluded in this study, future work will concentrate on compiling a similar map for Ireland. The mapping and interpretation of glacial landforms reported here and depicted in the Glacial Map of Britain is patchy in coverage, nonsystematic in approach and spans more than 100 years of field research. As a result Britain presently has no comprehensive map of glacial landforms and this study has highlighted the extent of the shortfall so that future research can target critical regions and work towards the compilation of a glacial map that compares with those of other countries. Of priority for future glacial landform research is the comprehensive mapping of subglacial bedforms at a regional scale from satellite imagery. If this is undertaken by a small team and from first principles, the existing variability in data quality can be eliminated. Finally, despite recent advances in the dating of the upper limits of the British Ice Sheet, the timing of British Ice Sheet growth, decay and oscillation is poorly constrained. Only a more systematic programme of moraine dating will allow Quaternary researchers to reconstruct the palaeogeography of the British Ice Sheet and assess the pulse and synchrony of its response to amphi-Atlantic climate events.

Acknowledgements

Thanks to Colin Whiteman and Clive Auton for their constructive comments on an earlier draft of this paper.

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David J.A. Evans is a Reader in Physical Geography at the University of Durham. He completed his undergraduate degree in Geography at the University of Wales, Lampeter in 1982 before moving to Canada to undertake research on the glacial history of part of northern Labrador for his MSc at the Memorial University of Newfoundland. He then completed a PhD at the University of Alberta based on the glacial and sea level history of NW Ellesmere Island. After a brief spell as Lecturer at King's College, University of London, he then moved to Glasgow in 1990. David has recently been appointed as Reader in Physical Geography at the University of Durham. His main areas of research are in glacial geomorphology, glacial sedimentology, palaeoglaciological reconstruction and related sea level histories, and the development of glacial landsystems.

Chris D. Clark is a Professor in the Geography Department at the University of Sheffield, UK. His research techniques involve the analysis of Earth Observation and digital elevation models to solve largescale issues in palaeo-glaciology. Use of Earth Observation and DEM techniques permits analysis at the ice-sheet scale. He primarily works on landform mapping of the beds of the former North American (Laurentide) and British-Irish ice sheets, from which it is possible to reconstruct the flow geometry, ice dome position and disposition of ice streams. Another research focus is in utilising palaeo ice stream beds to elucidate basal processes of ice stream motion and shutdown. Dr Clark received his degree in physical geography from the University of Wales, Aberystwyth, and his PhD from the Geology and Geophysics Department, University of Edinburgh. He has worked at the University of Sheffield since that time and currently teaches courses on earth observation, glaciology and glacial geomorphology.

Wishart A. Mitchell is a Lecturer in Geography at the University of Durham. His research interests include ice sheet reconstructions, landslides and mountain geomorphology. Wishart received his MA in geography from the University of St Andrews and PhD from Royal Holloway, University of London.