Geomorphology and dynamics of the British-Irish Ice Sheet in western Scotland

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Declaration of Authorship

This thesis has been entirely written by myself, and is composed of work that has been undertaken by me, unless otherwise specified. Much of the content has been submitted for external publication in peer-reviewed journals; however, the work has not been submitted for any other degree or professional qualification.

Signed:

Date:

Lay Summary of Thesis

Present-day glaciers and ice sheets are undergoing change: their margins are retreating; their surfaces are lowering; and they are altering the speed and direction in which they flow. Because humans have only observed these changes for a short period of time, we do not fully understand what they might mean for long-term ice sheet behaviour and stability. Investigating the geological record can help. By piecing together evidence from landscapes that were formerly covered by ice sheets, we can start to understand how they evolved over hundreds to thousands of years. Because the ice is no longer present, we can also investigate the important processes that occurred in the *subglacial environment* underneath the ice – an area that is largely inaccessible at modern ice sheets, but often implicated in glacier flow.

This thesis examines the geological record from the last British-Irish Ice Sheet (BIIS) in western Scotland. Evidence from the landscape is used to investigate the magnitude and timing of large-scale ice sheet changes. Characteristics of the subglacial environment are examined in a novel way, using information from thousands of boreholes that have been drilled in order for our society to build cities and develop infrastructure on the former ice sheet bed. We find that the last BIIS underwent numerous changes in western Scotland, and that conditions under the ice played an important role in its evolution. The thesis contributes to our overall understanding of how the BIIS system evolved over thousands of years. These kinds of evidence-based *ice sheet reconstructions* are crucially important if we are to test the computer models that are used to predict long-term ice sheet behaviour.

Abstract

Predicting the long-term behaviour of present-day ice sheets is hampered by the short timescales of our observations and restricted knowledge of the subglacial environment. Studying palaeo-ice sheets can help by revealing the nature and amplitude of past centennial- to millennial-scale ice sheet change. This thesis uses glacial sediments and landforms to examine the evolution of the partly marine-based British-Irish Ice Sheet (BIIS) and its bed, in western Scotland. Three zones of the former BIIS are considered: ranging from a mountain ice cap, to a core area of the ice sheet, to a peripheral marine-terminating sector. The topography of the subglacial landscape was an important influence on the location of dynamic and stable components of the ice sheet. At an ice cap scale, zones of glacier inception and retreat were linked to catchment elevation and size. At the ice sheet scale, the migration of ice divides and thermal boundaries were focused through corridors of low relief subglacial topography. The main west-east ice divide of the BIIS in central Scotland migrated by 60 km, $\sim$10% of the ice sheet's width, through one such corridor during the glacial cycle. A major change in the flow regime of the BIIS in western Scotland accompanied the development of a marine-based sector on the Malin Shelf. As the BIIS advanced to the shelf edge, ice flow was drawn westwards – orthogonal to the earlier, geologically controlled, flow pattern. Retreat of the BIIS from the shelf edge occurred at an average rate of $\sim$10 m a^{-1}, but was punctuated by at least one episode of accelerated retreat at $\sim$100 m a^{-1}. In each zone of the BIIS examined, a rich palimpsest landscape is preserved and the role of earlier glaciations in conditioning or priming the landscape is highlighted. Western Scotland in particular is dominated by features relating to a 'restricted' mountain ice sheet, suggested to have been the prevailing ice sheet mode during the Early and Middle Quaternary. Where the last BIIS was underlain by soft sediments, glacier movement at the bed was facilitated by a combination of basal sliding and a localised mosaic of shallow deforming spots, allowing landform and sediment preservation. In places, till deposition was focused over permeable substrates acting to seal the bed, promote lower effective pressures, and enhance motion by basal sliding. The modern land surface in western Scotland provides an approximation for the relief of the former glacier bed, and can be used for conceptual palaeoglaciological reconstructions. Areas of focused postglacial deposition have, however, obscured parts of the ice sheet bed, with demonstrable implications for quantitative palaeoglaciological analyses. Methods to improve the representation of former ice sheet bed in these areas are discussed and may be pertinent to future palaeo-ice sheet modelling exercises.

Acknowledgements

Without a doubt, my interest in natural landscapes has been shaped by adventures with friends and family, and it has led to me onto a career at BGS that I very much enjoy. In 2009, I thought that it would be a good idea to start a part-time PhD, investigating the glacial landscape of western Scotland. I always knew that combining a PhD with a full-time job, and a growing family, would be a challenge. However, the various highs and lows have contributed to an extremely worthwhile experience, and I wouldn't change anything. As a friend of mine said at the start of a climbing trip several years ago, 'if this goes too smoothly, I'll not be satisfied!'

Without the help and support from a number of people this PhD would not have happened:

Tom Bradwell was instrumental in helping me get the PhD set up. Tom has provided endless advice, discussed the ideas in this thesis with me more than anyone else, and has read pretty much every sentence. For that, I am extremely grateful. I thank David Sugden for his enthusiasm and confidence in the project, and his ability to spark so many ideas – sometimes from a single comment. Thanks also to Pete Nienow, who was involved at the initiation of the project, and whose advice ensured that I tried to keep a focused, coherent research theme throughout.

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Part I

Introduction

Chapter 1

Introduction and aims

1.1 Rationale

Glaciers, ice caps and ice sheets presently cover ∼10% of the Earth's surface [Benn and Evans, 1998]. They are dynamic components of the cryosphere, capable of influencing and responding to climate change, storing and releasing large volumes of fresh water, and driving sea-level fluctuations [Clark et al., 1999, 2002; Jansson et al., 2003; Alley et al., 2005]. Modern studies have detected recent and ongoing changes in present-day ice masses, including variations in ice stream flow, rapid ice mass thinning and glacier front retreat, and switches in subglacial processes [Retzlaff and Bentley, 1993; Conway et al., 2002; Joughin et al., 2002; Cook et al., 2005; Motyka et al., 2006; Nesje et al., 2008; World Glacier Monitoring Service, 2008]. The significance of these changes in relation to longer-term patterns of ice sheet behaviour is difficult to assess, due to the restricted (generally <0.1 ka) timescales of observation. While there has been some success in unravelling longer, centennial- to millennial-scale changes at the margins of extant ice sheets [e.g. Bentley et al., 2005; Livingstone et al., 2012b; Roberts et al., 2013], fewer studies have been able to describe the extended evolution of their interior and core areas [Scherer, 1991].

The evolution and behaviour of ice sheets has been intimately linked to subglacial conditions, which are strongly implicated in glacier flow [Clark et al., 2002; Clarke, 2005; Marshall, 2005]. Modern ice sheet beds, however, are difficult to access and direct observations of substrate characteristics, basal thermal regime, and detailed bed topography are restricted. As a result, uncertainty surrounds many aspects of ice sheet basal conditions, such as the extent to which subglacial sediments might undergo spatially pervasive shear to high strains, enhancing ice motion [Boulton, 1979; Boulton and Jones, 1979; Alley et al., 1987; Engelhardt and Kamb, 1998; Iverson et al., 2003]. Recent geophysical investigations are beginning to provide a more detailed picture of modern ice sheet beds, providing clues about the processes that are occurring there [Smith et al., 2007; King et al., 2009], and offering some insight into longer-term ice sheet organisation [Bingham et al., 2012; Ross et al., 2014]. However, these types of detailed study are still relatively rare; so clear comparison datasets must be obtained elsewhere.

The lack of long-term observations of modern ice mass change and the inaccessibility of present-day ice sheet beds can be partially addressed by studying past ice masses. The geological record in landscapes formerly affected by glaciation can provide a wealth of evidence for the extent, thickness, and flow of past glaciers, ice caps and ice sheets. Relative age constraints combined with a growing collection of absolute age controls make it possible to deduce conceptual reconstructions of ice sheet evolution through growth and decay cycles over 1-100 ka timescales, providing insights into the nature, magnitude and rates of large-scale and long-term changes in ice mass organisation [Dyke et al., 2002; Hughes et al., 2011a; Kleman et al., 2010; Clark et al., 2012]. This approach can provide useful information about the potential sensitivity, or stability, of components of past ice sheets systems and form a powerful accompaniment to numerical simulations [Golledge, 2008; Hubbard et al., 2009; Stokes and Tarasov, 2010].

Former ice sheet beds also offer a solution to help improve our understanding of the subglacial processes that operate during ice mass evolution. Exposed ice sheet beds can be mapped over large areas using remotely sensed datasets [Boulton and Clark, 1990; Lidmar-Bergström et al., 1991; Hughes et al., 2010], and their sedimentological characteristics and properties assessed in detail at numerous accessible point localities [e.g. Lee and Phillips, 2008; Thomason and Iverson, 2009]. Recent systematic surveys over vast areas of formerly glaciated terrain are now providing valuable datasets that describe large populations of bedforms generated under ice sheets [Dunlop and Clark, 2006; Clark et al., 2009a; Stokes et al., 2013]. These represent a major step forward in trying to understand widespread processes and rates of ice and sediment movement at the base of ice sheets. The use of high-resolution digital surface models in the last decade has been a significant development in the investigation of ice sheet beds in deglaciated terrain. However, as analyses increase in resolution and become more quantitative, we also need to understand the assumptions and potential errors that can arise from their use.

In areas that have seen considerable subsurface geological investigation for societal development and exploration, large borehole datasets can provide another dimension to the study of past ice sheet beds. These types of resources have been used before in glacial geology [e.g. Menzies, 1981; Boyce and Eyles, 1991], but their potential has probably not yet been fully realised. The increasing availability of geological modelling software now makes it possible to map the subsurface architecture, sediment type and sediment thickness over wide areas. This could yield information about the patterns and volumes of sediment movement during glacial cycles, required to help understand subglacial sediment and ice movement mechanisms [Alley, 1991].

This thesis aims to contribute to the issues highlighted above, by using the landscape and sediments from beneath the mid-latitude former British-Irish Ice Sheet (BIIS) to reconstruct its long-term behaviour and its interaction with, and modification of, the subglacial bed. The first part of the thesis investigates three zones of the BIIS in parts of western Scotland, ranging from: (i) an isolated ice cap, to (ii) a core area of the BIIS, to (iii) a marine terminating margin (Figure 1.1). In each case the evolution and organisation of the ice masses are studied through a full growth-decay cycle. The second part of the thesis combines aspects of the derived ice sheet reconstructions with subsurface (borehole) datasets to examine the patterns and volumes of sediment that were moved during the ice sheet cycle. Three-dimensional lithostratigraphic models generated from subsurface data are also used to address some methodological questions, concerning scenarios when the modern land surface may not be fully representative of past subglacial topography. The sites chosen in this work possess excellent preserved glacial landscapes, and in some cases are uniquely combined with the wide availability of densely-spaced borehole information.

1.2 Research questions

For the regions of the BIIS examined, this work seeks to address the research questions outlined below, with the overall goal of reconstructing the evolution of ice masses and their beds during the last glacial cycle.

1. What were the regional patterns of ice mass growth and decay, and what changes in ice mass organisation occurred during their evolution?
2. Were particular phases of ice mass evolution dominant in their effects on the landscape?
3. How did the soft sediment ice sheet bed evolve in the Clyde basin during the last ice sheet cycle? What were the patterns and volumes of sediment moved,

and how does this compare with proposed mechanisms of subglacial ice/sediment motion?

4. What problems and uncertainties are associated with using the modern land surface to represent former ice sheet beds, and how can these problems be reduced?

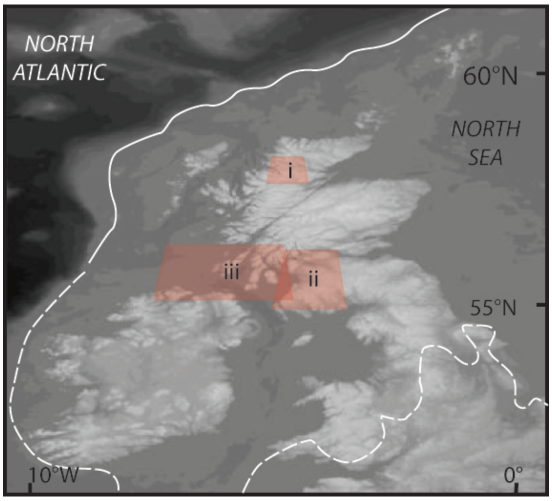

FIGURE 1.1: Zones of the former British-Irish Ice Sheet (BIIS) investigated in this project. The white line gives the approximate maximum extent of the last BIIS, based on Bradwell et al. [2008b] (solid line) and Clark et al. [2012] (dashed line).

1.3 Why the British-Irish Ice Sheet?

The former BIIS has been the subject of glacial geological research for more than a century, and a wealth of evidence has been documented [see Evans et al., 2005, for a review]. The BIIS has been one of the most intensively investigated former ice masses in the world, and given the existing body of work, it is not unreasonable to ask why still study it? Two key reasons are given here, which will be briefly discussed below in Sections 1.3.1 and 1.3.2. First, in the light of new developments that have taken place over the last decade, there are now opportunities to significantly improve our scientific understanding of how large-scale components of the BIIS evolved. This type of information is required to test and refine numerical models that are used to predict the future behaviour of modern ice masses. Second, continual planning and development for infrastructure in Britain and Ireland often require that new surface and subsurface (glacial) geological information is collected. This information can often be better interpreted and applied when there is an understanding of the sediment depositional environment, the (glacial) processes involved, and the wider ice sheet history.

The work reported in this thesis is primarily concerned with the first of these reasons. Separate more applied investigations of the glacial geology in the study sites have, however, been informed by outputs from this work [e.g. Finlayson, 2010; Finlayson et al., 2012; Finlayson, 2013].

1.3.1 Modern palaeo-ice sheet research in Britain and Ireland

Compilation of the BRITICE database in 2004 [Clark et al., 2004] triggered a renewed focus on BIIS research. By presenting a synthesis of existing geomorphological evidence at an ice-sheet-wide scale, it provided an initial template from which larger-scale ice sheet reconstructions, similar to those of the former Laurentide and Fennoscandinavian Ice Sheets, could be based. The synthesis also highlighted sectors of the ice sheet where little evidence had been described, and therefore could be used to direct new work. Shortly after the initial BRITICE work, two important new remote sensing datasets became available: the NEXTMap Britain elevation data from Intermap Technologies, and the OLEX bathymetric database, which provided a detailed sea-bed image for large parts of the British and Irish continental shelf. These datasets enabled an unprecedented view of the glacial landscape. They were used in key papers by teams of researchers; notably Bradwell et al. [2008b], who described the extent and decay of the BIIS over the northern UK continental shelf, and Clark et al. [2012], who, for the first time, reconstructed the pattern and timing of the retreat of the BIIS as a whole. In both these overviews, comparisons were drawn between the marine sectors of the last BIIS and the present-day West Antarctic Ice Sheet, highlighting the wider relevance of understanding how past ice sheets, such as the BIIS, evolved.

The availability of new high-resolution datasets has enabled analysis of the BIIS bed in more quantitative ways [e.g. Clark et al., 2009a; Spagnolo et al., 2012], and triggered new combined remote sensing and targeted field research, aimed at unravelling ice sheet dynamics through large parts of the last glacial cycle [e.g. Livingstone et al., 2009]. At the same time, the use of terrestrial cosmogenic nuclide (TCN) analysis to determine landform age has been steadily increasing. This technique has not only allowed new age constraints to be placed on ice sheet limits [e.g. Bradwell et al., 2008a; Ballantyne, 2010; McCarroll et al., 2010] , but has also demonstrated the long-term survival of some landforms under the ice sheet in upland areas [e.g. Phillips et al., 2006]. In this thesis, the use of subsurface (borehole) data is added to enhance our understanding of the former ice sheet bed in a soft sediment lowland area. The Clyde basin is probably one of the most densely drilled former glacier beds in the world, presenting an excellent opportunity to investigate its subsurface architecture and characteristics.

The merging of these developments and new datasets described above make it an interesting time to be involved in BIIS research. Indeed, the BIIS has recently been described as a 'conceptual playground' for glaciologists [Boulton, 2012], and it is fast becoming one of the best test beds for numerical ice sheet models, which are used to predict the future evolution of modern ice sheets [Clark et al., 2012].

1.3.2 Applications of glacial geological information

Glacial deposits cover 8% of the world's land surface, including approximately one third of Europe and a quarter of North America [Flint, 1971]. These deposits underlie many major cities and much of their associated infrastructure networks, and exert a significant influence on present-day groundwater systems. In Britain, ongoing development (e.g. energy supply, waste storage, and transport networks) mean that there are increasing demands on the underlying, glacially-shaped landscape. To inform planning and development decisions, there is a requirement to understand the likely ground conditions at and beneath the land surface. In glaciated terrain, this can be achieved by mapping glacial landforms and deposits, and identifying *glacial landsystems*, which comprise characteristic surface and subsurface materials. In this way ground investigations can be better targeted and the results more usefully interpreted [Eyles, 1983].

In Britain, the present need to upgrade transport and power networks requires an assessment of the surface and shallow subsurface materials over long distances. Detailed mapping of the glacial geology along these routes can aid interpolation between individual investigation sites, to better inform project planning and costing. In urban areas where geological mapping is more difficult, densely spaced borehole datasets can be used to map the distribution and geometry of glacial sediment packages beneath the land surface. Around the River Clyde in west central Scotland, three-dimensional models of glacial and post-glacial sediments are helping planners to anticipate ground conditions and identify contaminant pathways for a variety of brownfield regeneration projects [Campbell et al., 2010]. Providing geoscience information to end users in this way has been a significant output of the BGS Clyde Urban Super Project (CUSP), with which this PhD has been affiliated.

Over longer timescales, the processes and rates of landscape change are relevant to the selection and design of sites for storing radioactive waste. In Britain, the evolution of past ice sheets, and their impacts on the landscape, are therefore factors considered in these very long-term planning activities [e.g. Bradwell, 2010; Finlayson, 2010].

1.4 Format of thesis

This thesis is formed of five peer-reviewed papers, each addressing one or more of the research questions posed in Section 1.2. Each paper stands alone, providing relevant background information from the literature, methodological descriptions, and discussions. However, all of the papers contribute to the common goal of reconstructing the evolution of ice masses and glacier beds of the BIIS.

The papers are presented as chapters in Part II and Part III of this thesis. Part II presents geomorphological evidence from northern and western Scotland, and focuses mainly on reconstructing ice mass evolution through growth-decay cycles. Part III builds on aspects of these reconstructions, adding subsurface datasets to investigate the development and characteristics of the former glacier beds. The papers are summarised below in the context of the overall project, along with information regarding author contributions.

1.5 Summary of papers

1.5.1 Papers in Part II (Chapters 3, 4, and 5)

The sediment and landform record left by former ice masses provides information about their extent, flow, and behaviour. The metachronous nature of these records means that aspects of former ice flow and geometry can be reconstructed through extended periods of past glacial cycles, particularly when informed by well established stratigraphies. This type of information can provide a context for shorter-term changes observed in modern ice masses, and can form a strong accompaniment to the development and testing of numerical ice sheet models.

Studies of the former BIIS have often focused on maximum ice extent and retreat, particularly for ice masses that existed during the Younger Dryas stadial. Fewer studies have addressed the longer term evolution of these ice masses, including build-up phases. Part II of this thesis uses well-preserved geomorphological records to reconstruct the build-up and decay of former ice masses in three different palaeoglaciological settings: (i) an independent mountain ice cap in north-west Scotland; (ii) a core area of the former BIIS in west-central Scotland; and (iii) a peripheral, marine-terminating ice sheet sector in western Scotland.

Chapter 3 tackles the evolution of the Beinn Dearg ice cap through the Lateglacial period. The relatively small scale of the ice cap is compensated by the fact that it can be considered in its entirety, allowing characteristics at different margins to be compared. Prior to the Younger Dryas stadial, ice had disappeared from much of the Beinn Dearg range, retreating towards high central or eastern catchments. Renewed cooling into the stadial caused the build-up of an ice cap, found here to be much more extensive than suggested in previous reconstructions. The geometry of the geomorphologically reconstructed ice cap, with a dome centred over the western part of the range, is similar to that produced by numerical simulations. However, results suggest that the numerical simulation over-estimates the extent of western ice cap sectors, and under-estimates the extent of eastern ice cap sectors – possibly due to the omission of wind redistribution of snow in the model. The landscape in the Beinn Dearg range is found to be a glacial palimpsest including landforms from the previous deglaciation, which were overridden but not erased by subglacial processes

Author contributions: I designed the research, carried out the fieldwork, performed the geomorphological analysis and interpretation, and wrote the paper. Nick Golledge provided output from a numerical simulation (shown in Figure 3.10), and contributed to discussions. Tom Bradwell was involved during fieldwork and contributed to discussions. It was also Tom who initially highlighted the Beinn Dearg range as a potential 'problem area'. Derek Fabel processed the three samples that were collected for cosmogenic isotope analysis and calculated the exposure ages. Some fieldwork in this paper had been carried out prior to the start of the PhD in June 2009. However, fieldwork was completed in 2010, and all data compilation, interpretation and analysis was done subsequently.

Chapter 4 extends the investigation of evolving ice masses to a core area of the BIIS. A wide ranging compilation of remotely sensed geomorphological data, lithostratigraphical information, and field data are synthesised to reconstruct ice sheet advance, dynamics and decay in west central Scotland. The evidence points towards large-scale shifts in ice sheet geometry over millennial timescales – a result of repeated ice divide migration over the relatively low-lying ground of west-central Scotland. As in the first paper, the landscape is found to comprise a palimpsest assemblage of sediments and landforms, supporting the concept of the glacier bed as a mosaic of shallow deforming and stable spots (characterised by ice-bed separation and basal sliding), during phases of warm-based ice flow.

Author contributions: I designed the research, performed the geomorphological analysis and some of the geological modelling, interpreted the results, and wrote the paper. Jon Merritt was involved in numerous discussions and contributed some of the background and lithostratigraphy text. Mike Browne and Andrew McMillan had carried out extensive field research in the area during the 1980s, and were able to provide notes, sketches and photos of several (now closed or filled) sand and gravel pits and temporary sections from construction work. Jo Merritt and Katie Whitbread were involved in some of the geological modelling work and contributed to Figure 4.4.

Chapter 5 builds on the previous chapter by extending the geomorphological investigation westwards to include a marine terminating sector of the BIIS. Once more, a diverse, composite landscape is identified, with elements that were formed prior to, and during early, maximum, and late stages of the last glacial cycle. A major change in ice flow is shown to have accompanied the growth of a marine terminating ice sheet sector over the Malin Shelf. Both migrating and stable components of the ice sheet are identified, having been conditioned by subglacial topography. In this paper, there is a slight difference in interpretation from the previous paper, regarding the timing of deglacial events. This is partly due to the use of revised production rates to calculate cosmogenic exposure ages (an external factor), but also because extending the area under investigation presented incompatibilities with the earlier interpretation. While this highlights a potential pitfall in producing a thesis comprising published output from different progressive (intellectual) stages of the project, it faithfully reflects the evolution of ideas during a wider, ongoing research programme.

Author contributions. I designed the research, collected all the geomorphological and sedimentological information, interpreted the results and wrote the paper. Derek Fabel processed the five samples that were collected for cosmogenic isotope analysis, calculated the exposure ages, and contributed text to describe these procedures. Tom Bradwell and David Sugden contributed ideas and discussions.

1.5.2 Papers in Part III (Chapters 6 and 7)

Chapter 6 adds another dimension to the ice sheet reconstruction presented in Chapter 4. Through the use of an extensive borehole dataset, a three-dimensional computation of subsurface sediment distribution is derived, and linked to the ice sheet reconstruction, to elucidate patterns and volumes of sediment distribution. Analyses suggest that ice marginal and submarginal processes were responsible for the bulk of glacial sediment movement in the Clyde basin. Sediment movement under the ice sheet was probably more restricted, with motion at the glacier bed focused at, or very near to, the ice-sediment interface. The paper describes how different substrate characteristics might influence ice motion, and how these can change during the course of a glacial cycle.

Chapter 7 demonstrates another use of borehole data by removing postglacial sediments from surface models of the landscape. The rationale here is to investigate how the surface representation of postglacial sediments affects quantitative analyses that use the modern land surface as a proxy for former glacier bed topography. This paper differs somewhat in its approach to the previous ones, in that methodological issues are considered. However, the issues discussed are relevant to all the previous papers – indeed a finding from this paper is that some streamlined bedform measurements in Chapter 4 probably do not reflect their true subglacial morphology. Wider ranging implications regarding palaeo-ice sheet modelling are also considered and potential solutions suggested.

I designed, carried out, and wrote up all of the research in Chapters 6 and 7.

Chapter 2

Methods

2.1 Introduction

This chapter presents simply an overview of the principles, methods, datasets and software used during the project. More detailed methodological descriptions are given in each of the papers presented in Parts II and III.

2.2 Geomorphology

Geomorphological mapping forms the basis of any glacial landsystem assessment and palaeoglaciological reconstruction. Geomorphologically derived ice sheet reconstructions are based on the glacial inversion principle – that former ice sheet properties can be extracted from glacial geomorphology and geology [Kleman and Borgström, 1996; Kleman et al., 2006]. The underlying reasoning is that glacial landforms and glacigenic sediments hold spatial, chronological and glaciodynamic information regarding the conditions and context of their genesis, and these can be used as a basis to reconstruct former ice sheets [e.g. Greenwood and Clark, 2009; Kleman et al., 2010; Clark et al., 2012]

Principles of the inversion approach have long been used at some level to identify former glacier flow or geometry, using features such as striations, drumlins, and moraines [e.g. Geikie, 1863; Charlesworth, 1955; Price, 1975; Sissons, 1977]. However, the method was formalised by Kleman and Borgström [1996], with particular emphasis on ice-sheet-scale reconstructions and the use of glacial lineations. Of particular significance in the methodological account by Kleman and Borgström [1996] was the recognition that the landform record was formed metachronously and not just during deglaciation from maximum ice extent, and that landforms could be preserved under ice sheets for prolonged periods (the role of basal thermal regime).

The following assumptions, employed in the inversion approach, were outlined by Kleman and Borgström [1996] and Kleman et al. [1997].

- The basic control on subglacial landform creation, preservation, modification and destruction is the location of the phase boundary between water and ice, separating frozen from thawed material at or under the ice sheet base (i.e. basal temperature).
- Basal motion requires a thawed bed.
- Glacial lineations (positive and negative streamlined bedforms) can only form if basal motion occurs.
- Glacial lineations are created parallel to local ice flow directions and perpendicular to the ice sheet surface contours at the time of formation.
- Frozen-bed conditions inhibit the reshaping of the subglacial landscape.
- Regional deglaciation is always accompanied by the creation of a spatially coherent but metachronous system of meltwater features such as channels eskers and glacial lake shorelines.

Here, the term 'basal motion' is preferred to 'basal sliding', which was originally used by Kleman and Borgström [1996]. This is because basal sliding is often associated with ice-bed separation, whereas some degree of ice-bed coupling (e.g. clast ploughing, sediment erosion / deposition, deformation) is implicated in doing the geomorphic work.

2.2.1 Remote sensing data

Regional to ice-sheet-wide reconstructions, because of the large areas involved, are dependent on the use of remote sensing data to capture geomorphological information. In this PhD, two sources of remotely-sensed data were used: (i) the NEXTMap Britain digital elevation dataset; and (ii) aerial photographs. The NEXTMap Britain dataset was collected by Intermap Technology's NEXTMap programme in which airborne Interferometric Synthetic Aperture Radar (InSAR) is used to create digital elevation model (and image) products with a vertical accuracy of 1 m and sample spacing of 5 m [Mercer, 2007]. The data were acquired in 2002/3 (England, Wales and Southern Scotland) and 2004 (rest of Scotland). Processed digital surface models (DSMs), digital terrain models (DTMs) and orthorectified radar images (ORI), obtained by BGS, were used in this study. In addition, digital, georeferenced, monoscopic colour aerial photographs at 0.25 m (for display at 1:2500) resolution were used throughout the areas investigated in this project. These photographs were acquired by Getmapping / UKP and are licensed to BGS. Mapping in the Beinn Dearg area also used 1:24,000-scale black and white stereo aerial photographs, flown during the All Scotland Survey in 1988/89.

2.2.2 Field mapping

While a remote sensing approach is clearly essential for the investigation of glacial landscapes over large areas, field investigation can also contribute significantly to overall glacier reconstructions. Landforms that are too small to be identified by remote sensing data, or are ambiguous, can be verified in the field. Sediment exposures in the field reveal the composition and internal structure of landforms, which can be used to assign their formative conditions. Furthermore, underlying sedimentary sequences that do not relate to conditions represented by surface landforms can enable the extension of ice sheet reconstructions further back in time. In this sense, an understanding of the time sequence of deposits (lithostratigraphy) can play an important role in constructing an overall ice sheet event stratigraphy.

Locations for targeted field mapping were selected following an initial analysis of remote sensing information. Field mapping was carried out by walkover survey during which observable geomorphological features, geological boundaries, and natural sediment exposures were recorded. Observations made in the field were captured using custom-built field mapping tools (SIGMAmobile) in a Geographical Information System (GIS) environment, on a ruggedised tablet PC with a built-in GPS. The use of a GIS in the field enabled interrogation of remote sensing datasets in the field, together with other geographically referenced information (e.g. existing BGS solid geology and superficial geology maps and scanned field maps from late 19th and early 20th century geological surveys).

2.2.3 Exposure dating

Remote sensing and field based mapping provided a geomorphological context for the selection of sampling sites for cosmogenic nuclide analysis, aimed at improving chronological constraints in the areas studied. This part of the research was carried out in collaboration with Derek Fabel at Glasgow University. Eight samples were collected in the course of this PhD; three from the Beinn Dearg Range and five from Arran and Kintyre. In each case I selected the sample locations, collected samples, and took field measurements. Sample preparation, laboratory measurements, and age calculations (described in Chapters 3 and 5) were undertaken by Derek.

2.3 Geological modelling

To map the three-dimensional geometry and distribution of sediment packages beneath the land surface, a geological modelling approach was employed. The rationale behind this approach was that the distribution of sediment packages could be linked with an ice sheet reconstruction, in order to elucidate patterns and volumes of sediment movement associated with the glacial cycle. Furthermore, the ability to 'peel away' layers of sediment from the land surface could allow better topographic representation of former glacier beds, with implications for quantitative palaeoglaciological and geomorphological studies.

The geological modelling relied heavily on the availability of densely spaced borehole information. For this reason it was restricted to the vicinity of the Clyde basin, where borehole records from geotechnical and hydrogeological investigations, sand and gravel assessments and coalfield investigations are numerous. Because the original focus of borehole investigations differed, the detail concerning the glacigenic deposits is variable. However, interpretations were informed by comparison with, and the use of, reliable logs from boreholes drilled solely for the investigation of glacial and postglacial sediments [Browne and McMillan, 1989b] (Figure 2.1 A,B). In addition, numerous 'type sections' from around the Clyde basin were visited, providing a grounding for borehole interpretation (Figure 2.1 C).

Two approaches for geological modelling were used. Over a large area (the Clyde basin), a model was derived from a dense digital cross section network (fence diagram), combined with envelopes that mark the lateral extent (surface and buried) of sediment units. The cross sections and envelopes were interpolated manually from borehole information and geological maps. Points on the cross sections and envelopes

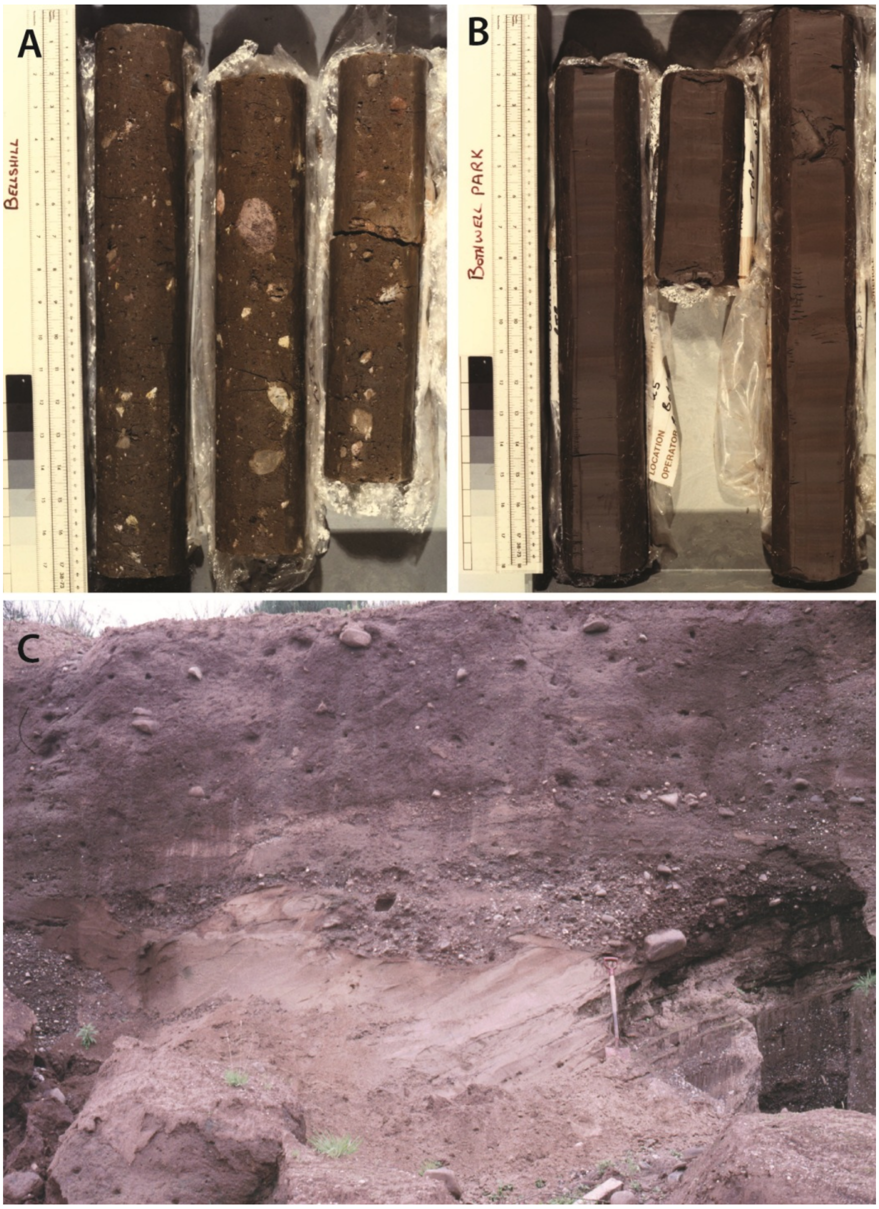

FIGURE 2.1: Examples of core records and sediment exposures that provided reference data during the interpretation of borehole logs. A. Stiff, poorly sorted, sandy clay diamict of the Wilderness Till Formation, from the BGS Bellshill borehole. B. Soft to firm, thinly laminated clays and silts of the Paisley Clay Formation, from the BGS Bothwell Park borehole. C. Stiff, poorly sorted, sandy clay diamict of the Wilderness Till Formation erosively overlying glacitectonised sands and gravels of the Cadder Sands Formation. Photo by Mike Browne.

were then triangulated to create triangulated irregular networks (TINs) for the tops and bases of each sediment unit. Over a smaller area in Glasgow, more uniform and densely spaced borehole coverage meant that a fully geostatistical interpolation could be adopted. Both approaches have their advantages and disadvantages: the former allows for geological / geomorphological reasoning, while the latter is more objective with quantifiable errors. These were considered in the selection of an appropriate geological modelling methodology for the problems tackled. Further descriptions of the geological modelling methods are given in Part III.

2.4 Data compilation and software used

All of the geomorphological information was compiled and managed using ESRI ArcMap and ArcCatalog. Data captured in the field were directly imported from the SIGMAmobile project into the master GIS project where all information was held. In the GIS environment, datasets were interrogated and processed using ArcTools. Further statistical treatment and analysis were performed in GNU Octave. Basin-wide geological modelling was undertaken using GSI3D. ASCII grids, geology shape files and geomorphological information were imported from ArcGIS, and borehole log and location files were imported from a nationwide database (the Borehole Geology Database) via an intranet portal. Interpolated surfaces were then exported to GOCAD for refinement and interrogation. Smaller-scale geostatistical modelling, including variogram analysis, was wholly carried out in GOCAD. This thesis was written and typeset using LaTeX.

Part II

Geomorphology and ice sheet reconstructions

Chapter 3

Lateglacial icecap evolution in northern Scotland

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Abstract

Detailed geomorphological mapping of the Beinn Dearg massif in northern Scotland was conducted to examine the maximum (Younger Dryas) extent, and earlier interstadial evolution, of an ice cap that existed during the Lateglacial period (14.7-11.7 cal. ka BP). Landform evidence indicates a plateau ice cap configuration, with radial outlet glaciers, during the Younger Dryas. The interpreted age is supported by new cosmogenic exposure ages, and previously reported interstadial sediments beyond the ice cap margin. The reconstructed Younger Dryas Beinn Dearg ice cap covered 176 km^2, with its summit positioned over the western side of the massif. Area-altitude balance ratio (AABR) equilibrium line altitudes (ELAs) of 570-580 m were calculated for the ice cap as a whole. The empirically reconstructed ice cap is compared with recent numerical model simulations; both methods produce an ice cap with a similar configuration. However, differences are apparent in the extent of eastern and western outlets ($\pm$1-5 km), and in the spatial variation of ELAs. Results suggest that the numerical simulation over-estimates the extent of western ice cap sectors, and under-estimates the extent of eastern ice cap sectors. We attempt to quantify these differences in terms of ice cap mass balance and assess their possible causes. Geomorphological evidence for pre-Younger Dryas ice cap configuration indicates that the Beinn Dearg massif remained an important source during earlier deglaciation. In contrast, the neighbouring Fannich mountains acted as an unzipping zone, and were ice free on their northern side by the Allerød (Greenland Interstadial 1c to 1a). Deglaciation continued over western parts of the Beinn Dearg plateau, with the possibility that glaciers remained in some central and eastern catchments, prior to (Younger Dryas) ice cap (re)growth.

3.1 Introduction

Reconstructions of palaeo-, or formerly more extensive, ice masses in northwest Europe have enabled inference of past glacier mass balance and climate, and allowed the causes of ice mass fluctuations to be assessed [e.g. Ballantyne, 1989; Dahl and Nesje, 1992; Carr, 2001; Rea and Evans, 2007; Golledge et al., 2009; Nesje, 2009]. In the Scottish Highlands, the last decade has seen a renewed focus of research into the extent and behaviour of ice masses during the Lateglacial Younger Dryas (YD), or Loch Lomond Stadial (Greenland Stadial 1 (GS-1)) (12.9-11.7 cal. ka BP [Lowe et al., 2008]) [e.g. Ballantyne, 2002, 2007a; Benn and Ballantyne, 2005; Finlayson, 2006; Golledge, 2007; Lukas and Bradwell, 2010]. Key outcomes of this research have been the identification of spatial trends in equilibrium line altitudes (ELAs) and palaeoclimate inferred from glacier reconstructions [Ballantyne, 2007b]. More recently, these data have provided important 'targets' for numerical glacier simulations, aimed at identifying the envelope of climatic parameters which allow growth of ice masses that best fit empirical reconstructions [Golledge et al., 2008]. The use of cosmogenic nuclide analysis has furthered this work by allowing more reliable assessment of landform age [e.g. Everest and Kubik, 2006; Golledge et al., 2007; Fabel et al., 2010]. Significantly, it has been demonstrated that a number of moraine sequences beyond the 'limits' of reconstructed Younger Dryas glaciers in northern Scotland were formed during the warmer Lateglacial interstadial (Greenland Interstadial 1 (GI-1)) (14.7-12.9 cal. ka BP [Lowe et al., 2008]), requiring the existence of larger ice masses at that time [Bradwell et al., 2008a; Ballantyne et al., 2009] (Figure 3.1). This has opened an exciting avenue of research into behaviour of Lateglacial interstadial (GI-1) ice caps in Scotland [e.g. Stoker et al., 2009; Bradwell and Stoker, 2010; Ballantyne et al., 2009].

Despite important advances arising from the work described above, there remain large areas in upland Britain that have not yet received detailed glacial geomorphological investigation. Thus, assessment of ice mass configuration during the YD is not complete. Furthermore, the geomorphological signature which relates to events immediately prior to the most recent (YD) glacial expansion has often been neglected. However, earlier landforms may shed light upon changes in glacier dynamics and configurations between GI-1 and the YD. Such evidence should yield important clues to whether or not significant ice masses survived throughout GI-1 [cf. Bradwell et al., 2008a]. This study attempts to address these issues for the Beinn Dearg massif (Figs 3.1, 3.2, and its surrounding area, in northern Scotland. Specifically, we: (i) map in detail the glacial landforms in, and surrounding, the Beinn Dearg massif; (ii) assess the age of landsystem components using existing and new dating evidence; (iii) reconstruct the

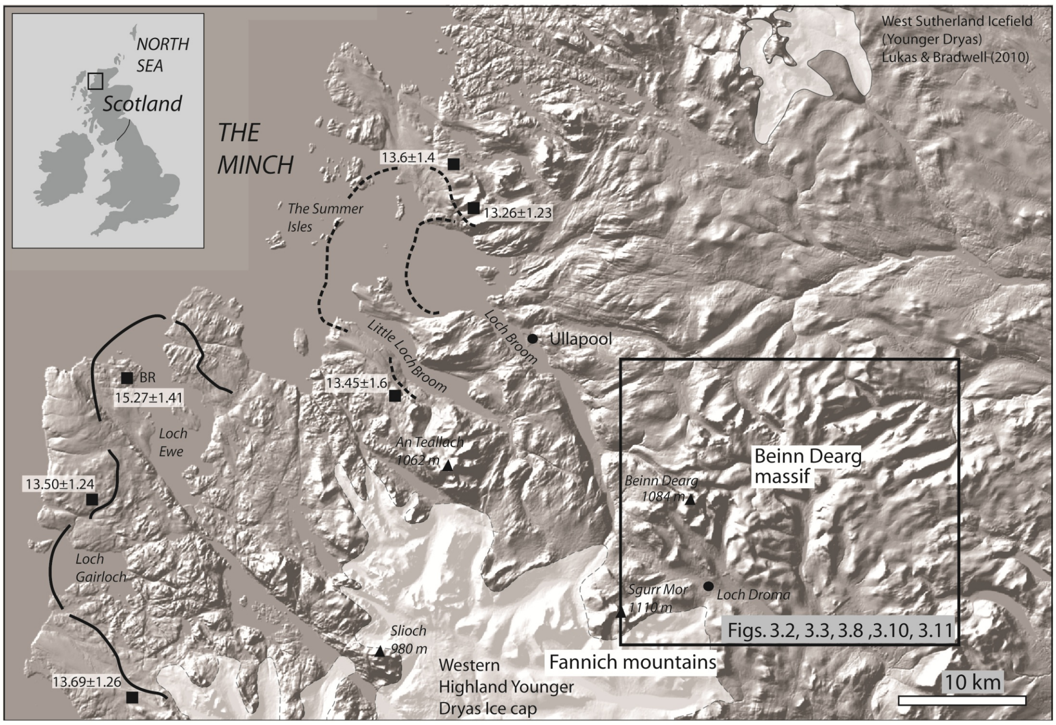

Figure 3.1: Regional context. Box (labelled Figs 3.2, 3.3, 3.8, 3.10, 3.11) shows location of study area. Mean cosmogenic exposure ages are shown for sites reported in Bradwell et al. [2008a], and Ballantyne et al. [2009]. All ages are for moraine ridges, except for the age labelled 'BR' by Loch Ewe. Black lines delimit reconstructed ice sheet margins in the vicinity during the early part of Greenland Interstadial-1. Dashed line – Bradwell et al. [2008a]; solid line – Ballantyne et al. [2009]. The main Western Highland Younger Dryas ice cap is taken from the BRITICE Map [Clark et al., 2004]. Hill-shaded digital surface model built from Intermap Technologies NEXTMap Britain topographic data. Northwest illumination.

approximate dimensions of the Beinn Dearg ice cap during the YD, making quantitative comparisons with recent simulations of a numerical ice sheet model [Golledge et al., 2008]; and (iv) use the geomorphological evidence to elucidate behaviour of the ice cap during the GI-1 to YD transition, developing a conceptual model of mountain ice cap evolution during this period of rapid climatic adjustment.

3.2 Study area

The Beinn Dearg massif in northern Scotland forms a broad, ∼300 km^2 upland plateau, lying to the east of Loch Broom (Fig 3.1). Approximately 150 km^2 (50%) of the massif lies over 600 m above sea level (a.s.l.), with the highest point, Beinn Dearg, rising to 1084 m. The plateau is dissected by a number of steep sided, radial valleys and corries. In contrast, the neighbouring Fannich mountains, to the southwest, are characterised by narrower, sharp peaks, amongst generally lower ground, with less than 30% of the area lying above 600 m a.s.l. Bedrock in the region is composed mostly of Neoproterozoic metasedimentary rocks (psammites and pelites), although a broad area of gneissose granite occurs in the south-eastern corner of the Beinn Dearg massif [B.G.S., 2004].

Glacial landforms in the area were first documented by Peach et al. [1912, 1913], who envisaged a final glacial stage where independent ice centres formed over prominent masses of high ground. The entire Beinn Dearg massif, and the Fannichs to the south-west, are within the limits of the 'Stage M' glaciation proposed by Charlesworth [1955], which elsewhere approximates several accepted Younger Dryas margins [Golledge, 2010a]. Interstadial sediments of at least Allerød (GI-1c to GI-1a) (14.0-12.9 cal. ka BP [Lowe et al., 2008]) age, discovered at Loch Droma (Figs 3.1, 3.2 [Kirk and Godwin, 1963], demonstrate that it was ice free by that time. These sediments had not been modified by any subsequent glacier advance, and thus conflict with the association of Charlesworth's 'Stage M' with YD glacier limits in this area. Kirk and Godwin [1963] and Kirk et al. [1966] suggested that a local glaciation of the Beinn Dearg massif (their 'Gharbhrain Stage') formed moraine complexes at Loch Gharbhrain, and Strath Vaich and Strath Rannoch (Fig. 3.2), inferring it to be the local equivalent of the Loch Lomond Readvance (YD). Bennett and Boulton [1993] also considered these localities to have been occupied by glacier ice at that time, with a separate glacier complex in the Fannich mountains (Fig. 3.2). However, Sissons [1977] considered lower Strath Vaich and Strath Rannoch to have been ice free during the Younger Dryas, and presented a more restricted reconstruction, consisting of thirteen independent glaciers in the Beinn Dearg massif, and a merged glacier complex sourced in the north and east facing corries of the Fannichs (Fig. 3.2). More recently, Reed [1988], Ballantyne [1997], and Finlayson and Bradwell [2007] have suggested that a more extensive ice field or ice cap existed over the Beinn Dearg massif during the Younger Dryas – a configuration that is also simulated by numerical modelling experiments [Golledge et al., 2008; Golledge, 2010b].

3.3 Methods

Glacial landforms in, and surrounding, the Beinn Dearg massif were digitally mapped in the field, using a ruggedized tablet PC, with a built-in GPS and ArcGIS 9.2 software. Using the GIS in the field enabled on-site interrogation of georectified digital aerial photographs, topographic maps, and NEXTMap digital surface models (DSMs), with

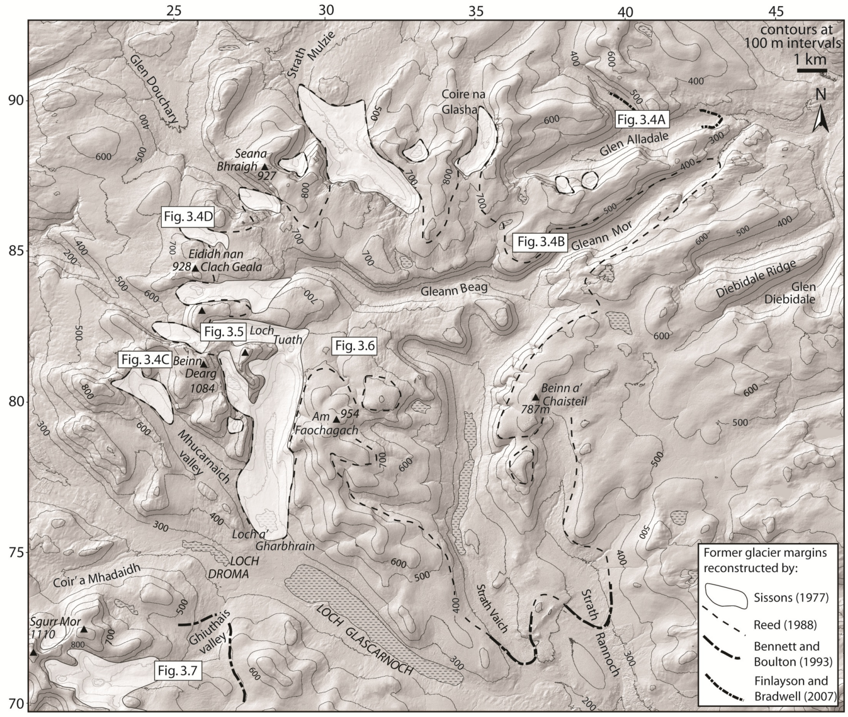

FIGURE 3.2: Topography of study area and previous glacier reconstructions. Locations of subsequent figures are shown. Contours taken from Intermap Technologies NEXTMap Britain topographic data.

1 m vertical and 5 m horizontal resolution. Aerial photographs and hill shaded DSMs were also studied, both prior to and following, field investigation.

During the field mapping, boulders from a conspicuous, well-preserved moraine complex in Glen Alladale [Finlayson and Bradwell, 2007] were sampled for cosmogenic nuclide analyses (A 1, 2, 3 in Figs 3.3 and 3.4A) (Table 3.1). This site was chosen because it (i) was beyond the limits of previously reconstructed YD glaciers in the area [Sissons, 1977; Reed, 1988], and (ii) is at the opposite (northeast) side of the massif from where existing chronological control exists (Loch Droma). Samples were chiselled from the upper surfaces of two psammite boulders and one vein quartz boulder on a single ridge of the moraine complex. Skyline topography was measured in the field at 15° increments, to allow calculation of topographic shielding. Samples were processed at the University of Glasgow's Centre for Geosciences cosmogenic isotope laboratory using methods adapted from Kohl and Nishiizumi [1992], Ditchburn and Whitehead [1994] and Child et al. [2000]. Beryllium ratios were determined at the Scottish Universities Environmental Research Centre (SUERC) AMS facility.

Sample (lab code)	Lithology	Elevation	Latitude (°N)	Longitude (°W)	Thickness correction [1]	Shielding factor [2]
A1(b3306)	Gritty psammite	272	57.8683	-4.7013	0.9833	0.9732
A2(b3307)	Gritty psammite	270	56.8687	-4.7017	0.9833	0.9673
A3(b3085)	Vein quartz	265	57.8689	-4.7029	0.9833	0.9683

TABLE 3.1: Sample locations for ^{10}Be exposure dating. [1]Calculated assuming a density of 2.7 g cm^{-3} and an attenuation length og 160 g cm^{-2}. The tops of all samples were exposed at the surface. [2]Calculated according to Dunne et al. [1999].

3.4 Glacial geomorphology

The results of the geomorphological mapping are summarised in Fig. 3.3. Some detail is lost in reproducing the map at this scale, although more detailed sections are described below. GIS shapefiles of the landform data are available from the corresponding author. The glacial landforms in, and surrounding, the Beinn Dearg massif are described below, under geographical zones.

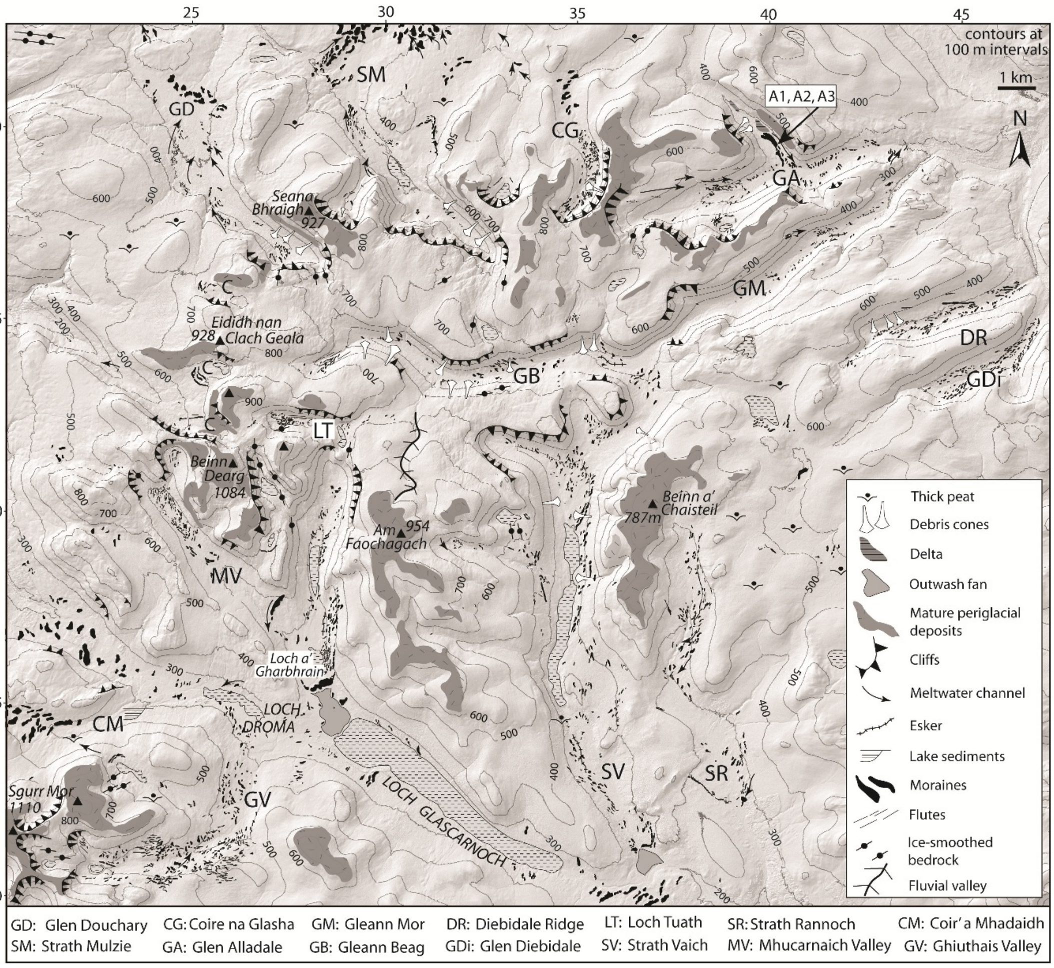

FIGURE 3.3: Glacial geomorphology of the Beinn Dearg massif, and immediate surroundings. A1, A2, A3 indicates location of sampling site for cosmogenic exposure dating. Note, 'mature periglacial features' are taken to include: relict bouldery solifluction sheets and lobes, blockfields, and thick talus deposits.

3.4.1 Northern valleys

Discrete, down-valley limits of closely spaced, small (10-30 m wide, 2-5 m high), hummocky recessional moraines are present in Glen Douchary, Strath Mulzie, and Coire na Glasha (Fig. 3.3), and were noted by Kirk et al. [1966]. These moraines contrast markedly with mounds farther to the north, which are more rounded, generally larger in size (many > 100 m width), with more gentle slopes. The abrupt transition between the two moraine types suggests a change in the characteristics of the ice mass that produced them. Lateral moraines and marginal meltwater channels continue up the valleys from the hummocky moraines, and eventually give way to debris flow-incised slopes, mantled with glacial sediment. A down-slope limit of thick talus coincides with the upper limit of thick sediment on the eastern side of Glen Douchary, and rises southward up the valley from 500 to 550 m a.s.l. The low-angle slopes of spurs between these valleys possess periglacial features in the form of blockfields and well-developed, relict, bouldery solifluction sheets and lobes. However, ice-moulded bedrock is abundant above the valley headwalls (600-650 m a.s.l.), indicating that plateau ice fed the valleys below.

3.4.2 Eastern valleys

Glacial landforms in Glen Alladale were described by Finlayson and Bradwell [2007]. In lower Glen Alladale, they found an assemblage of moraine ridges trending perpendicular to the valley axis, and a large boulder-strewn, multi-crested moraine ridge complex against the eastern side of a tributary valley to the north (Fig. 3.4A). Together with asymmetric recessional moraine distribution, high level meltwater channels, and bouldery lateral moraines, Finlayson and Bradwell [2007] concluded that the deposits most likely related to an outlet of an ice cap centred over the plateau to the southwest. Previous studies had only reconstructed small corrie glaciers in this vicinity [Sissons, 1977; Reed, 1988], thus the site was selected for cosmogenic nuclide dating (see below).

At the eastern end of Gleann Mor, suites of recessional hummocky moraines suggest westward ice margin retreat from a down-valley limit at NH 430 879 (200 m a.s.l.). The moraines can be traced up the northern valley side to a very well-preserved lateral moraine, which at 500 m a.s.l. merges with the plateau surface (Fig. 3.4B). Thus, the assemblage is consistent with that in Glen Alladale [Finlayson and Bradwell, 2007], suggesting a plateau ice mass that fed into the valleys.

Suites of closely spaced, hummocky, recessional moraines, and well-preserved lateral moraines occupy the valleys flanking the Diebidale Ridge. The moraines terminate close to the eastern end of both valleys, at approximately 250 m a.s.l.; an end moraine ridge bounds a small loch in lower Glen Diebidale. The down-valley limits of these moraines probably indicate former glacier margins, broadly contemporaneous with those in Glen Alladale and Gleann Mor to the north, which terminate at similar altitudes. The moraines can be traced up the valleys toward an upper basin at approximately 550 m a.s.l. Thick peat has accumulated in the basin and glacial landforms are sparse. However, it is likely that the basin once acted as a source area for glacier ice in the valleys below, particularly given the evidence for plateau-sourced ice to the north.

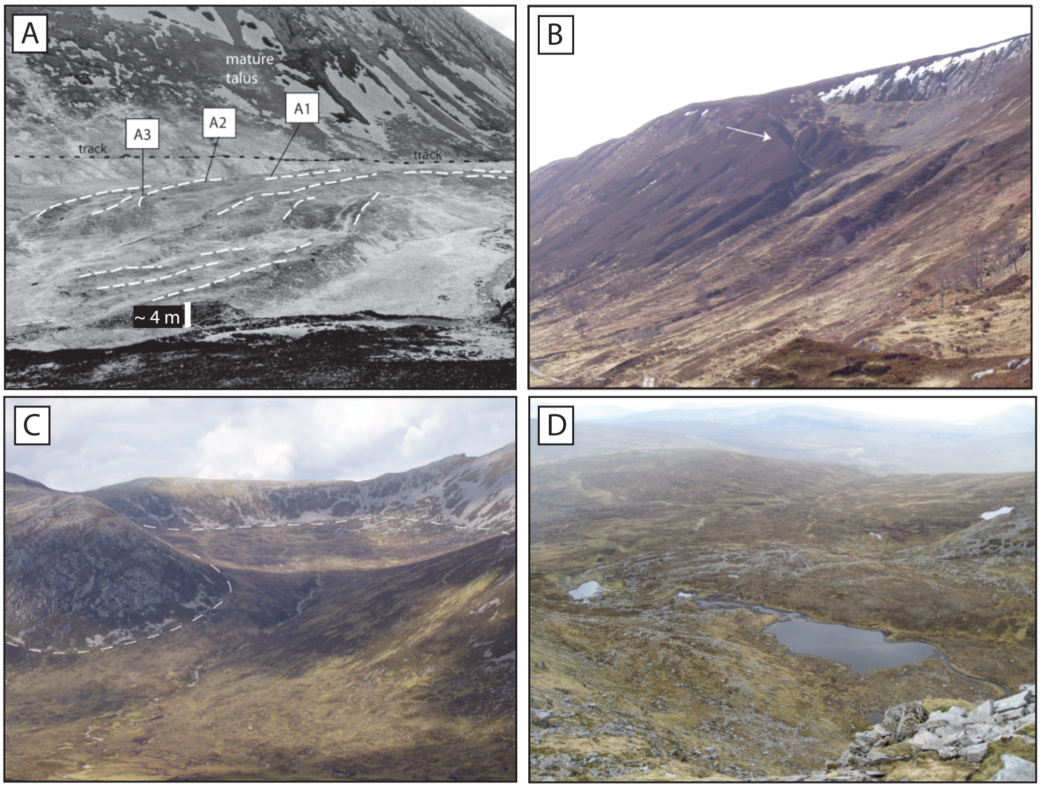

FIGURE 3.4: Field evidence for former glacier configurations: A. Multi-crested moraine ridge complex in the northern tributary to Glen Alladale (locations of samples A1, A2, and A3 are shown). B. Lateral moraine merging with plateau surface on the northern side of Gleann Mor. C. Down-slope limit of talus in the northern corrie of the Mhucarnaich valley, suggesting the minimal height of a former ice surface. D. Abruptly terminating limit of boulder moraines in the west-facing Coire an Lochain Sgeirich.

3.4.3 Southern valleys

Landforms around the southern valleys of the Beinn Dearg massif are relatively well documented [Kirk et al., 1966; Reed, 1988; Bennett and Boulton, 1993; Gordon, 1993b], and are briefly summarised here. A large (∼ 200 m wide, 800 m long), multi-crested, end moraine (Cnoc a' Mhoraire) with a distal outwash fan, dams Loch a' Gharbhrain (Fig. 3.3). It is immediately succeeded up the eastern valley side by closely-spaced, well-preserved, lateral and recessional moraines, which are generally composed of loose, sandy diamicton. The moraine probably marks a former readvance or still-stand margin, and morphologically resembles a hill-hole pair [cf Evans and Benn, 2001; Evans and Wilson, 2006]. The Gharbhrain moraine can be traced towards a conspicuous lateral moraine ~2 km to the northwest, allowing good delineation of the former glacier snout.

A clear down-valley termination of closely spaced, hummocky, recessional moraines indicates a former glacier margin at ~420 m a.s.l. in the Mhucarnaich valley. A sharp, lateral moraine trending up the south-western valley side from 460 m a.s.l. to 540 m a.s.l. approximates the former glacier surface. Clear down-slope limits of mature talus infer the dimensions of a source area in the corrie to the north (Fig. 3.4C), while ice-smoothed bedrock in the north-western col suggests former occupation by active glacier ice (Fig. 3.3).

A terraced outwash fan, approximately 0.4 km^2 in area, occupies the mouth of Strath Vaich, originating at a cross-valley terminal moraine at ~230 m a.s.l. Rising lateral moraines and associated meltwater channels continue north, up the valley from the terminal moraine, indicating a coherent former glacier margin. The valley floor is occupied by morainic mounds, an esker, and isolated terrace fragments which disappear a short distance up the valley from the terminal moraine. Larger, bouldery moraines, chaotic mounds, and boulder-scattered bedrock outcrops are present beyond the outwash fan, in the confluence of Strath Vaich and the Glascarnoch River, perhaps indicating a former area of in situ decay, as ice margins detached and retreated away from the confluence. Closely-spaced, hummocky, recessional moraines and lateral moraine fragments are present along the sides of Loch Vaich and a dense moraine assemblage occupies the valley head. Slopes here are affected by debris flows, which have incised sediment-mantled valley sides. Higher up the eastern valley sides, a transition (between 600 and 650 m a.s.l.) to relict bouldery solifluction lobes, then a summit blockfield at Beinn a' Chaisteil is apparent. This transition indicates the upper limit of the most recent episode of glacial modification.

A well-formed lateral and end moraine assemblage indicates a former glacier limit at the mouth of Strath Rannoch. Moraines and marginal meltwater channels continue up the valley, where a particularly clear lateral moraine can be traced continuously for almost 2.5 km, up onto higher ground to the east.

3.4.4 Western corries

Three corries occupy the western side of the massif (marked 'C' on Fig. 3.3), and are characterised by relatively high (650-700 m a.s.l.) corrie floors partially surrounded by steep cliffs, but with open cols linking to the plateau. These corries were considered by Sissons [1977] and Reed [1988], to have been occupied by glaciers during the YD. Prominent boulder moraines (e.g. Fig. 3.4D) are present in each, terminating abruptly within ∼2 km of the corrie heads. The angular nature of the boulders suggests a rockfall source, and that blocks were transported supraglacially down the valley. Their abrupt down-valley termination provides strong evidence for glacier limits during the last phase of intense rockfall activity. Sparse glacial landforms exist in the few kilometres to the west of these boulder spreads, generally comprising isolated moraines and meltwater channels.

3.4.5 Central valley and plateau surface

An ∼ 16 km-long valley, comprising Gleann Beag and Gleann Mor, dissects the plateau of the Beinn Dearg massif. The valley is occupied by numerous suites of recessional, hummocky moraines, although their presence is not continuous. Debris cones, sourced from thick glacial sediment deposited on valley flanks, are frequent, and sparse esker fragments occur on the valley floor. The head of Gleann Beag (NH285 838) is occupied by conspicuous, broad, morainic accumulations (up to 50 m in width), and superimposed fluted sediment, streamlined in a north-eastward direction. Similar accumulations are present at up to 700 m a.s.l. at the head of the corrie of Loch Tuath (NH 280 823), a tributary to Gleann Beag. Here, broad mounds are fluted towards the east, and possess a flat, streamlined surface (Fig. 3.5), suggesting that they have been smoothed by eastward ice flow, subsequent to their initial deposition.

Plateau areas immediately above the headwalls of outlet valleys are commonly characterised by smooth, ice-moulded bedrock – most likely the consequence of increased basal abrasion under accelerating, wet-based, ice flowing steeply into the valleys. Flatter, and more central parts of the plateau do not possess such features, and in places are occupied by relict landforms, such as the preserved (possible pre-Quaternary) fluvial valley at NH 30 81 (Fig. 3.6). Higher ground, around the western summits of Beinn Dearg, Eididh nan Clach Geala, and Seana Bhraigh is characterised by mature periglacial features, such as blockfields and relict boulder sheets and lobes. Where such features occur above, but in close proximity to, glacially modified terrain (characterised, for example, by ice-smoothed bedrock or moraines), upper limits of the most recent phase of glacial alteration may be inferred.

Figure 3.5: Large sediment accumulations with streamlined upper surfaces at the valley head west of Loch Tuath. The sediment is interpreted to have originally banked up against a thinning ice cap margin which pushed into the valley (black arrow). Subsequent ice cap growth and ice flow from local centres of high ground (white arrow) led to smoothing and streamlining of the surface. Black broken line broadly defines the crests of the large sediment mounds

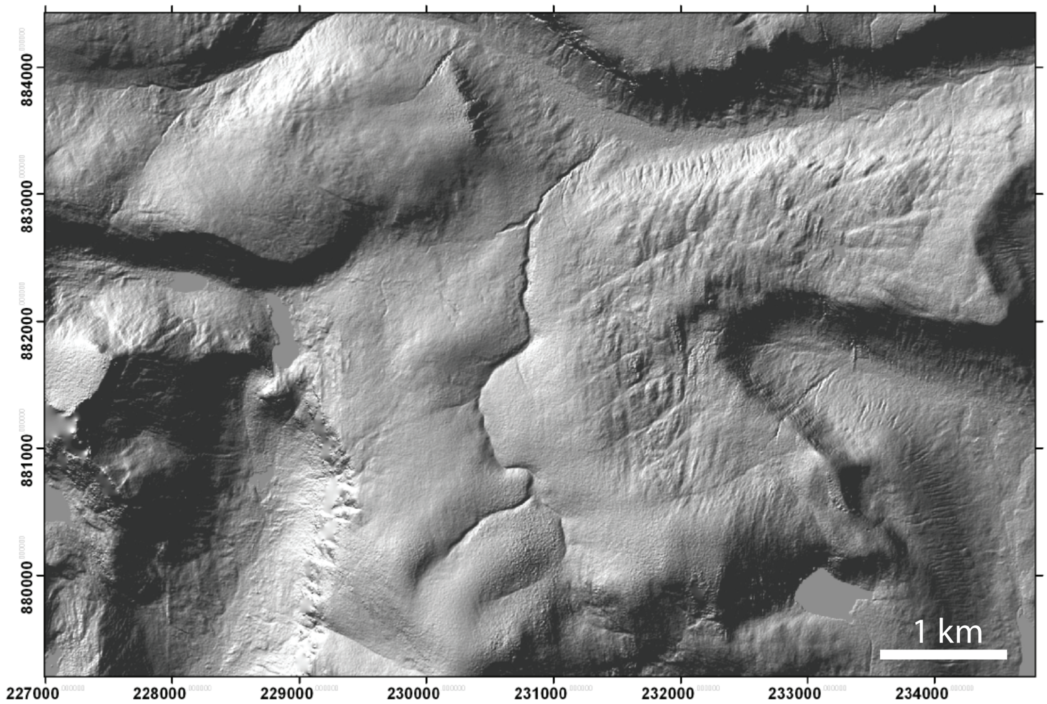

FIGURE 3.6: Meandering V-shaped valley on the plateau surface of the Beinn Dearg massif, between 600 m and 750 m a.s.l. This feature may have a pre-glacial origin. Hill-shaded digital surface model built from Intermap Technologies NEXTMap Britain topographic data. Northwest illumination.

3.4.6 Loch Droma and the northern Fannichs

Large bouldery moraines, sparse kettle holes, and meltwater channels extend west-north-westward down the valley from Loch Droma; these landforms document ice margin retreat back towards Loch Droma. In Coir' a Mhadaidh, $\sim$ 100 m-wide mounds arc up into a col, and can be traced back down valley towards lower ground (Fig. 3.3). At NH 237 749, a section revealing sheared, laminated, fine sands may indicate former ponding against an active ice margin which retreated out of the valley towards the northeast. A complex palimpsest moraine assemblage is present in the Ghiuthais valley (Figs 3.3, 3.7). Here broad, cross-valley moraine ridges that arc up into the valley, are overprinted in places by an assemblage of smaller, bouldery moraines, flutes, and meltwater channels, which descend down the valley to the northeast. The broad moraines suggest ice margin retreat out of the valley towards the north. In contrast, the latter morainic assemblage suggests an ice advance from high ground to the south, terminating at $\sim$320 m a.s.l.

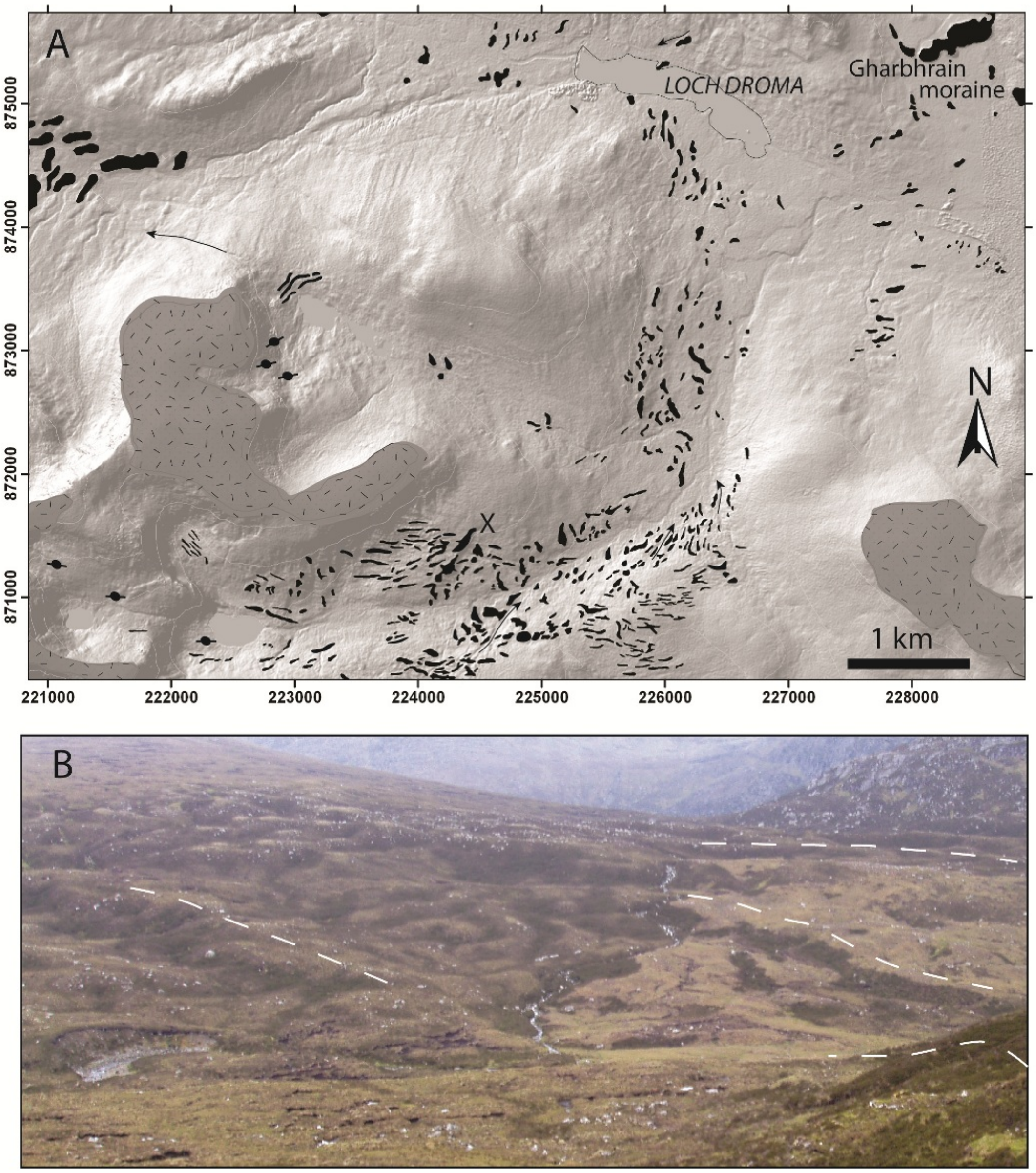

FIGURE 3.7: Palimpsest moraine assemblages in the Ghiuthais valley of the northern Fannich mountains: A. Geomorphological map (see Fig. 3.3 for legend). B. Photograph taken from point 'X' on map looking south-eastward. White broken line shows broad, cross-valley moraines.

3.5 Chronology

Combining geomorphological evidence to make inferences about former glacier configurations requires an assessment of landform age. Loch Droma is a well-known site where sediments of pre-YD age were discovered during excavations for a dam embankment [Kirk and Godwin, 1963]. A single, bulk sample from the base of an organic layer in the profile yielded a radiocarbon age of 12810 ± 155^{14}C a BP (15456 ± 900 cal. a

BP (2σ)(Calib 6.0, Stuvier and Reimer)). However, Kirk and Godwin [1963] stated that the radiocarbon age is probably at least 1000 years too old, and may have been affected by incorporation of older material, or by hard water error in an open water environment. Based on stratigraphic and pollen analyses Kirk and Godwin [1963] suggest that the sediments are of Allerød age (GI-1c to GI-1a: 14-12.9 cal. ka BP). Thus landforms to the west of Loch Droma, and those in the northern Fannichs indicative of northward ice margin retreat, are considered to be older than 12.9 cal. ka BP. Three boulders sampled from the Glen Alladale moraine complex (Fig. 3.4A) (Tables 3.1, 3.2), yielded overlapping exposure ages that are strongly suggestive of a YD age of deposition. The oldest sample, A2 (12.9 ± 1.6-12.4 ± 1.3 ka BP), suggests deposition at, or close to, the start of the YD. Limited, localised moraine degradation may have occurred in the vicinity of sample A3, closest to the moraine edge (Fig. 3.4A), giving an analytical uncertainty that does not overlap with the oldest sample at 1σ (Table 3.2), and a slightly younger minimum exposure age (11.0 ± 1.4-10.6 ± 1.0 ka BP). Importantly, the moraine complex is indicative of non-topographically constrained ice flow [Finlayson and Bradwell, 2007], thus suggesting a plateau ice source, and ice cap-style configuration [cf. Golledge, 2007] at the time of deposition.

Landforms can also be assessed on a morphostratigraphical basis [cf. Lukas, 2006], whereby particular landsystem components provide clues to landform age [Ballantyne, 1989, 2002, 2007a; Benn and Ballantyne, 2005; Finlayson, 2006; Finlayson and Bradwell, 2007; Lukas and Bradwell, 2010]. Suites of closely-spaced, 'hummocky', recessional moraines are commonly, although not always [e.g Clapperton et al., 1975; Everest and Kubik, 2006], observed within the limits of former YD glaciers which have been dated by other evidence [e.g Walker et al., 1988; Ballantyne, 1989]. Their presence in several valleys in the Beinn Dearg massif suggests the possibility of a YD age, an interpretation consistent with the cosmogenic exposure ages discussed above. Mature periglacial landforms can also be useful indicators of age. The last period of intense periglacial activity in upland Britain was the YD [Ballantyne and Harris, 1994]. The distribution of mature, relict, periglacial forms over upper parts of the Beinn Dearg massif may thus indicate areas that were not protected, or modified by, glacier ice at that time.

Landform overprinting also enables interpretation of relative age. The thick sediment mounds at the head of Gleann Beag and by Loch Tuath (Fig. 3.5) are fluted and, in places, planed off on their surfaces, suggesting that a later glacial event modified the mounds following their initial deposition. Our interpretation is that these mounds were originally deposited by eastward ice cap retreat and thinning (see below), allowing sediment to accumulate between the ice margin and reverse slope, prior to subsequent ice cap thickening, overriding and streamlining. Hättestrand et al. [2008] have reported morphologically similar features, termed 'cirque infills' on the Kola Peninsula in Russia. Abundant sources for the debris would have existed during ice cap thinning and retreat, including: rockfall, fluvial transport from the upper valley, and debris flows / fluvial transport from valley sides and ice margin. Smoothing and fluting of the mounds presumably occurred during subsequent (YD) glacier overriding. Similarly, moraine cross-cutting in the Ghiuthais valley of the northern Fannichs (Fig. 3.7), suggests that an initial phase of ice margin retreat towards the north was followed by renewed advance of ice from high ground in the south. The most likely period for this renewed advance would be the YD.

Sample (lab code)	Quartz (g)	Be carrier (mg)	$^{10}Be/^{9}Be$ [1,2] ($x10^{-15}$)	^{10}Be conc. [2,3,4] ($x10^{3}$ atoms g^{-1})	Min exp. age (ka): De [6]	Min exp. age (ka): Du [6]	Min exp. age (ka): Li [6]	Min exp. age (ka): Lm [6]
A1(b3306)	24.517	0.237	122.7±3.7	74.8±3.0	12.0±1.5	12.0±1.5	11.6±1.2	11.6±1.2
A2(b3307)	22.583	0.238	119.6±3.2	79.2±3.0	12.9±1.6	12.9±1.6	12.4±1.3	12.4±1.2
A3(b3085)	22.075	0.234	103.2±4.2	68.1±3.5	11.0±1.4	11.0±1.4	10.7±1.2	10.6±1.0

TABLE 3.2: Analytical data and exposure ages. [1]Isotope ratios were normalised to NIST SRM 4325 using $^{10}Be/^{9}Be = 3.06 \times 10^{-11}$ and a ^{10}Be half-life of 1.51×10^{6} years [Yiou and Raisbeck, 1972; Hoffmann et al., 1987; Inn et al., 1987]. [2]Uncertainties are reported at 1σ confidence level. [3]A mean procedural blank value of $10.9 \pm 1.6 \times 10^{4}$ ^{10}Be atoms ($^{10}Be/^{9}Be = 7.0 \pm 1.0 \times 10^{-15}$) was used to correct for the background. [4]Propagated uncertainties include error in the procedural blank, carrier mass (2%), and counting statistics. [5]Calculated with the CRONUS-Earth online calculator [Balco et al., 2008, http://hess.ess.washington.edu/): wrapper script version 2.2, main calculator version 2.1, constants version 2.2.1, muons version 1.1.]. [6]Results are provided for different time-dependent cosmogenic nuclide altitude/latitude scaling schemes according to Balco et al. [2008]): De [Desilets and Zreda, 2003; Desilets et al., 2006]; Du [Dunai, 2001]; Li [Lifton et al., 2005]; Lm [Lal, 1991; Nishiizumi et al., 1989; Stone, 2000]
.

3.6 The Younger Dryas Beinn Dearg ice cap

3.6.1 Reconstruction

Collectively, the evidence presented above allows an empirically-based, three-dimensional representation of the ice cap that occupied the Beinn Dearg massif during the YD (Figs 3.8A,B). The reconstruction is a synthesis based on the cumulative landform evidence, and thus represents the integrated maximum glacier extent in all individual outlets. However, it cannot be demonstrated that all outlet glaciers occupied their maximum positions simultaneously. Ice margins and surface contours were reconstructed following procedures outlined in previous studies [Sissons, 1974; Ballantyne, 1989, 2002, 2007a; Benn and Ballantyne, 2005]. Over high ground, such as the western summits of Beinn Dearg and Eididh nan Clach Geala, zones of mature periglacial deposits are considered not to have been covered by active ice, although a non-erosive, cold-based ice cover cannot be ruled out [Gordon et al., 1987; Gellatly et al., 1988; McDougall, 2001; Rea and Evans, 2003]. The preserved (possible pre-Quaternary) fluvial valley (Fig. 3.6) on the plateau surface is considered to have been covered by cold-based ice. This is supported by the distribution of glacially-modified landforms that show an outwardly radial pattern in the vicinity. The Beinn Dearg ice cap, as reconstructed here, has a total area of 176.2 km^2. One small additional (2.74 km^2) independent glacier (the Mhucarnaich glacier) is also reconstructed on the south-western corner of the massif.

The reconstruction is based solely on geomorphological evidence. To test the theoretical range of basal shear stresses required by the reconstruction, the 'Profiler v.2' spreadsheet of Benn and Hulton [2010] was used. The glacier reconstruction for Glen Douchary was used, as it includes the widest range of reconstructed glacier surface slopes (from a relatively 'flat' plateau area to a zone where steeply flowing ice fed into the valley below). The glacier profile and shear stresses, assuming perfectly plastic ice rheology, are shown in Figure 3.9. Low basal shear stresses (tending towards zero) were calculated on the plateau, consistent with observations of landform preservation over parts of the plateau. Shear stresses increased (up to 250 kPa) where ice accelerated, flowing steeply into the valley below (possibly as an ice fall). This is also consistent with landform evidence, where moulded bedrock was observed immediately above the valley headwall. In the valley, shear stress remained constant at $\sim$50 kPa. A slight reduction in reconstructed shear stress over 1.5 km at the glacier margin suggests that wet-based basal sliding or lubricated sediment deformation may have occurred.

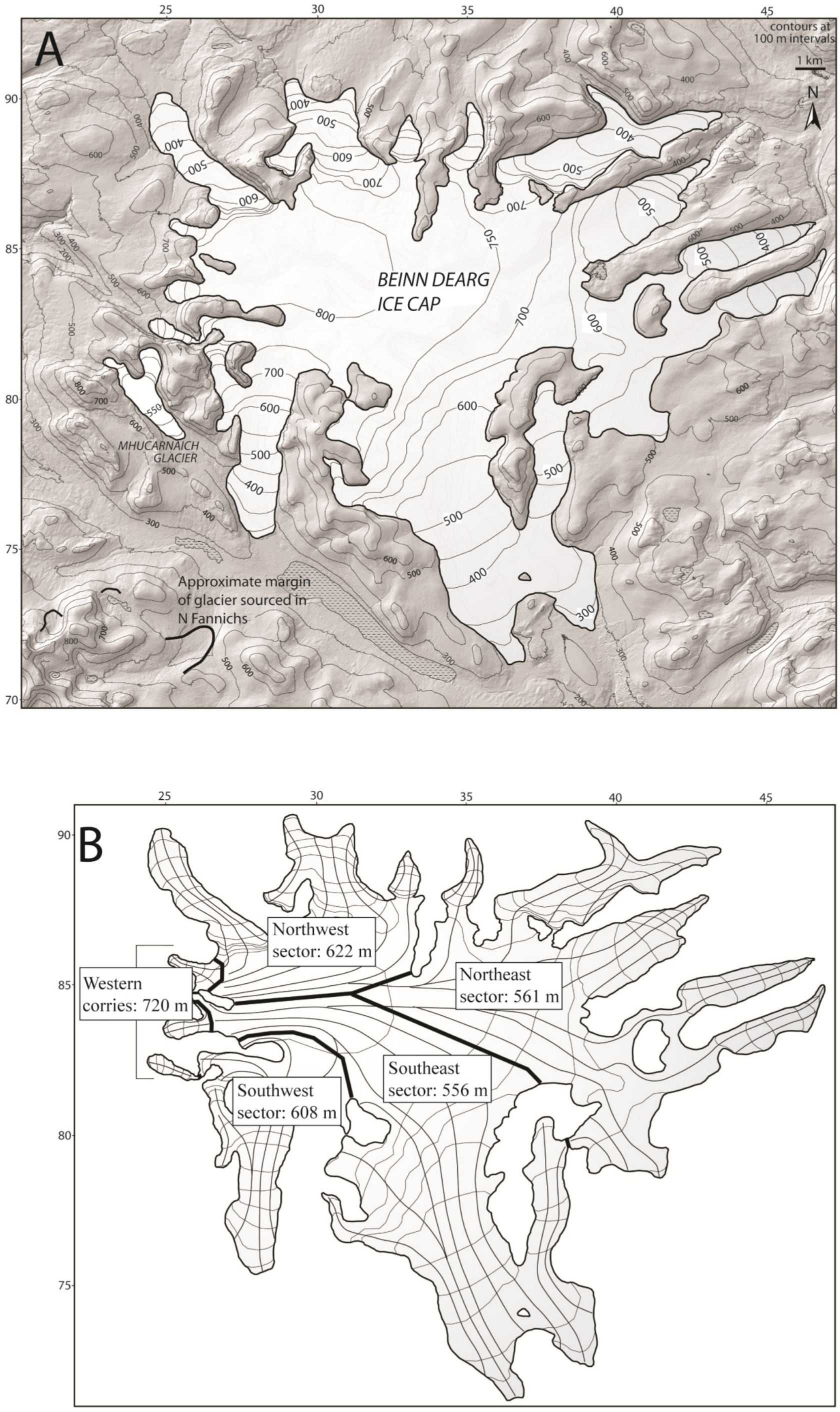

FIGURE 3.8: Three-dimensional reconstruction of the YD Beinn Dearg ice cap, based on empirical evidence. Note that the reconstruction is derived from the cumulative landform record; it cannot be demonstrated that all outlets were at their maximum position simultaneously. Ice surface contours at 50 m intervals. B. Ice cap reconstruction showing schematic flowlines and sectors used in ELA calculations. The ELAs shown are those calculated using the AABR (1.8) method described in Section 3.6.2.

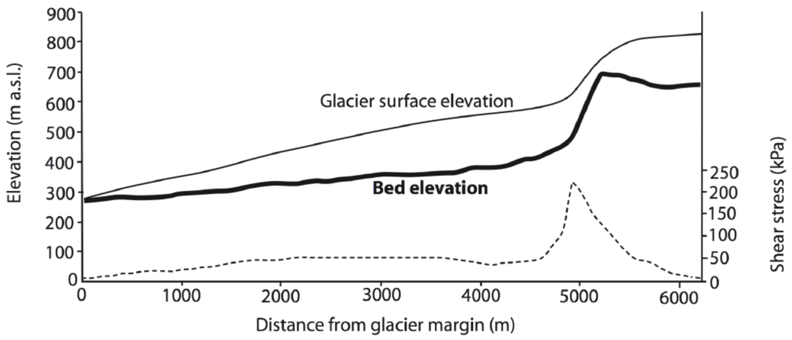

Figure 3.9: Glacier surface elevation and shear stress in Glen Douchary, calculated using the 'Profiler v.2' spreadsheet of Benn and Hulton [2010].

3.6.2 Equilibrium line altitudes (ELAs)

Equilibrium line altitudes (ELAs) on ice masses indicate theoretical lines where annual accumulation and ablation are balanced. Mass balance studies on modern ice masses have shown that ELAs are strongly linked to regional ablation-season temperature and accumulation-season precipitation, providing a link between glaciers and climate. For comparison with data from elsewhere in Scotland, we present ELAs calculated by the area weighted mean altitude (AWMA), accumulation area ratio (AAR), and area altitude balance ratio (AABR) methods (Table 3.3). The AWMA method assumes identical, linear accumulation and ablation gradients, and thus tends to overestimate ELAs, as ablation gradients are often steeper in reality. The AAR method can account for net ablation occurring over a smaller proportion of an ice mass, but does not consider ice mass hypsometry [see Osmaston, 2005, for review]. AABR methods are generally assessed to be more reliable [Ballantyne, 2002; Benn and Ballantyne, 2005; Osmaston, 2005], and incorporate an assumed balance ratio, defined as ablation gradient/accumulation gradient (thus a balance ratio of 1 yields an ELA equivalent to that of the AWMA method). Previous Scottish studies use balance ratios of between 1.67 and 2.0 [Ballantyne, 2002; Benn and Ballantyne, 2005; Finlayson, 2006; Lukas and Bradwell, 2010] – a range that includes AABRs considered to be representative of modern mid-latitude maritime glaciers [Rea, 2009]. Using these values the AABR method yields ELAs of between 570 m and 580 m for the reconstructed Beinn Dearg ice cap as a whole. As the ice cap was drained radially, its overall ELA gives a good approximation of the regional temperature-precipitation dependent ELA, independent of the effects of deflation and snow drifting [Dahl and Nesje, 1992]. In order to evaluate spatial variations in ELA within the ice cap, it was divided into sectors based upon former ice flow direction (Fig. 3.8B). A general eastward decline in ELA across the ice

Ice mass	Area	ELA					
		AWMA	AAR (0.5)	AAR (0.6)	AABR (1.67)	AABR (1.8)	AABR (2.0)
Western corries	2.6	739	720	698	722	720	716
Northwest sector	28.2	666	725	610	628	622	614
Northeast sector	64.0	593	560	538	565	561	555
Southeast sector	63.2	599	575	528	562	556	549
Southeast sector	18.2	650	700	679	613	608	600
Beinn Dearg Icecap	176.2	616.2	598	548	581	567	569
Mhucarnaich Glacier	2.74	586.9	555	542	570	568	565

TABLE 3.3: Areas and equilibrium line altitudes (ELAs) for the empirically reconstructed YD Beinn Dearg ice cap and independent Mhucarnaich glacier. AWMA = area-weighted mean altitude; AAR = accumulation area ratio; AABR = area altitude balance ratio. AABR values calculated using the spreadsheet of Osmaston [2005].

cap is evident – a characteristic common to reconstructed YD ice masses in Scotland [e.g. Ballantyne, 1989, 2002; Benn and Ballantyne, 2005]. The eastward decline in ELAs is probably a result of eastward snow redistribution across the ice cap by strong westerly winds during the YD.

Previous studies have used glacier reconstructions to make palaeoclimatic inferences, based upon relationships between total precipitation and summer temperature at the ELA [Benn and Ballantyne, 2005; Finlayson, 2006; Lukas and Bradwell, 2010]. These studies have favoured a non-linear relationship, derived from a global sample of modern glaciers by Ohmura et al. [1992], expressed as:

$$P = 9T_a^2 + 269T_a + 645 \tag{3.1}$$

where P is the sum of winter accumulation (bw) plus summer precipitation at the ELA, and Ta is the mean ablation season air temperature at the ELA (usually derived from biological proxies and extrapolated to the location under investigation). However, recent studies by Golledge [2008] and Golledge et al. [2010] have shown that use of such a global dataset neglects the regional influence of seasonality upon former Scottish ice masses during the YD, and may have resulted in over-estimates of palaeoprecipitation. As a result, Golledge et al. [2010] used numerical model output to propose an alternative precipitation-temperature function specific to the Scottish YD environment. The alternative function allows for an annual temperature range (mean July T − mean Jan T = 30°C) that was three times greater than today, and so is more aligned with evidence from some palaeoclimatic proxies [Atkinson et al., 1987; Isarin et al., 1998; Witte et al., 1998; Lie and Paasche, 2006]:

$$P = S(14.2T_a^2 + 248.2T_a + 213.5) \tag{3.2}$$

S is a scaling coefficient allowing for seasonality in precipitation ($S = 1$ for neutral precipitation seasonality where daily precipitation = 1/365$*$ total annual precipitation; $S = 1.4$ for summer dominated precipitation; and $S = 0.8$ for winter dominated precipitation; Golledge et al. [2010]. It is also useful to consider a function that assumes a modern maritime annual temperature range (mean July $T-$ mean Jan $T = 10$°C; Golledge [2008], which is derived from the original modelling experiments presented by Golledge et al. [2008]:

$$P = S(25.3T_a^2 - 4.7T_a = 17.9) \tag{3.3}$$

Using an ablation-season sea level palaeotemperature of 6.38°C [Golledge, 2008] for the coldest part of the YD in northwest Scotland, and an assumed lapse rate of 0.006°C m^{-1}, palaeoprecipitation values at the ELA can be calculated using equations 3.1, 3.2, and 3.3 (Table 3.4). A variety of climatic regimes is represented, clearly highlighting the range of scenarios for seasonality that could have resulted in a mass balance permitting the existence of the YD Beinn Dearg ice cap.

Function	**Ohmura et al. [1992] Eq. 3.1**	Golledge et al. [2010]: 30°C annual temp. range, Eq. (3.2)			Golledge [2008]: 10°C annual temp. range, Eq. (3.3)		
		Summer-type P	Neutral-type P	Winter-type P	Summer-type P	Neutral-type P	Winter-type P
P (mm a^{-1})	1587	1485	1061	849	3184	2319	1886

TABLE 3.4: Possible icecap ELA palaeoprecipitation values calculated using Equations 3.1, 3.2, and 3.3. An icecap ELA of 576 m (AABR = 1.8) is used.

3.6.3 Comparison with model simulations

Numerical modelling experiments aimed at simulating a 'best fit' to accepted YD glacier limits in Scotland [Golledge et al., 2008] produced a YD ice cap over the Beinn Dearg massif with a similar overall configuration to the empirically reconstructed ice cap presented here (Fig. 3.10), although key differences are apparent, as discussed below. The numerical simulation assumed initial ice-free conditions, and was forced by a GRIP-pattern temperature depression (10°C lowering of annual temperature at YD maximum). Seasonality was considered to follow present (maritime) annual temperature ranges, and modern precipitation values were invoked over western Scotland, but with 60% northward and 80% eastward reductions [see Golledge et al., 2008, for further discussion]. In subsequent experiments, Golledge et al. [2010] and Golledge [2010b] found that similar modelled glacier configurations across Scotland could equally be produced with doubled or trebled seasonality, combined with reduced total precipitation, and more relaxed precipitation gradients; thus demonstrating that different climatic regimes can produce similar overall ice mass dimensions, albeit with different glaciological conditions.

For the purposes of this discussion, comparisons are made with the most fully described simulation, reported in Golledge et al. [2008, 2009]. The numerical simulation produced an overall ice cap configuration, drained radially by valley outlets, with a summit (∼ 900 m a.s.l.) in the west, and glacier ice in the elevated basin in the south-eastern corner of the massif. A similar overall configuration is empirically reconstructed here. The numerical model simulates negligible basal velocities and low (-2 to -4.5°C) basal temperatures over much of the plateau [Golledge et al., 2009], with the exception of 'drawdown' zones which feed valley outlet glaciers. This is also consistent with the empirical evidence. Key differences are that the numerical simulation produced more extensive western glacier limits and less extensive eastern glacier limits (generally ±1-5 km). We now discuss some possible reasons for these discrepancies.

Possible underestimates in the empirical reconstruction for western areas may arise due to preferential production/preservation of landforms during latter parts of the stadial. In eastern outlets of the Beinn Dearg range, over-estimation using empirical evidence may occur by mistakenly assigning a YD age to older landforms. However, the available dating evidence suggests that the empirical reconstruction is correct. Since the numerically modelled ice cap impinges upon Loch Droma (known to have been ice-free during the YD) in the southwest, and does not reach the dated limits in Glen Alladale (Figs 3.3, 3.4A)(Table 3.2), it is considered to be in error in those areas. Although increased dating control is desirable, the generalised conclusion made for the

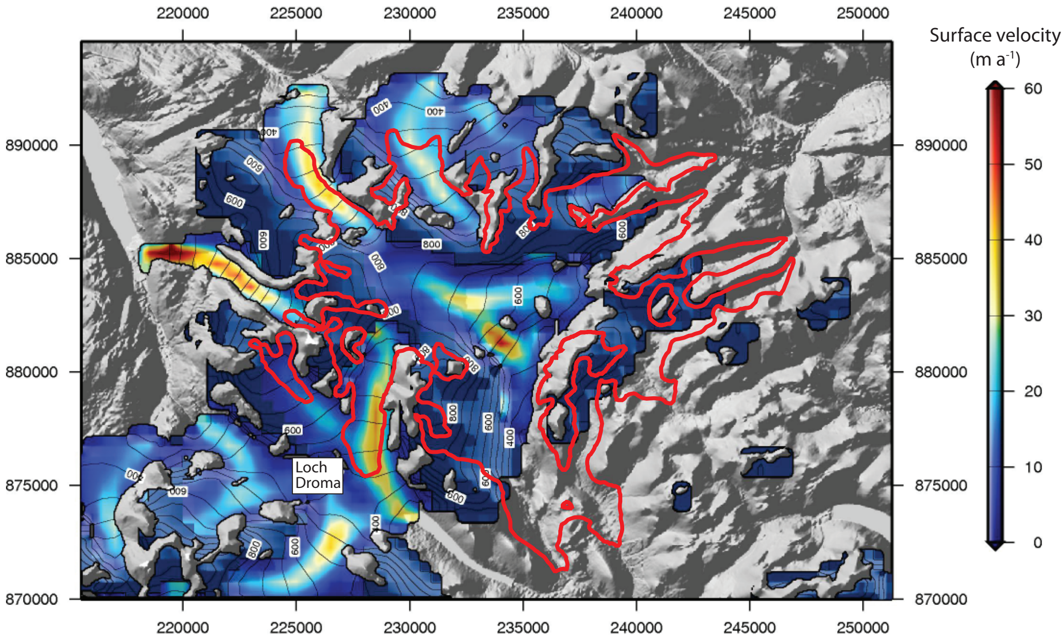

FIGURE 3.10: Comparison of empirical ice cap reconstruction (red line, this study) with numerical simulation (colour ramp, Golledge et al. [2008]) Note that the numerical simulation was originally run at 500 m resolution. The output presented here is interpolated to 50 m.

Beinn Dearg range, is that the numerical model over-predicts western glacier extent, and under-predicts eastern glacier extent. Golledge [2008] indicates that accumulation in the model does not take into account mass redistribution by wind, and suggests that this may introduce a degree of error. As the overall configuration and catchments of the empirically reconstructed and modelled ice caps are similar (Fig. 3.10), it is useful to compare the spatial variations in their ELAs (Table 3.5). Modelled ELAs are obtained by identifying cells in glacier flow lines where net mass balance = 0. A general eastward rise in ELA is apparent for the numerical simulation – largely a consequence of the precipitation gradients that were imposed to fit the regional (Scotland wide) ice distribution. In contrast, an eastward ELA decline is calculated across the ice cap from the empirical reconstruction. This supports the contention that eastward wind redistribution influenced the ice cap's mass balance at a local scale. To gauge the required mass transfer (δb_w) that would account for the ELA differences (Δh), the relation given by Hooke [2005] can be used:

$$\frac{\partial b_w}{\partial z}\Delta h + \delta b_w = \frac{T}{L}\left[\frac{\partial R}{\partial z}\Delta h + \delta R + \gamma\left(\frac{\partial T_a}{\partial z}\Delta h + \delta T_a\right)\right] \tag{3.4}$$

$\partial b_w/\Delta z$ is the winter balance-elevation gradient (kg m^{-2} m^{-1}). T is the length of the

Sector	Empirically reconstructed ELA (m)	Modelled ELA (m)	Δh (m)	δb_w (kg m^{-1})	Proportion of winter P (%)	Proportion of total P (%)
Western corries	720	590	-130	-492	-35.9	-21.2
Northwest sector	622	555	-67	-253	-18.5	-10.9
Southwest sector	608	600	-8	-30	-2.2	-1.3
Northeast sector	561	615	+54	+204	+14.9	+8.8
Southeast sector	556	610	+54	+204	+14.9	+8.8

Table 3.5: ELA variations between the empirical reconstruction and numerical simulation (Δh), and representative change in winter balance (δb_w), winter precipitation, and total precipitation. Note, these calculations carried out based on output from the model of Golledge et al. [2008], assuming a modern annual temperature range (mean July T - mean Jan T = 10°), and neutral precipitation seasonality.

melt season in days (~150 days close to the former ELA elevation, derived from an annual sinusoidal temperature variability curve [Golledge, 2008]), L is the latent heat of fusion of ice (334 kJ kg^{-1}), R is net radiation (MJ m^{-2} d^{-1}), γ is a constant of proportionality (suggested to be 1.7 MJ m^{-2} d^{-1} K^{-1}, for glacier ice [Kuhn, 1989]), and T_a is mean ablation season air temperature. Based on modern data from northwest Scotland, Ballantyne [2002] calculated a proportional increase in precipitation of 5.8% for every 100 m of elevation gain. Since the numerical model being compared here is based on a 10°C annual temperature range, combining the neutral-type P value of Equation 3.3 (Table 3.4) (adjusted to represent winter balance over ~215 days) with the precipitation gradient calculated by Ballantyne [2002], allows a winter balance-elevation gradient of 0.8 mm m^{-1}, or 0.8 kg m^{-2} m^{-1} to be approximated. Assuming that Δh is not due to differences in R or T_a ($\delta R = \delta T_a = 0$), and that $\partial R/\partial z = 0$ (as radiation input does not vary significantly with elevation [Hooke, 2005]), Equation 3.4 can be solved to give values that could account for the mismatch between modelled and empirically reconstructed ELAs (Table 3.5).

Attributing the required change in δb_w solely to wind redistribution, would require ~36% of winter snowfall to be stripped from catchments of the exposed western corries. The similarly exposed northwest sector, and relatively less exposed southwest sector, would have to lose ~19% and ~2% of their winter snowfall, respectively. The north-eastern and south-eastern sectors of the ice cap would both need to gain an additional ~15% to their winter snowfall from wind redistribution. Further work involving snow redistribution modelling is desirable in order to build upon these initial simple comparisons. However, the above discussion highlights the potential importance of wind redistribution when modelling ice caps at a more local scale.

A further source of error that may have contributed to the mismatches in eastern glacier extents is the possible survival of glaciers throughout the interstadial in eastern catchments. Although difficult to quantify, any surviving glaciers would have become incorporated within the ice cap that re-grew during the YD. However, glacier survival would have become relatively less important if the ice cap began to approach equilibrium.

3.7 Ice cap retreat prior to the Younger Dryas readvance

Using moraines to approximate former ice-marginal positions, retreat of the Beinn Dearg ice mass prior to the YD can be partially reconstructed (Fig. 3.11). Given the location of these former ice fronts (inside moraines of GI-1 age, see Fig. 3.1), they are considered to represent ice cap retreat during GI-1. Landform evidence from the north, west, and south suggests ice margin retreat back towards the Beinn Dearg massif, indicating that it acted as an important source area at that time. In contrast, ice margin retreat in the northern Fannich mountains was out of the valleys (away from local high ground) towards the Beinn Dearg massif. This demonstrates that the limited proportion of high ground in the Fannich mountains (characterised by sharp summits, with little plateau area), was insufficient to sustain glacier ice in those northern valleys throughout GI-1, and suggests that the Fannich mountains may have acted as an 'unzipping' corridor during deglaciation. An important inference is that valleys of the northern Fannich mountains had completely deglaciated by the Allerød (GI-1a to GI-1c) (when ice-free conditions are suggested by the Loch Droma sediments), and that any YD glacier growth in the northern Fannichs [Sissons, 1977; Bennett and Boulton, 1993] was therefore from initial ice-free conditions.

The large, streamlined sediment mounds present in upper valleys in western parts of the Beinn Dearg massif are of importance (Fig. 3.5). They are interpreted here as primarily ice marginal accumulations, formed as sediment banked up between ice margins and reverse slopes during retreat across the Beinn Dearg massif, prior to subsequent YD ice cap thickening and streamlining. Interpreted in this way, they demonstrate that active ice eventually thinned to below the height of the plateau in the western massif during GI-1, and that glacier source areas during the latter part of the interstadial (~13.5 cal. ka BP) were towards central or eastern parts of the massif.

This interpreted pattern of late-stage ice margin decay towards the east, contrasts to the modelled YD ice build up pattern (centred on the watershed in the west), where an ice-free interstadial is assumed [Golledge et al., 2008]. We speculate that glacier ice persistence in the Beinn Dearg massif would have been most likely in high, or shaded, central and eastern localities, where interstadial mass turnover and response times might have been slowest. Had glacier ice survived throughout GI-1 in those places, it

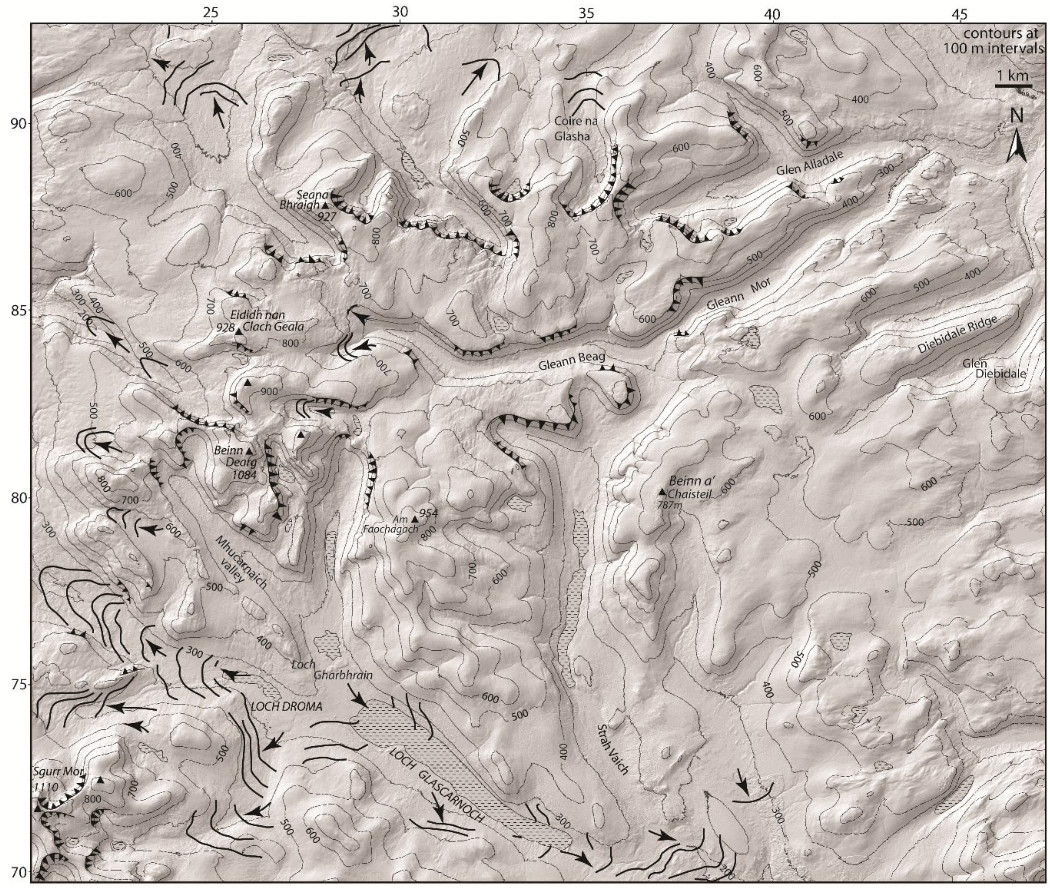

FIGURE 3.11: Reconstruction of former ice fronts (black lines) prior to the YD, based on this study. Black arrows indicate former ice flow direction.

could then have become incorporated within the ice cap that re-grew during the YD – providing a possible additional contribution to explaining the discrepancy between the empirical reconstruction and numerical simulation in the east. The oldest exposure age from the Glen Alladale moraine (Table 3.2) suggests deposition near the start of the YD, lending some support to the concept of glacier ice survival in the east.

3.8 A conceptual model of Lateglacial mountain ice cap evolution in northern Scotland

In synthesising the above, it is useful to present a simple conceptual model of ice cap evolution in the region during the GI-1 to YD transition of the Lateglacial (Fig. 3.12). Ice cap thinning, and marginal retreat towards key source areas, such as the Beinn Dearg massif, probably occurred during GI-1, following dated ice sheet oscillations along the Scottish north-western seaboard [Bradwell et al., 2008a; Ballantyne et al., 2009] (Fig. 3.1). Ranges, such as the Fannich mountains, with little accumulation area, acted as 'unzipping' zones, accompanied, in places, by deposition of broad ice marginal landforms, which arc into their valleys (Fig.3.12A). Loch Droma became ice-free by the Allerød (GI-1a to GI-1c) (within error of the dated ice sheet limit in Wester Ross, *c.* 14 cal. ka BP [Ballantyne et al., 2009]), suggesting that relatively rapid ice margin retreat from low-lying coastal areas occurred. Reduced sea ice cover would have allowed near-modern precipitation values at that time, enhancing mass turnover, melt, and debris evacuation, and leading to localised formation of large morainic accumulations and ice-contact glaciofluvial topography. As the ice mass thinned over the Beinn Dearg massif, plateau areas in the west became ice-free first (Fig. 3.12B), with the possibility that more slowly responding glacier ice survived in less maritime, eastern parts of the massif. At that time the northern Fannich mountains had probably completely deglaciated. It is not known how much ice survived in the Beinn Dearg massif during late stages of GI-1. However, cooler temperatures at the end of GI-1 and during the YD, resulted in re-growth of the Beinn Dearg ice cap, with the summit now positioned over the western part of the massif (Fig. 3.12C). Strong westerly winds [Brauer et al., 2008] probably redistributed snow from western to eastern parts of the ice cap, resulting in relative lowering of ELAs in the east. The lateral extent of the ice cap was more limited compared to earlier stages of the preceding GI-1. In the northern Fannich mountains glaciers probably grew afresh. Preservation of landforms interpreted as pre-YD in age, combined with the profile of the ice cap, suggests that motion was predominantly by internal deformation over the plateau, and locally by basal sliding (with perhaps some bed deformation) in outlet areas [Golledge et al., 2009]. Retreat of the YD ice cap left regular, closely-spaced recessional moraine assemblages in many outlet valleys, indicative of active, oscillatory retreat [Benn and Lukas, 2006; Ballantyne, 2007b] or high mass loss, due to enhanced surface melting supplying ice-marginal debris flows during retreat [Golledge, 2010b]. The resultant landscape (Fig. 3.12D) now encompasses both regular recessional moraine assemblages and palimpsest assemblages, indicative of changes in ice mass configuration during the Lateglacial period. Testing this model would enhance our understanding of glacier dynamics during the rapid environmental changes characteristic of the Lateglacial period, and provide further insight into ice cap response during transitional warm phases, such as GI-1. We propose that future work should include targeting eastern margins of former Scottish ice caps for cosmogenic exposure dating, in order to refine empirical reconstructions and guide numerical glacier simulations of the Lateglacial period.

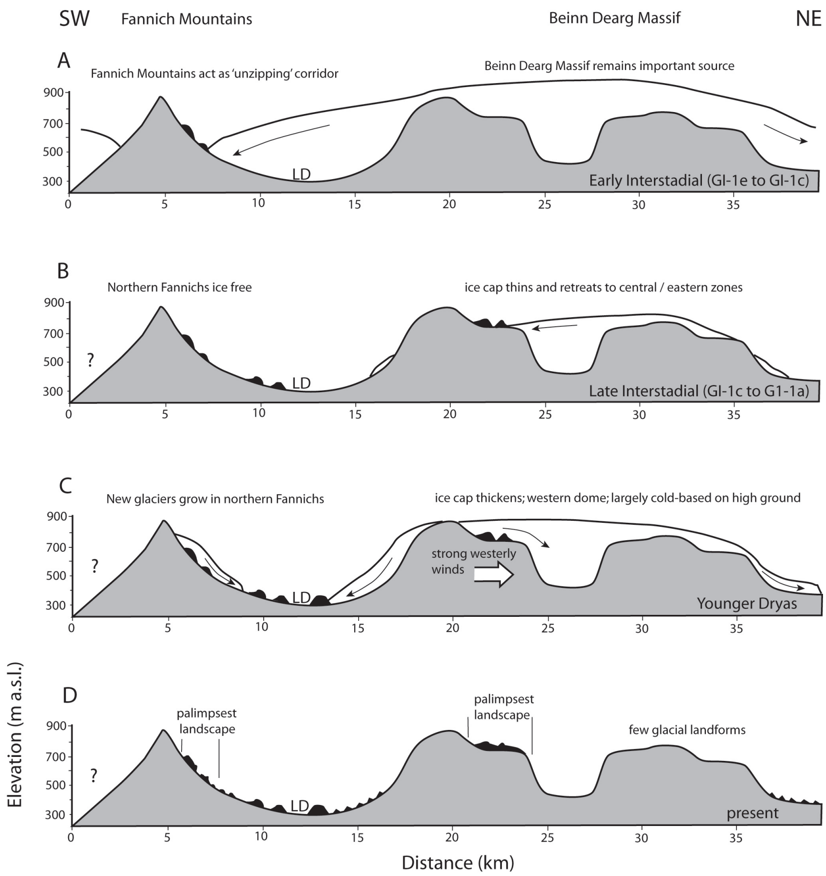

FIGURE 3.12: Schematic cross section through the Beinn Dearg massif showing proposed ice cap evolution during the Lateglacial period. See text for details. LD = Loch Droma.

3.9 Conclusions

Detailed geomorphological mapping of the Beinn Dearg massif in northern Scotland has enabled reconstruction of the evolution of a former ice cap during the Lateglacial period (c 14.7-11.7 cal. ka cal BP). The age of glacial landforms is constrained by previously recorded interstadial sediments at Loch Droma [Kirk and Godwin, 1963]), new cosmogenic exposure ages from a north-eastern outlet of the Beinn Dearg massif, and morphostratigraphic assessment. Based on empirical evidence, a 176 km^2 ice cap is reconstructed over the Beinn Dearg massif during the YD. The ice cap was centred over the western side of the Beinn Dearg massif and ELAs (AABR method) are calculated to have been between ~570 and 580 m a.s.l. over the ice cap as a whole.

The empirical reconstruction bears many similarities to a modelled ice cap produced in a recent numerical simulation of YD glaciation in Scotland [Golledge et al., 2008]. However, differences are apparent, predominantly in the form of glacier ice extent in western and eastern outlets. Comparison of empirically reconstructed and modelled

ELAs for different ice cap sectors reveals contrasting patterns. As the numerical model did not allow for wind redistribution, one explanation for these contrasts is that winter balance over western sectors was reduced (2%-36% loss depending on exposure) by deflation. Eastern sectors could have gained an additional 15% to their winter snowfall from deposition of blown snow. Differences in eastern glacier extent may also be complicated by the assumption of initial ice-free conditions in model runs. Significantly, however, both empirical and numerical reconstructions concur on the style and scale of glaciation in this area.

Landform evidence indicative of ice cap configuration prior to the Younger Dryas readvance suggests that the Beinn Dearg massif was an important source during interstadial ice cap retreat. In contrast, the neighbouring Fannich mountains acted as an 'unzipping' zone [cf. Greenwood and Clark, 2009; Clark et al., 2012], and had probably completely deglaciated on their northern side by the Allerød (GI 1c-1a). As the ice cap thinned over the Beinn Dearg massif, plateau areas in the west became ice-free, with the possibility that more slowly responding glacier ice survived in sheltered, central and eastern parts of the massif. Renewed cooling in the YD enabled regrowth of glaciers in the northern Fannich mountains from ice-free conditions, and growth of a thicker ice cap over the Beinn Dearg massif.

Chapter 4

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

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Abstract

A 3700 km^2 area adjacent to the Firth of Clyde, Scotland, is examined to constrain the development and dynamics of the western central sector of the last British and Irish Ice Sheet. Results from geomorphological mapping, lithostratigraphic investigations, three-dimensional geological modelling and field observations are combined to produce an empirically constrained, five-stage conceptual model of ice sheet evolution. (A) Previously published dates on interstadial organic deposits and mammalian fossils suggest that the Main Late Devensian (MLD) (MIS 2) glaciation of central Scotland began after 35 ka cal BP. During build-up, ice advanced from the western Scottish Highlands into the Clyde and Ayrshire basins. Glaciomarine muds and shelly deposits scavenged from the Firth of Clyde were redeposited widely as shelly tills and glacial rafts. Ice advance against reverse slopes generated, and subsequently overtopped, ice-marginal sediment accumulations. We hypothesise that some of these formed pre-cursor ridges which were moulded into suites of ribbed moraine during the glacial cycle. (B) Sustained stadial conditions at the Last Glacial Maximum (LGM) (c. 30 - 25 ka cal BP) resulted in development of a major dispersal centre over the Firth of Clyde and Southern Uplands. This dispersal centre locally preserved previously-formed subglacial bedforms, and fed a wide corridor of fast-flowing ice east towards the Firth of Forth. (C) Initial deglaciation promoted a substantial re-configuration of the ice surface, with enhanced westward drawdown into the outer Firth of Clyde and eastward migration of an ice divide towards the Clyde-Forth watershed. (D)

Renewed ice sheet thickening over the Firth of Clyde may have accompanied growth of the Irish Ice Sheet during the Killard Point Stadial (c. 17.1 - 15.2 cal ka BP); it was associated with limited bed modification over west central Scotland. Subsequent ice sheet retreat was characterised by substantial meltwater production, ponding and erosion. (E) Late stages of MLD ice sheet retreat were punctuated by one or more significant ice margin oscillations. Discovery of De Geer moraines at the site of a former proglacial lake in western Ayrshire allows glacier flow at the ice margin to be approximated as $\leq$290 m a^{-1} during one such oscillation. Such velocities were probably enabled by basal sliding and shallow sediment deformation. At this stage those parts of the MLD ice sheet margin that were grounded in the Firth of Clyde were extremely vulnerable to collapse. Final disintegration of glacier ice in the Clyde basin probably occurred early in the Lateglacial Interstadial (Greenland Interstadial-1), coinciding with marine incursion to $\sim$40 m above present day sea level.

4.1 Introduction

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

This study attempts to address these issues for west central Scotland (Figs 4.1, 4.2) – an area of some 3700 km^2 which was subjected to interactions between major accumulation zones of the last BIIS. By combining geomorphological, lithostratigraphical, and three-dimensional geological modelling investigations with existing research, we reassess the palaeoglaciology of this formerly dynamic ice sheet zone. Specifically, this paper aims to: (i) identify evidence for spatially and temporally variable ice flow patterns, major geometry changes, and oscillatory events that accompanied build up and decay of the BIIS in west central Scotland; (ii) take account of published evidence from surrounding areas to test for indicators of more widespread ice sheet reorganisation(s); and (iii)

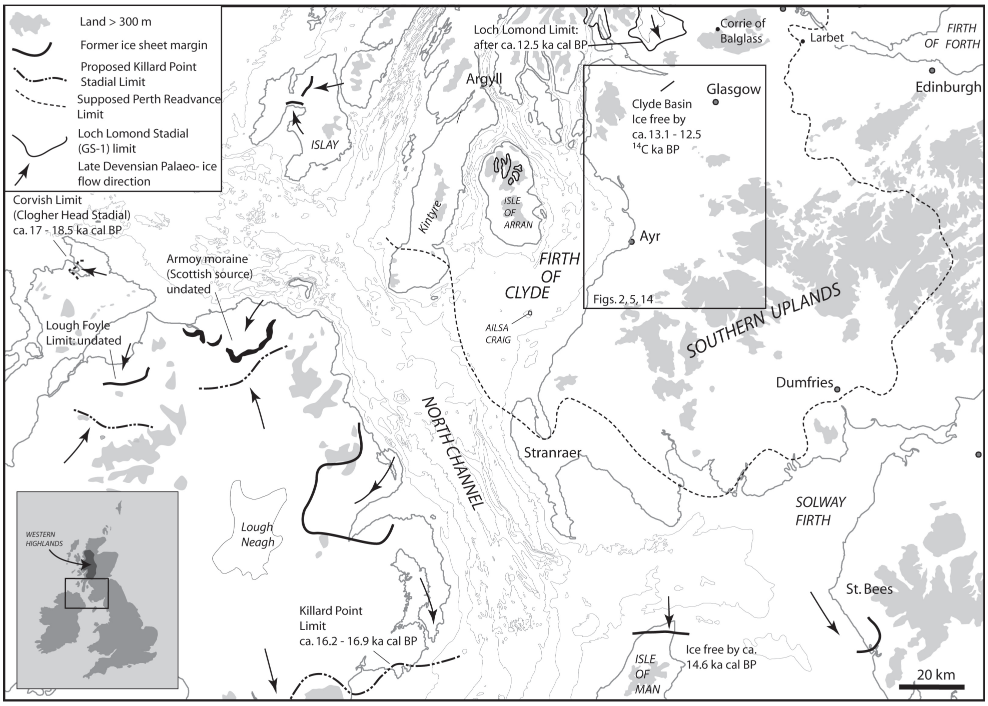


FIGURE 4.1: Regional context. Box (labelled Figs. 2, 5, 14) shows location of study area in west central Scotland. Proposed former glacier ice limits and ages from Sissons [1967b]; Dawson [1982]; Rose et al. [1988]; Peacock and Merritt [1997b]; McCabe et al. [1998]; Thomas et al. [2004]; Ballantyne [2007a], and McCabe [2008]. Key place names are shown.

provide a coherent, conceptual model of ice sheet evolution for this zone of the last BIIS.

4.2 Background

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

The western central lowlands of Scotland have been recognised as an area affected by complex ice-flow patterns since Geikie [1863] described it as the 'debatable ground over which glaciers of the Highlands or Southern Uplands (Fig. 4.1) prevailed according to their contemporary strengths'. Highland-sourced ice initially extended south-eastwards to the northern flanks of the Southern Uplands as suggested by the distribution of indicator erratics and the widespread occurrence of 'lower' tills of north-western provenance underlying 'upper' tills derived from the south [Price, 1975; Sutherland and Gordon, 1993]. Southern Uplands ice became more dominant later, deflecting Highland ice both to the east and south-west. The easterly diverted flow left a strong imprint on the landscape of the Clyde basin, in the form of extensive, drumlin assemblages [Rose, 1987; Hall et al., 1998; Rose and Smith, 2008].

Former changes in ice flow direction are particularly apparent in central Ayrshire where evidence from erratics, stratigraphy, cross-cutting striae and roches moutonées demonstrates that ice firstly flowed onshore towards the south-east and then offshore south-westwards [Richey et al., 1930]. Highland-sourced ice initially penetrated at least as far south as Nith Bridge (Fig. 4.2), carrying shells that were probably scavenged from the Firth of Clyde [Holden and Jardine, 1980; Sutherland, 1993]. This early, on-shore flow of ice resulted in widespread deposition of shelly tills and rafts of glaciomarine clay, notably at Afton Lodge and Greenock Mains (Fig. 4.2) [Smith, 1898; Holden, 1977; Abd-Alla, 1988; Gordon, 1993a,c]. Evidence for a later, onshore readvance of Highland-sourced ice at the latter locality [Holden, 1977] implies that active ice occupied the Firth of Clyde and Ayrshire Lowlands after the eastern central lowlands had deglaciated [Sutherland, 1984].

Most evidence for these switches in flow relates to the last, MLD glaciation when the BIIS is thought to have reached the continental shelf edge to the north-west of the British Isles and merged with ice from Fennoscandia in the North Sea Basin [Graham et al., 2007; Bradwell et al., 2008b]. However, there is an unusually high concentration of mammalian fossil occurrences and other organic remains within glacigenic sequences that have survived the MLD glaciation, particularly in the Ayrshire Lowlands and the lower Clyde valley [Bishop and Coope, 1977; Sutherland and Gordon, 1993]. Early age determinations from bones of woolly rhinoceros from Bishopbriggs [Rolfe, 1966] and reindeer antler fragments from Sourlie (Fig. 4.2) [Jardine et al., 1988] yielded ages of c. 27-30 ka ^{14}C BP. A revised age of 31.1 ka ^{14}C BP (c. 35 ka cal BP) has recently been published for the Bishopbriggs sample, following ultrafiltration pre-treatment [Jacobi et al., 2009]. This age is similar to those obtained from organic remains within fully investigated interstadial profiles beneath till at Balglass (Fig. 4.1) [Brown et al., 2007]) and Sourlie [Bos et al., 2004]. Collectively, these dates suggest that the MLD ice sheet did not become established in the area until after c. 35 ka cal BP, contrary to the conclusions of Bowen et al. [2002]. However, numerical modelling experiments simulate minor glacial advances into the area prior to the main sustained advance in the Late Devensian [Hubbard et al., 2009]. Indeed, the moderately-weathered Lawthorn Diamicton that underlies the interstadial deposits at Sourlie, and the Ballieston Till

Formation (see below), which occurs within concealed depressions beneath Glasgow, must relate to an earlier expansion of glacier ice.

There is general agreement that the MLD ice sheet withdrew towards the west and northwest during deglaciation of the area [Price, 1983; Sutherland, 1984]. Ice-marginal lakes formed where ice impeded drainage, firstly on the watershed between the catchments of the Avon Water and River Irvine, in Glengavel (Fig. 4.2), where laminated glaciolacustrine silts occur to an elevation of at least 205 m above sea level (a.s.l.) [Nickless et al., 1978]. This lake came into existence shortly after Highland and Southern Uplands-sourced ice had separated [Phemister in Richey, 1926; Mclellan, 1969; Martin, 1981]. Water held within the upper Avon Valley merged with a much larger lake, 'Lake Clydesdale' [Bell, 1874], which eventually occupied the Clyde valley and its tributaries upstream of Glasgow. The level of Lake Clydesdale probably dropped in stages as north-westward retreat of the ice margin in the lower Clyde Valley made available spillways to the east at progressively lower elevations of 200, 165, 102, and 85 m a.s.l. respectively [Paterson et al., 1998]. The lake finally drained eastwards via a col in the upper Kelvin valley at about 45 m a.s.l. [Forsyth et al., 1996; Hall et al., 1998].

Lake Clydesdale probably existed during the creation of the Main Perth Shoreline in the Forth estuary [Sissons and Smith, 1965; Sutherland, 1984], and possibly into the beginning of the Lateglacial Interstadial (GI-1) [Peacock, 1999, 2003]. The timing, contemporary sea level and manner in which the late-glacial sea eventually invaded Lake Clydesdale is disputed [see Peacock, 2003, for review], but it is generally accepted that the transgression had occurred by 13.1 - 12.514C ka BP (based on uncorrected, reported ages from marine shells) [Peacock, 1971, 2003; Peacock et al., 1977; Browne et al., 1977; Rose, 2003]. The presence of a radiocarbon plateau at this time, and uncertainty regarding the reservoir correction at the Greenland Stadial-2 (GS-2) to GI-1 boundary, preclude a more precise chronology based on ^{14}C dates alone [Peacock, 2003]. Final disintegration of ice blocking the Clyde estuary resulted in the level of Lake Clydesdale falling from 45 m to no less than 40 m, the contemporary sea level. Sea level then fell rapidly, but further discussion of the subsequent, complex sea-level history of the area is not presented here, nor discussion of events during the Loch Lomond Stadial (Greenland Stadial-1, GS-1), when ice readvanced from the Loch Lomond basin to the northwest of Glasgow [Rose et al., 1988; Evans, 2003; Rose and Smith, 2008].

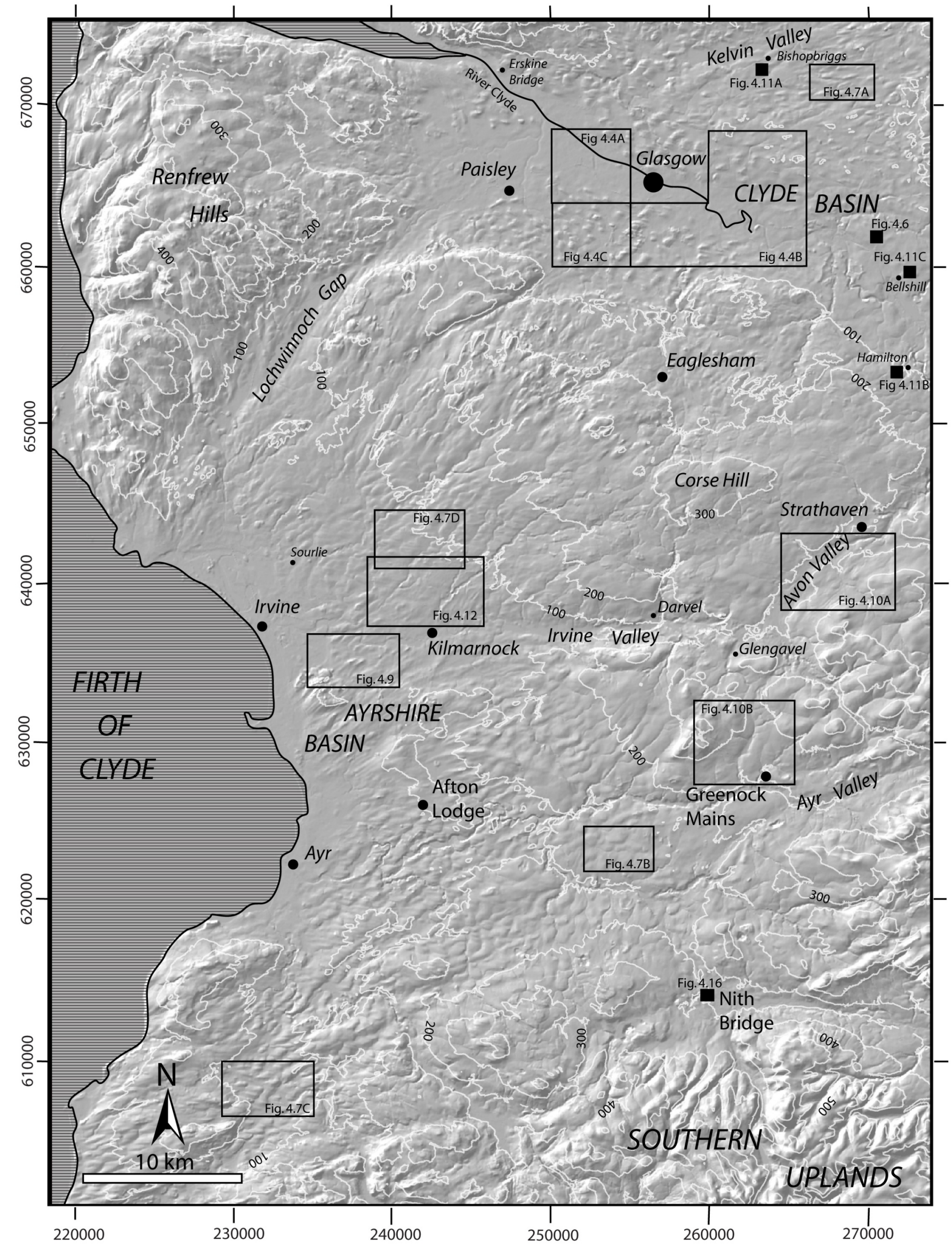


FIGURE 4.2: Topography of study area and place names mentioned in text. Locations of subsequent figures are shown. Hill-shaded digital surface model built from Intermap Technologies NEXTMap Britain topographic data. Northwest illumination.

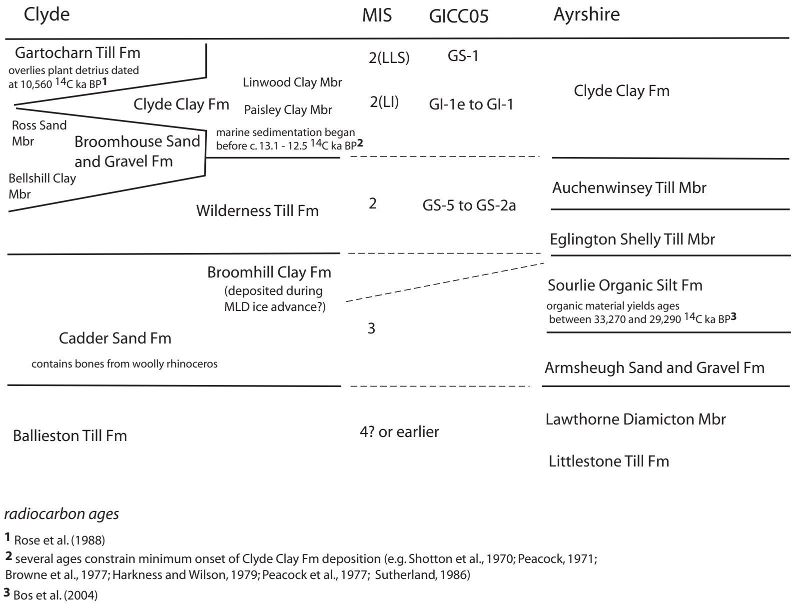


Figure 4.3: Simplified lithostratigraphy for the period spanning the Main Late Devensian glaciation for the Clyde and Ayrshire basins. Based on McMillan et al. [2010]. For clarity only formations of primary relevance to this study are included. Fm - Formation, Mbr - Member. GICC05 - Greenland Ice Core Chronology 2005 events, after Lowe et al. [2008].

4.3 Lithostratigraphy of west central Scotland

Despite the success of modern geomorphological analysis in distinguishing the relative ages of landforms, it is essential also to consider the known sequence of deposits (lithostratigraphy) in order to determine a robust event stratigraphy. Detailed lithostratigraphical knowledge also underpins three-dimensional modelling of Quaternary deposits [e.g. Merritt et al., 2007]. The lithostratigraphy for the Clyde and Ayrshire basins presented here (Fig. 4.3) develops that of Sutherland [1999] and follows a new top-down, nationwide framework [McMillan et al., 2010, , www.bgs.ac.uk/lexicon].

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

In west central Scotland, raised glaciomarine deposits of late-glacial age are assigned to the **Clyde Clay Formation** [McMillan et al., 2010] in which two principal members are recognised, the **Paisley Clay** and **Linwood Clay**. The former member generally comprises thinly laminated clays and silts with dropstones. This member is generally poor in fauna, only yielding the cold-water foraminifera *Elphidium clavatum* in significant numbers. It has been mapped in areas around the Clyde estuary up to altitudes of ~40 m a.s.l. [Browne and McMillan, 1989b]. The Linwood Clay Member is confined to western areas of the Clyde estuary where it commonly overlies the Paisley Clay. It consists of more thickly bedded silts and clays with a richer faunal assemblage [Browne and McMillan, 1989b; Peacock, 2003].

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

The Quaternary lithostratigraphy of central Ayrshire follows McMillan et al. [2010] and is based mainly on the succession that was exposed in an opencast coal site at

Sourlie, near Irvine (Fig. 4.2) [Jardine et al., 1988; Sutherland, 1999]. The site was excavated into the north-western side of Sourlie Hill, one of a swarm of broadly eastward orientated drumlins. The importance of the site lay in the discovery of thin lenses of organic material (**Sourlie Organic Silt Formation**) occurring between two units of till. These lenses yielded a very rich flora and fauna deposited within a shallow pond in a treeless, low-shrub to sedge-moss tundra environment, and included bones of woolly rhinoceros and reindeer. Radiocarbon dates on antler fragments, plant debris and bulk organic matter suggest a Middle Devensian age [Bos et al., 2004].

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

4.4 Methods

4.4.1 Remote sensing evidence

Remote sensing datasets were interrogated within ESRI Arc Map 9.2. Digital surface models (DSMs) and georectified 1:10,000 monoscopic aerial photographs were analysed to identify glacial landforms in the study area. The surface models, built from

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

4.4.2 Three-dimensional geological evidence

Over 60,000 borehole records exist for Glasgow and the surrounding area. The British Geological Survey is currently creating a suite of three-dimensional Quaternary and bedrock models, based on the borehole data [Merritt et al., 2005, 2007] using the modelling software tool, GSI3D [Sobisch, 2000; Kessler et al., 2006]. In this study, outputs from these three-dimensional geological models were used for two purposes: (i) to confirm the basic composition of landforms, thereby enabling more confident discrimination between true glaciogenic features and bedrock controlled features; and (ii) to aid interpretation and identification in areas where subglacial landforms are masked by younger deposits (Fig. 4.4), or modified at the surface due to urban development. A comprehensive UK database containing borehole records (BGS Borehole Geology) was interrogated throughout the investigation to provide additional information about the surface and subsurface sediments in the mapping area.

4.4.3 Field evidence

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

4.4.4 Data compilation

All landforms and mapped sediment distributions were captured in a spatially attributed ArcGIS database. Landforms mapped include: ribbed moraine, streamlined bedforms, meltwater channels, moraine ridge complexes and narrow transverse ridges. Existing Quaternary geological maps of the area [B.G.S., 1993a, 1994, 1993b, 1987a, 2002; Browne and McMillan, 1989a] were consulted throughout the study. A recently compiled 2-D digital geological map of Britain at 1:50:000 scale (DiGMapGB 50) was interrogated in the GIS and forms the basis for mapped distributions of glaciofluvial, deltaic, and raised marine sediments, and areas where bedrock occurs at or near the surface.

4.5 Results

A glacial geological and geomorphological map of the greater Clyde basin is shown in Figure 4.5. Some detail is lost reproducing the map at this scale, and numerous smaller features such as minor meltwater channels and individual moraine crests, are not shown. The morphological, spatial and, where known, basic sedimentological characteristics of landform assemblages are described below.

4.5.1 Ribbed moraine

Suites of southwest to northeast aligned, broad transverse ridges occupy the Clyde and Ayrshire basins up to an elevation of $\sim$ 200 m a.s.l. These ridges are 0.4-1.2 km in width, 0.4-6.5 km in length and up to 40 m in height. On a morphological basis, the ridges can be described as ribbed moraine, and their dimensions are entirely consistent with those reported by Dunlop and Clark [2006]. Occupying areas principally underlain by Carboniferous sedimentary rocks and basalts, the ribbed moraine maintain a long-axis alignment that does not concord with variations in bedrock strike. Three-dimensional geological models (Fig. 4.4) in the Clyde basin indicate that these ridges commonly consist of glacial sediments assigned to the Wilderness Till Fm; thus their form is not considered to be controlled by bedrock structure.

The ribbed moraine are extensively remoulded with development of, or modification into, elongate streamlined bedforms (described below). A 50-m-long temporary section within a broader zone of drumlinised ribbed moraine ridges was described by McMillan and Browne [1983] and Browne and McMillan [1989b]. Here, a 2- to 5-m thick surface

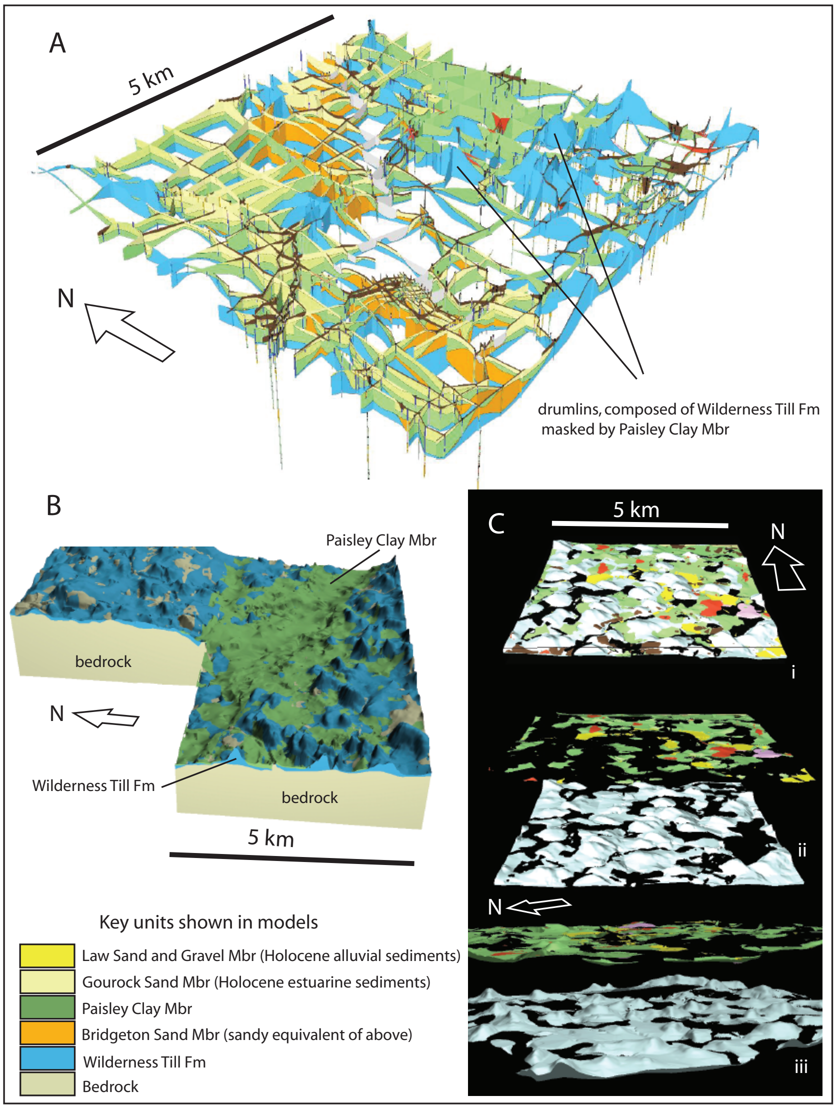


FIGURE 4.4: Three-dimensional Quaternary geological models revealing basic composition of geomorphological features in the Glasgow area. A. Fence diagram revealing three-dimensional geology of Erskine-Renfrew area. Note drumlins entirely comprise sediments of the Wilderness Till Formation (in blue). Cross sections are based on borehole records. Vertical sticks represent individual boreholes B. Complete three-dimensional geological model for the central Glasgow area showing Paisley Clay Mbr draped over drumlins comprising Wilderness Till Fm. C. Three dimensional geological model of Paisley area. Paisley Clay Formation sediments (in green) and alluvial sediments (in yellow) are removed to more clearly reveal bedforms in the Wilderness Till Formation (blue). All images are vertically exaggerated between 5 and 10 times.

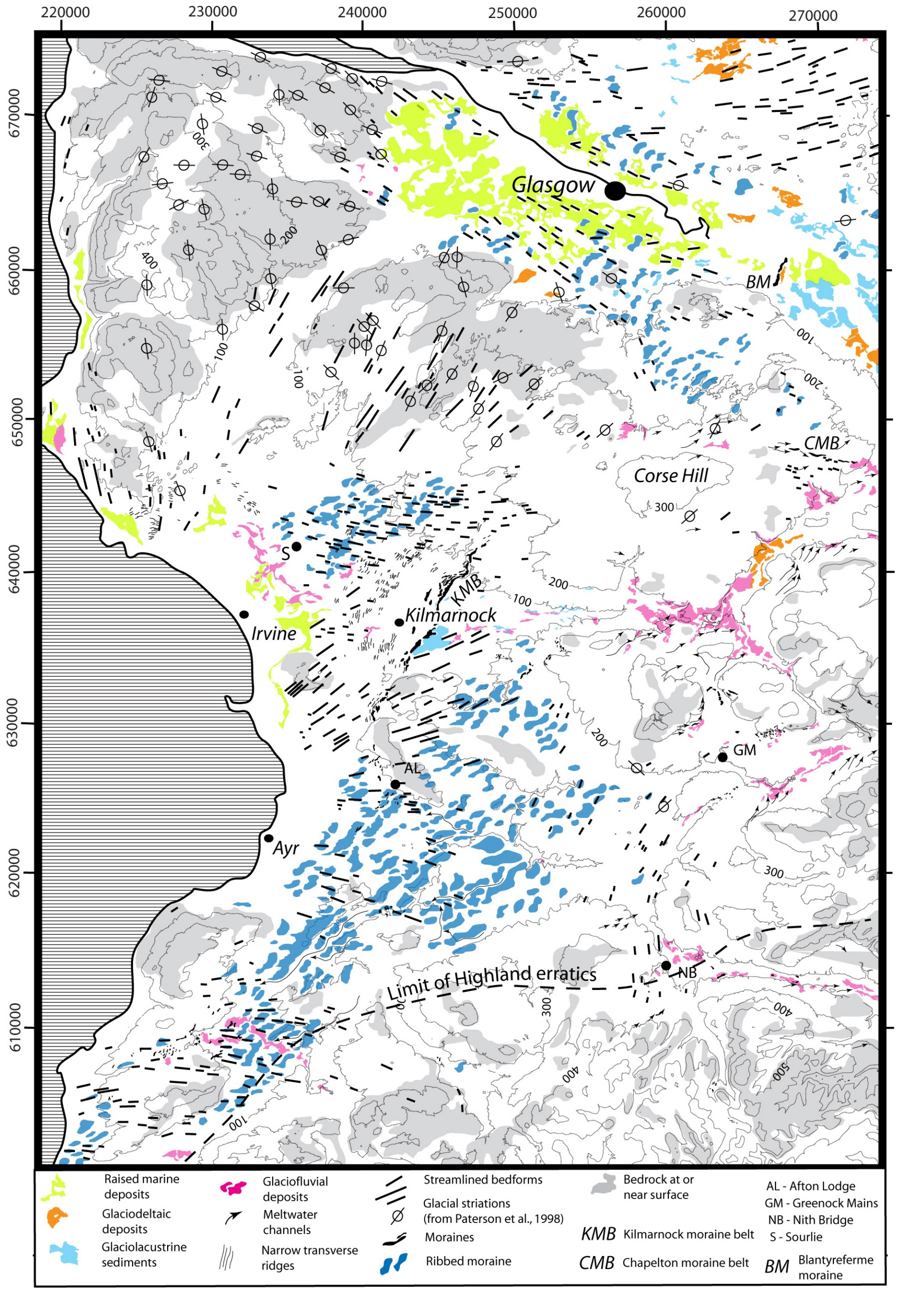


Figure 4.5: Glacial geomorphology and geology of the Clyde and Ayrshire basins. Erratic limits from Eyles et al. [1949] and glacial striations from Paterson et al. [1998].

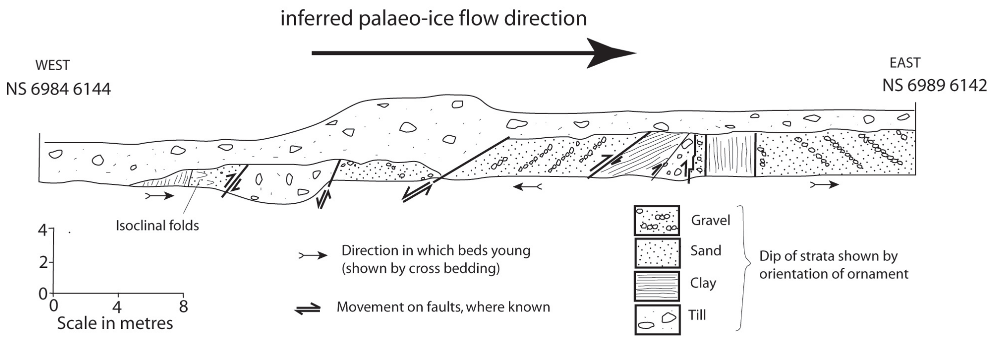


FIGURE 4.6: Temporary section within zone of drumlinised ribbed moraine at Holmbrae Road, Glasgow. Ice flow direction inferred from regional streamlining and sense of compression. From McMillan and Browne [1983].

carapace of red-brown sandy clayey till of the Wilderness Till Fm sharply truncates a series of underlying sediments (Fig. 4.6). The lower sediments comprise till, bedded gravels, sands and clay, and form a series of thrust slices dipping steeply westwards. Two normal faults occur in these lower sediments, on the western side of the thrust stack.

4.5.2 Streamlined bedforms

Streamlined bedforms are well developed within the Clyde and Ayrshire basins, around the margins of the Southern Uplands, and to the southeast of the Lochwinnoch Gap (Figs. 4.2, 4.5). In the latter area they occur where till is thin and patchy, and locally are strike parallel to the gently dipping Clyde Plateau Volcanic Formation. In that locality, bedforms are probably influenced by bedrock structure. Cover of Quaternary sediments is much thicker over the basins to the north and south, which are underlain mainly by Carboniferous sedimentary rocks. That bedforms in the Clyde Basin are of glacial origin is supported by three-dimensional geological models, which show that the landforms principally consist of glacial sediments assigned to the Wilderness Till Fm (Fig. 4.4). Many of the streamlined bedforms in the basins are superimposed on, or consist of, re-shaped sections of the ribbed moraine (Fig. 4.7).

Numerous geomorphologically-based ice sheet reconstructions use the approach of grouping bedforms into coherent 'flowsets' or 'swarms' [e.g. Boulton and Clark, 1990; De Angelis and Kleman, 2007; Stokes et al., 2009; Greenwood and Clark, 2009]. The streamlined bedforms identified here can be broadly divided into six flowsets based on their geographical distribution, trend, morphology and spatial relationships with other geomorphological features (Fig 4.8, Table 4.1).

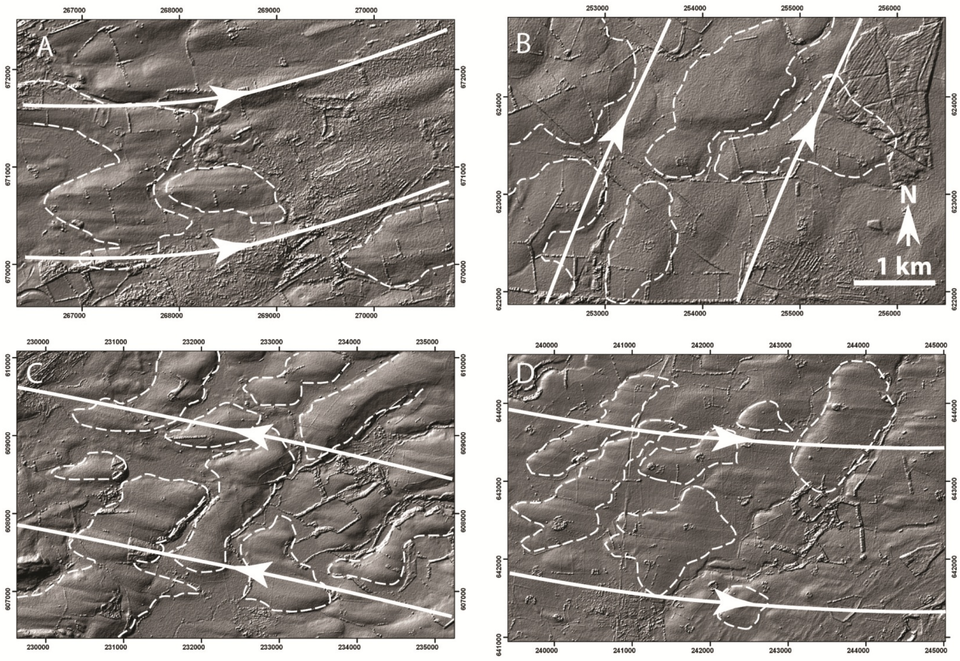


FIGURE 4.7: Streamlined bedforms superimposed on ribbed moraine. See text for description. Hill-shaded digital surface models built from Intermap Technologies NEXTMap Britain topographic data. Northwest illumination.

4.5.2.1 Flowset-I

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

4.5.2.2 Flowset-II

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

4.5.2.3 Flowset-III

Flowset III comprises ice-moulded bedrock and crag-and-tail forms over higher elevations and drumlins in the lower basin areas. They are well-preserved, and trend in a south to west-northwest direction, forming an overall convergent pattern towards the southwest (Fig. 4.8). Subsets III-a to III-e are identifiable within flowset-III on the basis of slight variations in alignment and differences in morphological characteristics. For example, subsets III-b and III-c have a considerably longer mean length (> 1 km) (Table 1). This probably reflects thinner till cover, and local concordance with strike of the gently dipping volcanic rocks. Although some of these bedforms may be influenced by bedrock structure over higher elevations, a consistent convergent trend is maintained in the lower-lying sediment-filled basins. The transition between individual subsets is largely gradational; thus, they are all incorporated within flowset-III.

4.5.2.4 Flowset-IV

Flowset-IV comprises a small cluster of streamlined bedforms ~ 12 km south from the Blantyreferme Moraine (see below). These bedforms trend in an east to east-southeast direction. They possess the shortest lengths (mean 321 m) of all the flowsets identified (Table 4.1).

4.5.2.5 Flowset-V

Flowset-V comprises a well-defined suite of streamlined bedforms trending towards the east. They are generally confined to the western side of the Kilmarnock Moraine Belt (Fig 8) (see below); only a few isolated bedforms occur on the immediate eastern side. Streamlined bedforms belonging to flowset-V overprint those of flowset-III (Figs 4.8, 4.9).

4.5.2.6 Flowset-VI

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

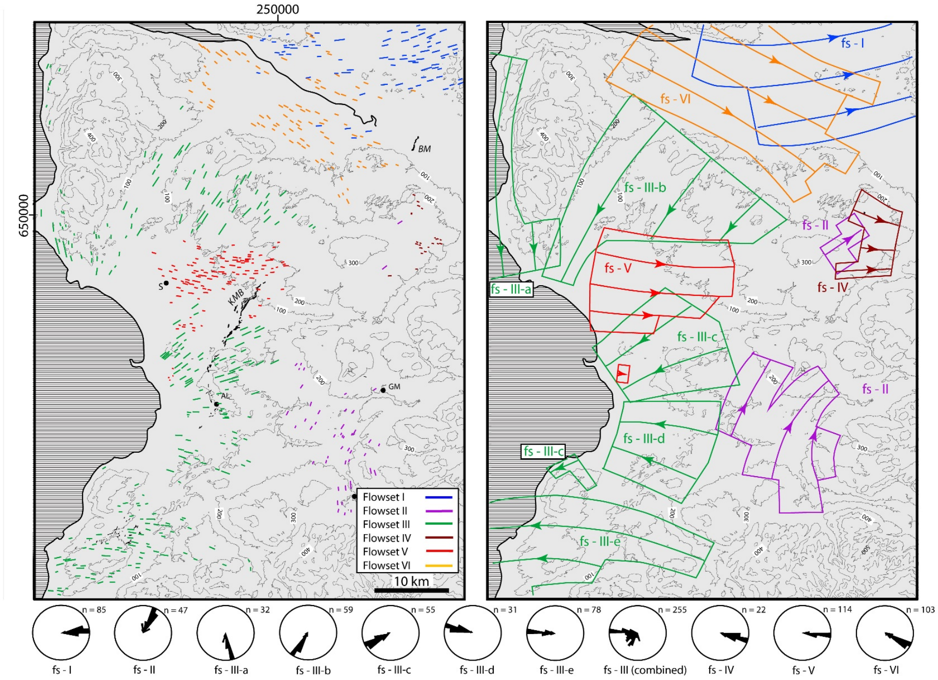


Figure 4.8: Streamlined bedforms and flowsets identified in this study. Fs - flowset. Bedform orientations are shown in the rose plots in the lower panel.

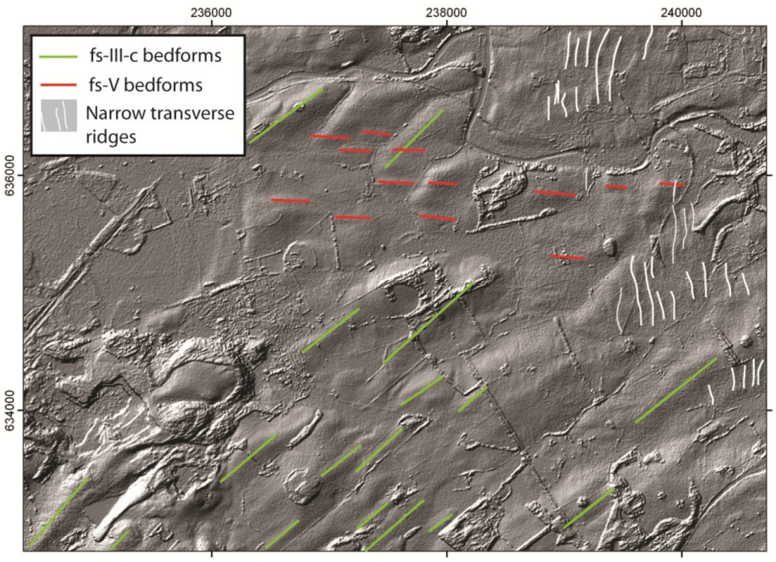


Figure 4.9: Cross cutting of streamlined bedforms in the Ayshire basin to the west of Kilmarnock.

Flowset	Length			Width			ER		
	Range	Mean	St Dev	Range	Mean	St Dev	Range	Mean	St Dev
Fs-I	370-1498	826	262	75-468	239	87	2.1-7.6	3.7	1.1
Fs-II	330-1195	563	187	65-443	307	83	1.6-7.7	2.9	1.0
Fs-III-a	350-2430	706	476	76-500	194	103	2.0-5.3	3.7	1.0
Fs-III-b	379-2391	1017	445	79-590	274	108	2.3-6.9	3.8	0.9
Fs-III-c	400-2600	1063	367	100-467	236	78	2.5-8.0	4.6	1.3
Fs-III-d	449-1214	766	226	118-469	250	88	1.7-5.7	3.3	0.9
Fs-III-e	372-1700	754	244	72-520	197	84	2.2-11.6	4.3	1.5
Fs-III (all)	350-2600	877	384	72-590	230	96	1.7-11.6	4.0	1.3
Fs-IV	204-487	321	86	51-134	76	22	3.1-5.8	4.3	0.9
Fs-V	167-1142	540	179	61-290	127	39	2.3-8.0	4.3	1.1
Fs-VI	290-1206	660	155	80-467	232	79	1.7-6.4	3.1 7 0.8	

TABLE 4.1: Morphological characteristics of streamlined bedforms in study area. ER - elongation ratio, St Dev - standard deviation.

4.5.3 Glaciofluvial assemblages

Glaciofluvial assemblages described here include both moundy, kettled, ice-contact deposits and terraced outwash spreads, both assigned to the Broomhouse Sand and Gravel Formation. It is worthy of note that glaciofluvial deposits portrayed on BGS maps may include deltaic sequences that formed in ice-marginal or proglacial lakes; only widespread, fine-grained deposits are generally identified as 'glaciolacustrine'. Major belts of glaciofluvial deposits occur principally in the valleys of the Clyde, Kelvin and Avon [B.G.S., 1993b,a, 1994]. Many of the deposits in the southeast of the area are associated with systems of north-easterly-descending ice-marginal meltwater channels, notably southeast of Eaglesham and south of Strathaven [Richey et al., 1930; Paterson et al., 1998], where they fall from about 320 to 260 m a.s.l. (Fig. 4.10A). Further significant belts of mounds and undulating spreads of sand and gravel occur in the upper Ayr and Nith valleys [B.G.S., 1982]. These deposits are also associated with coherent systems of ice-marginal meltwater channels, which in the upper Ayr valley descend from about 300 to 250 m a.s.l. towards the east. Further sets of eastward descending marginal meltwater channels exist in the Nith valley at altitudes from about 300 m down to 200 m a.s.l.

4.5.4 Moraine ridge complexes

Three major moraine complexes have been identified in the Ayrshire and Clyde basins, at Kilmarnock, Blantyreferme and Eaglesham respectively (Figs. 4.2, 4.5). The Kilmarnock Moraine Belt (KMB, Fig. 4.5) extends for approximately 14 km in a south-southwest to north-northeast direction, and ranges from 5 to 20 m in height. To the northeast of Kilmarnock, the belt reaches a maximum width of nearly 800 m, where it forms multiple crests. Borehole records indicate that at least part of the complex

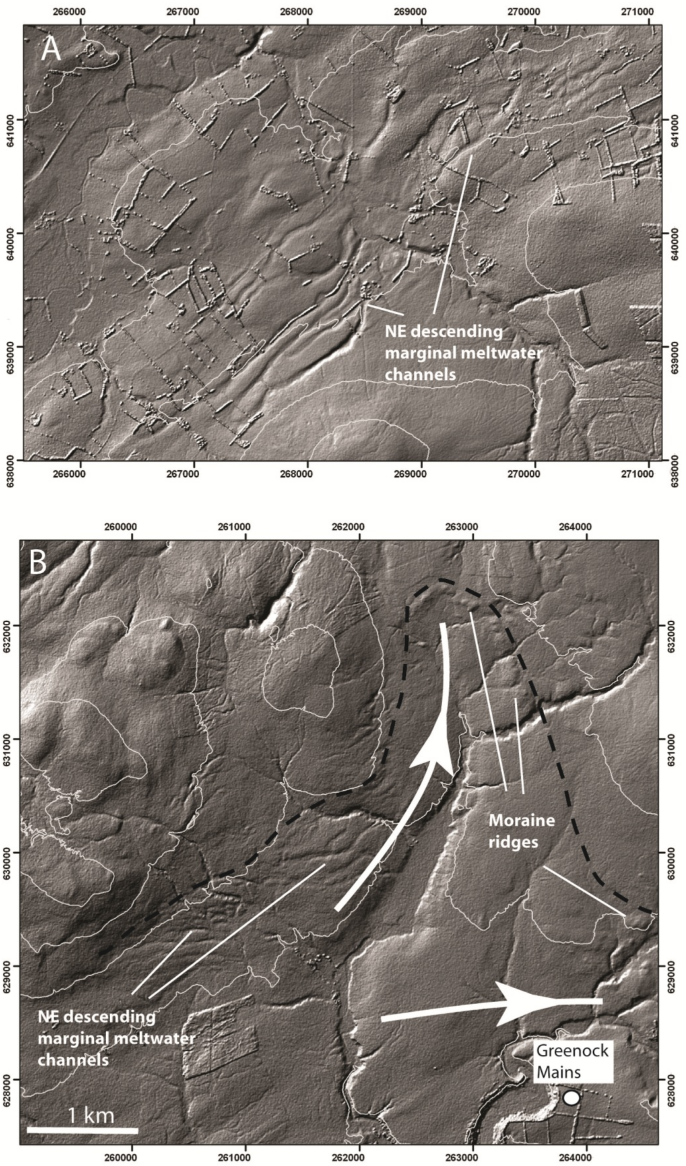


FIGURE 4.10: A. Ice-marginal meltwater channels clearly descending towards the NE, to the south of Strathaven. B. Assemblage of moraine ridges and NE declining marginal meltwater channels in the vicinity of Greenock Mains. White arrows denote inferred final ice flow direction. Hill-shaded digital surface models built from Intermap Technologies NEXTMap Britain topographic data. Northwest illumination.

comprises clay and sandy clay, while records immediately northwest of the ridge reveal till interbedded with sandy clay. Flat terrain immediately southeast of the complex is underlain by up to 7 m of sands, laminated silts and clays.

The Blantyreferme Moraine (BM, Fig. 4.5) was first recognised by Clough et al. [1911]. Forming a near-symmetrical, cross-valley ridge, it is aligned south-southwest to north-northeast extending for over 2 km, and reaching nearly 20 m in height. Field mapping has revealed the feature to be of variable lithology, comprising till, sand and gravel and also laminated clay and silt [Browne and McMillan, 1989b].

The Chapelton Moraine Belt (CMB, Fig. 4.5) comprises a string of ridges and mounds that lie to the east of Eaglesham on the northern slopes of Corse Hill [Richey et al., 1930; Paterson et al., 1998]. These landforms, which include esker fragments, were formed at the southern margin of Highland-sourced ice early in the deglaciation of the area. They descend eastwards from about 305 to 260m a.s.l. and were described by Sissons [1963, 1964, 1967a] as part of more widespread evidence for the supposed 'Perth Readvance' [Simpson, 1933].

A further, near-coherent chain of moraine ridges occurs in the upper Ayr valley, above Greenock Mains. Individual ridges, up to 250 m in length, extend from $\sim$ 4 km north of Greenock Mains eastward for $\sim$ 8 km, declining in altitude from 310 to 270 m a.s.l. Suites of eastward declining, marginal meltwater channels occur on the northern and southern flanks of the upper Ayr valley, at altitudes of about 305 to 270 m a.s.l. Those on the northern side merge with the moraine ridges to the north of Greenock Mains (Fig. 4.10B).

4.5.5 Glaciolacustrine assemblages

Extensive spreads of fine-grained glaciolacustrine sediment of the Bellshill Clay Member occur at surface southeast and east of the Blantyreferme Moraine (Fig. 4.5). Glaciolacustrine sediments also crop out locally on the western flanks of the Kelvin valley [Browne and McMillan, 1989b; Hall et al., 1998] and along the margins of the Irvine valley downstream of Darvel [Nickless et al., 1978].

The glaciolacustrine assemblages in the Clyde basin commonly pass up and laterally into flat-topped, deltaic deposits of the Ross Sand Member [Browne and McMillan, 1989b; Martin, 1981]. Formerly exposed sections revealed sands and gravelly sands forming dipping foresets of Gilbert-type deltas (Fig. 4.11A, B). The deltaic deposits locally exceed 20 m in thickness in the eastern Clyde basin [Browne and McMillan, 1989b].

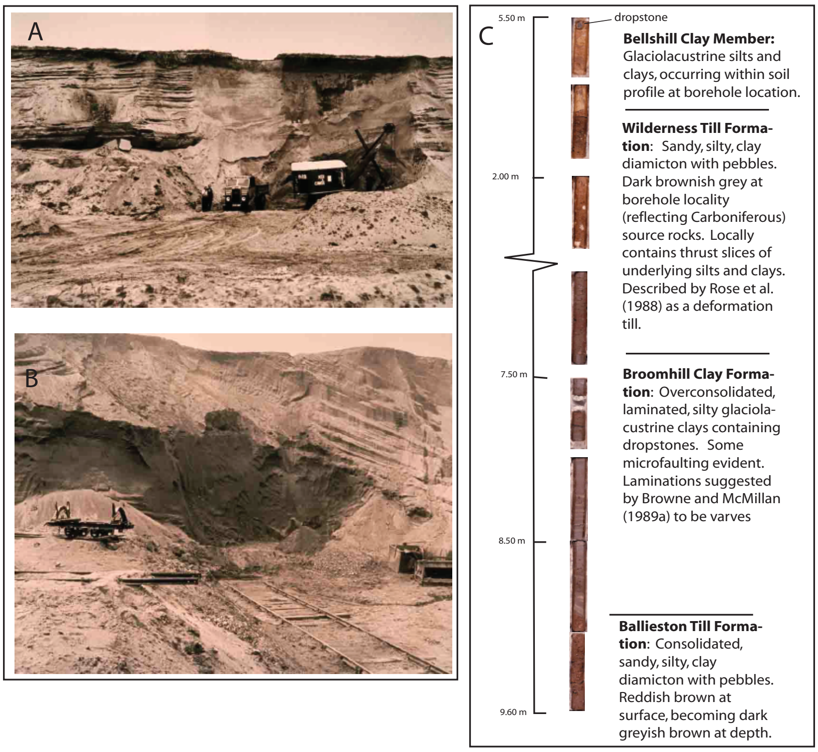


FIGURE 4.11: Deposits associated with ice-dammed lakes in the Clyde basin. Deltaic sediments of the Ross Sand Mbr revealed in former sand and gravel pits near Bishopbriggs (A) and Hamilton (B). Photos from BGS archive image base. C. Lithostratigraphy including surface and buried glaciolacustrine deposits, revealed in the BGS Bellshill borehole.

4.5.6 Narrow transverse ridges

Two suites of previously unreported, closely spaced, narrow, linear transverse ridges occupy parts of the lower Irvine valley (Fig. 4.12) and the southern entrance to the Lochwinnoch Gap. The former, situated on the western side of the KMB trend broadly from south-southwest to north-northeast, and the latter trend broadly west to east. Those in the lower Irvine valley occur between about 30 m and 150 m a.s.l., while those south of the Lochwinnoch Gap lie between about 35 m and 140 m a.s.l. The former have a mean width of 75 m, a mean height of 2.4 m, and generally possess a symmetrical cross-profile (Fig. 4.13). Many of the ridges are continuous for over 400 m, maintaining their alignment across topographic undulations of up to 20 m in height. No sections have yet been observed within any of the landforms. However, borehole evidence from the Kilmarnock area demonstrates that surface sediments in the area of these narrow ridges comprise silts, sands and till. In places, the narrow transverse ridges are clearly superimposed on streamlined bedforms assigned to flowset V (Fig 4.12).

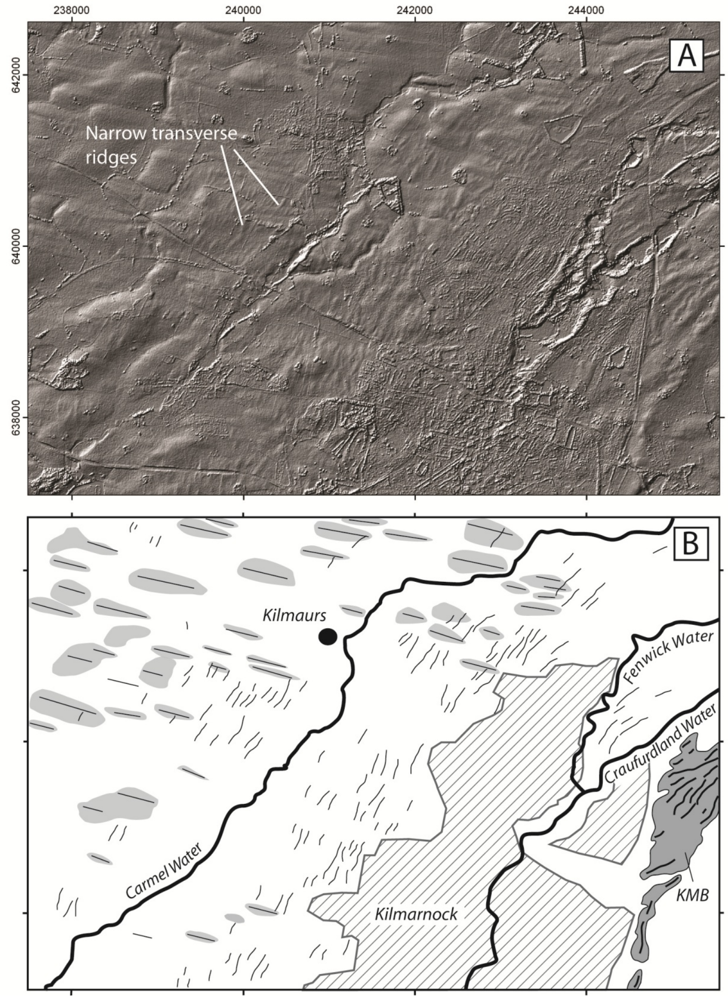


FIGURE 4.12: . A. Hill-shaded digital surface model, built from Intermap Technologies NEXTMap Britain topographic data, revealing narrow transverse ridges in the vicinity of Kilmarnock. Note overprinting of ridges on streamlined bedforms. B. Interpretation of same area.

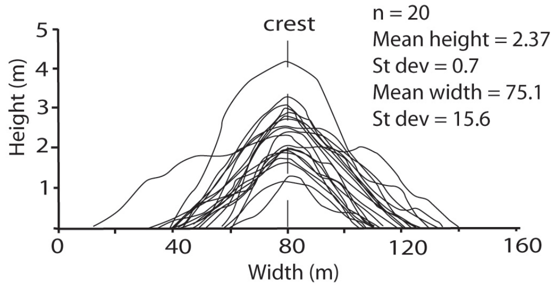


FIGURE 4.13: Cross-profile data extracted from the digital surface model revealing dimensions of the narrow transverse ridges.

4.5.7 Raised marine deposits

The lithostratigraphy of raised marine deposits in the Clyde basin has been described above. Field mapping has identified deposits of the Paisley Clay Member occupying extensive areas of the Clyde basin up to $\sim$ 40 m a.s.l. [Browne and McMillan, 1989b]. Three-dimensional geological modelling supports the field interpretation and reveals thick spreads of silts and clay, often partially masking the underlying, drumlinised landscape (Fig. 4.4). Further discussion of the distribution of raised marine deposits and associated features is not presented here.

4.6 Interpretation of events in west central Scotland

4.6.1 Pre-Late Devensian Glaciation

Evidence for a pre-Late Devensian, MIS 4 or older glacial advance-retreat cycle has been briefly discussed earlier in this chapter. The earlier glacial event led to the deposition of the Ballieston Till Formation in the Clyde basin, and the Littlestone Till Formation in Ayrshire; it probably involved a substantial advance of ice from the northwest. The apparently weathered top of the Ballieston Till suggests that there was a significant period of exposure before deposition of the overlying, glacitectonised, thinly laminated, glaciolacustrine sediments of the Broomhill Clay Formation in the Bellshill area (Figure 4.11C).

4.6.2 Late Devensian Glaciation; Stage A. (Fig.4.14A; build-up to LGM)

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

Instances exist where streamlined bedforms from each flowset are superimposed on the ribbed moraine ridges (e.g. Fig 4.7). Therefore, ridge formation must have occurred prior to the earliest phases of preserved streamlining in the study area. Regional

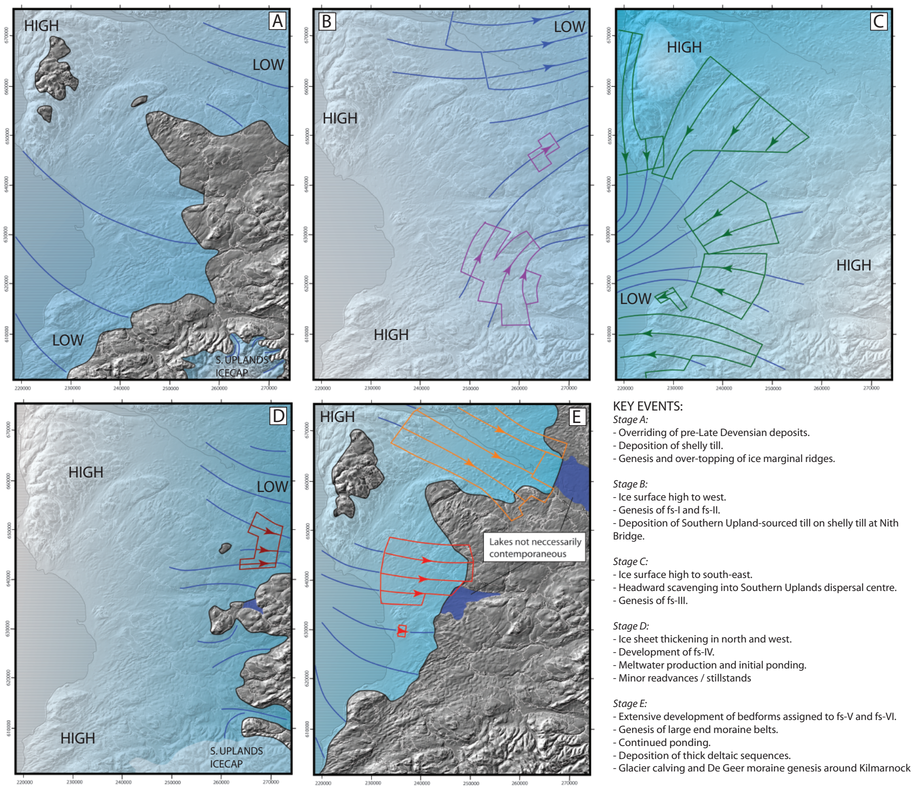


FIGURE 4.14: Reconstructed stages, showing the evolution of the last BIIS in west central Scotland. See text for discussion. Hill-shaded digital surface models built from Intermap Technologies NEXTMap Britain topographic data. Northwest illumination.

geological evidence [e.g. Price, 1975; Sutherland, 1984; Sutherland and Gordon, 1993] (Fig. 4.15) along with numerical ice sheet models [Hubbard et al., 2009] indicate that initial MLD ice-sheet advance into the area was from the northwest, broadly perpendicular to the ribbed moraine crest lines. We suggest this was the period of ribbed moraine formation (Fig. 4.14A), when the ice front advanced against a reverse slope, building (then overtopping) sediment ridges through folding and thrusting of proglacial sediments. A similar mechanism is invoked for the formation of 'cupola hills' elsewhere [Benn and Clapperton, 2000; Benn and Evans, 1998]. The sediments formerly exposed at Holmbrae Road in Glasgow (Fig. 4.6) [McMillan and Browne, 1983] are consistent with this interpretation. Initial advance led to thrusting of the gravel, sand and clay beds in the eastern side of the section. The two normal faults may have been activated during a minor ice margin retreat, prior to overriding and deposition of the upper (Wilderness) till. The concept that some ribbed moraine ridges originate as overridden ice marginal moraines has been proposed by Möller [2006]. However, rigorous investigation of the sediments would be required to test this hypothesis for the suites of ribbed moraines in the present study.

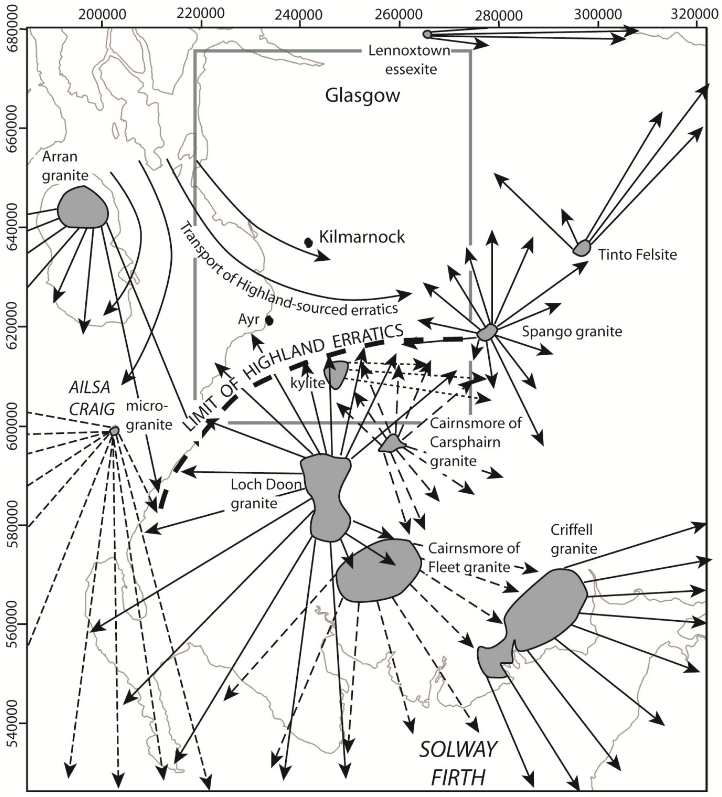
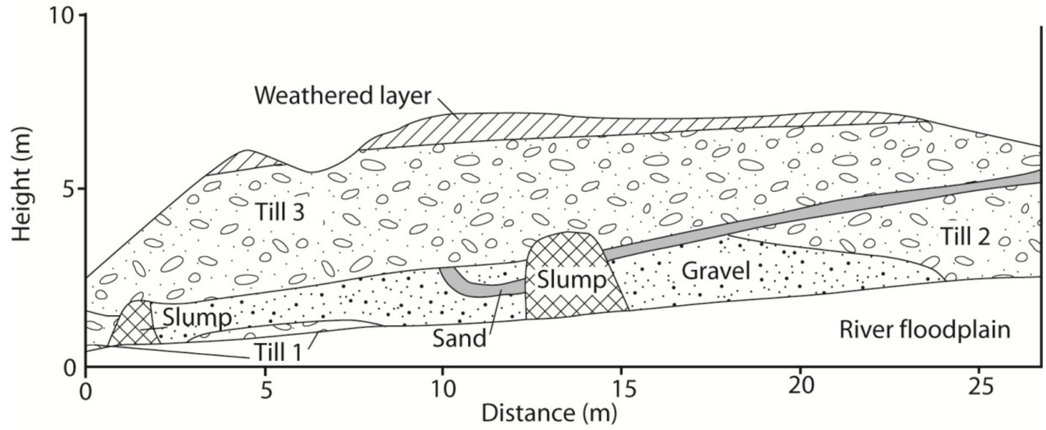


FIGURE 4.15: Erratic transport paths in SW Scotland. From [Eyles et al., 1949] and [Sissons, 1967a]. Note, erratic paths do not imply contemporaneous flow. Rectangle delimits main study area as shown in Figs. 4.2 and 4.5.

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

4.6.3 Late Devensian Glaciation; Stage B. (Fig.4.14B; LGM)

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



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

FIGURE 4.16: Section at Nith Bridge, from Holden and Jardine [1980].

dispersed from their source near Lennoxtown (Fig. 4.15) into the Firth of Forth [Peach, 1909; Shakesby, 1978; Evans et al., 2005]. Striae patterns [Paterson et al., 1998](Fig. 4.5) also document this flow.

Streamlined bedforms belonging to flowset-II were formed by north-eastward flow towards the Firth of Forth, driven by thicker ice to the southwest. This flow would have deposited Southern Upland till, with north-orientated clasts over the lower shelly tills (possibly of the Eglington Shelly Till Member) as described at Nith Bridge [Holden and Jardine, 1980; Sutherland, 1993] (Fig. 4.16).

4.6.4 Late Devensian Glaciation; Stage C. (Fig.4.14C)

Streamlined bedforms from flowset-III document convergent south-westward and westward flow into the Firth of Clyde, and are consistent with patterns of glacial striations in the north Ayrshire basin [Paterson et al., 1998](Fig. 4.5). This evidence suggests that a major change in the ice sheet configuration had occurred, largely caused by drawdown to the west. Subsets of flowset-III probably document a transgressive phase where part of the Southern Upland dispersal centre was scavenged by this increasing drawdown as ice flowed westward over the south of Arran. Westward transportation of Ailsa Craig erratics [Sissons, 1967a] (Fig. 4.15) would have occurred during this flow phase. By this stage, the north-eastward flow that generated flowset-II (Fig 4.14B - see above) must have switched off, allowing eastward migration of the ice divide, beneath which minimal subglacial modification was occurring.

4.6.5 Late Devensian Glaciation; Stage D. (Fig.4.14D)

A further, substantial alteration in ice sheet configuration and local basal conditions is indicated by the following suite of landforms: limited east-trending streamlining (flowset-IV), eastward descending marginal meltwater channels, eastward descending suites of ice contact glaciofluvial landforms (including the Chapelton Moraine Belt), and minor eastward-pushing morainic assemblages in the Ayr valley. Collectively, they demonstrate ice-divide migration back towards the west, coupled with ice-sheet thickening in the vicinity of the Firth of Clyde. The configuration is roughly that originally proposed for the 'Perth Readvance' in central Scotland [Sissons, 1963, 1964, 1967b]. Of note is the limited bedform development from this stage, with only a small patch of Group IV bedforms occurring in the east. A minor readvance is apparent in the Ayr valley near Greenock Mains where a coherent assemblage of meltwater channels and moraine ridges indicates a late, north-eastward push (Fig. 4.10B) [Holden and Jardine, 1980].

4.6.6 Late Devensian Glaciation; Stage E. (Fig.4.14E)

A subsequent phase of more persistent streamlining is indicated by bedforms assigned to flowset-V and VI. These bedforms are generally longer than those of flowset-IV to the east (Table 4.1). Abrupt termination of flowset-V and VI just beyond (to the east of) the Kilmarnock and Blantyreferme moraine complexes demonstrates that the ice flow phase that generated them extended to these areas. This flow phase probably included, or was followed by ice margin stabilisation at the moraines. Cross-cutting of drumlins in the Glasgow area [Rose and Letzer, 1977] indicates that flowset-I drumlins must have stopped developing prior to formation of flowset-VI bedforms.

To the east of the Kilmarnock and Blantyreferme moraine complexes (which may or may not be contemporaneous features), vast suites of glaciolacustrine and deltaic sediments in the Clyde and Irvine valleys demonstrate existence of ice-dammed lakes. The narrow, closely-spaced ridges observed in the Kilmarnock area (Figs. 4.5, 4.12) and to the west of the Lochwinnoch Gap (Fig. 4.5) are similar in morphology to De Geer moraines, which form at, or close to the grounding line of calving glaciers [e.g. Lindén and Möller, 2005]. The scale of the landforms is consistent with that of

De Geer moraines, as is their pattern trending across topographic undulations [Todd et al., 2007]. Given the abundant evidence for ice-dammed lakes in the area, and local borehole records of interbedded sands, silts and till, the landforms are interpreted as De Geer moraines.

A simple estimate of calving speed can be calculated across a hypothetical calving margin similar to the one indicated by the De Geer moraines at Kilmarnock. Warren and Kirkbride [2003] described an empirical linear relationship between water depth (D_W) and calving speed (U_C) for glaciers terminating in freshwater bodies: $U_C = 17.4 + 2.3D_W$. D_W can be approximated from the $\sim$ 150 m a.s.l. upper altitude of De Geer moraines (proxy for lake surface altitude) and the base of the Ayr valley (30 m a.s.l.). Assuming a similar calving margin relationship, former calving rate is calculated to have been 293 m a^{-1} across the deepest part of the lower Ayr valley. Under steady state conditions (ice front not retreating, nor advancing), ice velocity at the margin would have been $\sim$ 290 m a^{-1}. These values are comparable with those of modern glaciers terminating in proglacial lakes on the eastern side of the South Patagonian Icefield [e.g. Warren and Aniya, 1999]. Ice flow velocities of this order are consistent with the local development of well-preserved streamlined bedforms assigned to flowset-V (Figs. 4.5, 4.8). The survival, locally, of extended Devensian sequences suggests that fast flow was enabled by basal sliding, and a relatively thin, near surface deforming layer.

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

4.7 Towards a regional synthesis

In order to put our results into a more regional context (Fig. 4.17) we briefly compare and test our deductions with some of those published recently for surrounding segments of the former BIIS. Importantly, our history of events for west central Scotland is consistent with the paradigm of a mobile, dynamic BIIS [Bowen et al., 2002; Bradwell et al., 2008b; McCabe, 2008; Evans et al., 2009; Greenwood and Clark, 2009].

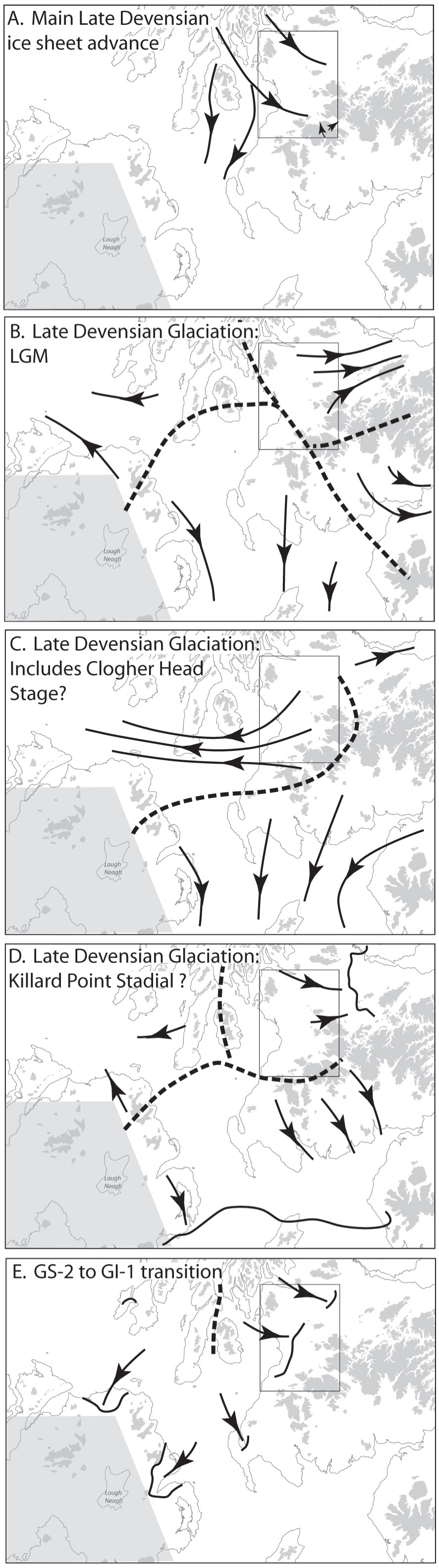


FIGURE 4.17: Proposed evolution of the western sector of the last BIIS. Arrows denote former ice flow directions. Dashed lines indicate approximate positions of ice divides. See text for discussion.

Recent numerical modelling experiments [Hubbard et al., 2009] simulate initial ice advance from the northwest into the Clyde and Ayrshire basins, accompanied by independent ice cap development over the Southern Uplands. Our hypothesis for precursor ribbed moraine development partially through accumulation and over-riding of ice marginal sediments also requires Southern Uplands ice to have remained a confined, independent mass while northwest-sourced ice entered the Clyde and Ayrshire basins(Fig. 4.14A). Further support for this early configuration comes from erratic transport paths (Fig. 4.15) where a distinct limit of Highland-sourced erratics has been identified (Fig. 4.5) [Eyles et al., 1949]. Joining of the two ice masses is unlikely to have occurred until northwest-sourced ice reached at least the south-eastern fringes of the Ayrshire basin. The subsequent development of a substantial ice divide over Arran and the Firth of Clyde with eastward flow across central Scotland by the LGM (Fig. 4.17B) has similarities with the recent BIIS reconstruction for northern England [Evans et al., 2009] in which ice sourced over southwest Scotland and the Lake District is driven eastwards through the Stainmore and Tyne gaps. Importantly, there is no evidence for eastward transport of Arran and Ailsa Craig erratics (Fig. 4.15), limiting the westernmost position of the ice divide to the vicinity of Arran.

There is no direct evidence to constrain the timing of ice divide migration towards the east following the LGM (Fig. 4.17C). However, there is evidence of post LGM enhanced drawdown of ice towards shelf-edge fans on the continental shelf to the northwest [Bradwell et al., 2008b] and into the Irish Sea basin [Eyles and McCabe, 1989; Roberts et al., 2007]. Our reconstruction of westward flow over the North Channel towards Ireland at this time is consistent with the view of Salt and Evans [2004]. Post-LGM, convergent westward flow may have been similar to that of an ephemeral ice stream [Stokes et al., 2009] responding to break up and calving offshore to the northwest of Ireland. This interpretation is consistent with the findings of Greenwood and Clark [2009] that once the ice sheet was established, geometry was largely controlled by fast flow / streaming corridors, which in this instance forced the ice divide to the east.

The past two decades have seen considerable advances towards understanding the dynamics and deglacial history of the Irish sector of the last BIIS, largely through the work of McCabe and co-workers [Knight and McCabe, 1997b; McCabe et al., 1998, 2005, 2007b]. Bedform patterns demonstrate that an ice sheet dome existed in the vicinity of Lough Neagh for much of the glacial cycle (Fig. 4.17B,C,D) [Knight, 2002] and a variety of inverse ice sheet models reconstruct an ice ridge over the North Channel linking the Southern Uplands and Lough Neagh dispersal centres during, and following, the LGM [Boulton et al., 1991, 2002]. Two major readvances interrupted decay of the Irish Ice Sheet: the Clogher Head Readvance (c. 15.0 - 14.2 ^{14}C, 18.5 -

16.7 cal ka BP), and the Killard Point Readvance (c. 14.2 - 13.0 ^{14}C, 17.1 - 15.2 cal ka BP), the latter believed to be a direct response to Heinrich Event 1 in the North Atlantic [McCabe et al., 1998; McCabe and Clark, 2003; McCabe et al., 2007b]. We speculate that the strong westward ice flow during stage C (Fig. 4.17C) may have been in operation during deposition of a moraine at Corvish, County Donegal, during the Clogher Head Readvance [McCabe et al., 2007b].

Our interpretation of subsequent westward migration of the Forth-Clyde ice divide towards Kintyre followed by topographically constrained eastward ice flow (Fig. 4.17D), is consistent with aspects of the reconstruction by Salt and Evans [2004] (their stages F and G). Renewed, climatically-driven ice sheet growth over northeast Ireland during the Killard Point Stadial has been suggested by McCabe et al. [1998], and is supported by recent cosmogenic exposure ages of 15.6 10Be ka BP from moraine sequences in north-western Ireland [Clark et al., 2009b]. It is possible that the thickening of ice over Arran and the Firth of Clyde deduced here occurred at a similar time (Fig. 4.17D).

It is noteworthy that the ice limits during our Stage D [and stage 3 of Paterson et al., 1998] are consistent with the ice sheet configuration in central Scotland suggested by Sissons [1963, 1964, 1967b] during the hypothesised 'Perth Readvance'. All require the presence of a large ice mass over the Firth of Clyde during deglaciation. Evidence for a significant readvance at Perth was questioned by Francis et al. [1970]; Paterson [1974]; Price [1983]; Sutherland [1984], and the concept was rejected by Sissons [1976]. McCabe et al. [2007a] recently interpreted evidence to support a readvance at Perth, which they correlate with the Killard Point Stadial in Ireland, concluding that it indicated an ice-sheet wide response to North Atlantic climate forcing. However, the evidence at that location remains open to interpretation (see comments by Peacock et al. [2007] and reply from McCabe et al. [2007c]). The evidence presented here cannot support nor refute that a more widespread readvance of the eastern ice margin took place at this time. However, the configuration depicted (Fig. 4.14D, 4.17D) would have had the effect of isolating ice masses on the eastern side of the Clyde-Forth drainage divide from their western source, possibly leading to development of widespread 'ice stagnation' glaciofluvial topography, initially cited as one piece of evidence for the readvance [Sissons, 1964].

Local readvances have been proposed to have occurred during deglaciation at Blackrock Ridge, at the head of Loch Indaal, Islay [Peacock and Merritt, 1997a], Stranraer [Charlesworth, 1926; Peacock and Everest, 2010], and at Armoy and east Antrim on the northeastern Irish coast [McCabe, 2008]. We suggest that moraine building at Blantyreferme and Kilmarnock (Fig. 4.5), occurred during the same overall phase of events. Ice retreat from Loch Indaal occurred possibly only a few hundred radiocarbon years before the beginning of the Lateglacial Interstadial (GI-1) [Peacock, 2008], placing tentative chronological constraints on these late ice margin oscillations. Work in progress suggests that the outer Firth of Clyde was probably deglaciated before the opening of GI-1, at c. 14.7 cal ka BP (J.D. Peacock, personal communication, 2009), with deglaciation of the Glasgow region occurring some time after.

The Irish record suggests radically different local ice sheet geometries during build-up and decay [McCabe, 2008; Greenwood and Clark, 2009]. In contrast, the ice sheet decay geometry in west central Scotland is reconstructed to have been similar to the build-up configuration (Fig. 4.14A, E). This was likely a result of proximity to the western Highlands, where the ice sheet was well situated to survive rises in equilibrium altitude during initial warming. Thus, the western Highlands and parts of Argyll were able to remain an important source area nourishing late stage ice margin oscillations.

4.8 Summary of regional events

Following ice sheet build up (Fig. 4.17A), a centre of relatively immobile ice existed over Argyll and west central Scotland (Fig. 4.14B). Ice from this centre later linked with dispersal centres over Lough Neagh, in Ireland, and over the hills of southwest Scotland and the Lake District. Ice was driven eastward towards the Firth of Forth and through the Stainmore and Tyne Gaps (Fig. 4.17C) [Evans et al., 2009]. Ice divides then migrated both eastwards and southwards as a result of enhanced drawdown of ice towards shelf-edge fans on the continental shelf, to the northwest, and into the Irish Sea basin (Fig. 4.17C). This reorganisation severely reduced the power of eastward flow towards the Firth of Forth and resulted in the generally accepted, relatively early deglaciation of eastern Scotland. A reversal of ice flow also occurred within the Vale of Eden and Solway Lowlands.

A major ice-surface high and ice divide developed over the outer Firth of Clyde, possibly during the Killard Point Stadial of Ireland. The ice divide probably linked with dispersal centres over Lough Neagh and the Southern Uplands (Fig. 4.17D). The ice sheet surface now descended from west to east over west central Scotland. On southern fringes of the Southern uplands, ice flow became topographically constrained [Salt and Evans, 2004], extending into the Solway Firth [Evans et al., 2009].

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

4.9 Conclusions

The main conclusions from this research are as follows:

- Published dates on preserved interstadial organic deposits show that the Main Late Devensian (MLD) (MIS 2) glaciation of central Scotland began after 35 ka cal BP. Some deposits of an earlier glaciation (MIS 4 or older) occur locally within the Clyde and Ayshire basins.
- During a sustained build-up phase, ice advanced from the western Scottish Highlands into the Clyde and Ayrshire basins. Glaciomarine muds and shelly deposits scavenged from the Firth of Clyde were redeposited widely across Ayrshire. Ice advance against reverse slopes enabled the build up of marginal sediment accumulations. Some of these accumulations probably formed pre-cursor ridges, moulded into suites of ribbed moraine by subsequent over-riding.
- Sustained stadial conditions at the Last Glacial Maximum (LGM) (30-25 ka cal BP) resulted in development of a major dispersal centre over the Southern Uplands and deflection of Highland ice towards the east and northeast. Relatively immobile ice beneath an ice-surface 'high' positioned over Ayrshire and the western Clyde basin, preserved previously-formed subglacial landforms and fed a wide corridor of fast-flowing ice towards the Firth of Forth.
- A substantial re-configuration of the ice surface over west central Scotland was caused by enhanced westward drawdown into the outer Firth of Clyde and eastward migration of an ice divide towards the Clyde-Forth watershed. This reorganisation is tentatively correlated with the Clogher Head Readvance established in the north of Ireland (c. 15.0 - 14.2 ^{14}C, 18.5 - 16.7 cal ka BP).
- Renewed ice sheet thickening over the Firth of Clyde may have accompanied growth of the Irish Ice sheet during the Killard Point Stadial (c. 14.2 - 13.0 ^{14}C, 17.1 - 15.2 cal ka BP). Subsequent ice sheet retreat was initially characterised by substantial meltwater production, ponding and erosion.

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

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

Chapter 5

Growth and decay of a marine terminating sector of the last British-Irish Ice Sheet: a geomorphological reconstruction

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Abstract

The boundary conditions that govern ice sheet dynamics can change significantly with the development of marine margins. This paper uses the glacial landscape in western Scotland to investigate changes in the British-Irish Ice Sheet (BIIS) that accompanied the growth and decay of a marine sector over the Malin Shelf. Ice advanced from a restricted mountain ice sheet with tidewater margins after ∼35 ka BP, and reached the continental shelf in ∼7 ka (average rate of ∼30 m a^{-1}). Early ice flow had been directed through north-south, geologically controlled, over-deepened fjords that were carved during previous 'restricted' glaciations. This flow regime was abandoned with development of the Malin Shelf ice sheet sector; ice flow direction changed by 90° and was drawn down towards the shelf edge. The marine ice sheet phase saw episodes of ice divide migration by up to 60 km over west central Scotland, possibly linked to ice streaming and calving events at the ice sheet margin. However, permanent and stationary ice divides and zones of cold-based ice, associated with subglacial topographic highs, also characterised the marine glacial stage over western Scotland. The North Channel ice divide remained a constant, though migratory feature while the BIIS occupied the Malin Shelf; it finally collapsed at the end of the Killard Point Stadial when the Irish Ice Sheet began to rapidly decay, ∼16.5 ka BP. This

permitted the Scottish Ice Sheet to temporarily advance over north-east Ireland (previously identified as the East Antrim Coastal Readvance) before it too retreated, at rates in the order of 10^2 m a^{-1}. Although the imprint of extensive shelf-edge ice sheet glaciation exists in the coastal landscape of western Scotland, the dominant landscape features relate to a restricted, marine-proximal mountain ice sheet with markedly different flow configurations. Similar first-order geomorphological features, relating to 'restricted' glacial conditions, are likely to be preserved in subglacial highlands under interior parts of modern ice sheets.

5.1 Introduction

The geological record left by past ice sheets provides information about their long-term evolution and interaction with the landscape over timescales beyond that of contemporary glaciological observations [Boulton and Clark, 1990; Kleman et al., 2008, 2010]. Large-scale ice sheet reorganisations identified in palaeoglaciological studies therefore add important context to recent changes seen in modern ice sheets [Retzlaff and Bentley, 1993; Conway et al., 2002], and can play a role in predicting their future evolution as we discover more about the landscapes they submerge [Ross et al., 2012]. Parts of the West Antarctic Ice Sheet (WAIS), for example, rest on complex topography, with deep basins in close proximity to subglacial highlands, which have been suggested to possess characteristics of former marine-proximal alpine glaciation (e.g. the Ellsworth Subglacial Highlands) [Holt et al., 2006; Vaughan et al., 2006; Ross et al., 2014]. Linking these new findings about the subglacial topographic setting of the WAIS with longer-term (10^4 yr) ice sheet dynamics is an exciting area of research, and one in which insights from former ice sheets can contribute.

The British-Irish Ice Sheet (BIIS) is known to have had marine or partially marine sectors, which have been suggested to be analogous to the present West Antarctic Ice Sheet, although smaller in scale [Bradwell et al., 2008b; Graham et al., 2009; Clark et al., 2012]. Recent systematic assessments utilising high-resolution elevation datasets have considerably advanced our understanding of the overall configuration and flow paths during retreat of the BIIS [Clark et al., 2012]. However, detailed time transgressive reconstructions of flow geometries and configurations during ice sheet build up and collapse do not yet exist for a number of important ice sheet sectors. Comprehensive investigations combining remote-sensing- and field-based investigations [eg. Livingstone et al., 2009] can provide this information and reveal how an evolving ice sheet interacted with its bed [e.g. Sugden, 1968; Hall and Sugden, 1987; Kleman and Glasser, 2007; Golledge et al., 2009], thereby providing a key link between long-term ice dynamics and the subglacial landscape.

In this paper we examine the geomorphological record from the peninsula of Kintyre and the adjacent island of Arran (combined area of $\sim$ 825 km^2), at the transition between the fjord-like coastal terrain of the western Scottish Highlands and the Malin Shelf to the west (Figs. 5.1, 5.2), in order to reconstruct BIIS behaviour through the last glacial cycle. The area is ideally suited for detailed palaeoglaciological examination since: (i) the position of western edge of the BIIS meant that it was particularly sensitive to changes in oceanic and atmospheric circulation that characterised the North Atlantic region during the last glacial cycle [Rahmstorff, 2002; McCabe, 2008]; (ii) Kintyre and Arran contain a variety of landforms and sediments, some of which have been suggested to pre-date the growth of the last ice sheet, therefore providing insight into the extent of landscape modification that took place during the last glacial cycle; (iii) the southernmost point of Kintyre, the Mull of Kintyre, lies just 20 km from the Irish mainland, providing a unique link between the terrestrial geomorphological records of south-west Scotland and north-east Ireland, with the potential to greatly improve our understanding of the break up of the BIIS over the North Channel; and (iv) published data exist for adjacent parts of the BIIS [e.g. Greenwood and Clark, 2009; Dunlop et al., 2010; Finlayson et al., 2010, this thesis' Chapter 4], which can be combined in a larger-scale synthesis of the advance and collapse its western margin. Despite these research opportunities, Kintyre and Arran have received little recent geomorphological examination in relation to the BIIS. The goal of this paper, therefore, is to review and re-examine the glacial geomorphology of Kintyre and Arran, and combine new data with published studies to examine the nature and scale of changes in the BIIS associated with the growth and decay of its western marine margin.

5.2 Setting

5.2.1 Geology and relief

Kintyre is a 68-km-long, north-south trending peninsula in the south-west of Scotland (Figs. 5.1, 5.2). It is no more than 19 km wide at any point and is bounded to the west by the Sound of Jura (200 m below sea level (b.s.l.)), to the east by the Kilbrannan Sound (120 m b.s.l.), part of the outer Firth of Clyde, and to the south by the North Channel, a tectonic basin up to 300 m b.s.l. [Maddox et al., 1993]. West Loch Tarbert separates Kintyre from the Knapdale region to the north. Most of the solid rocks underlying Kintyre consist of psammites, semipelites and pelites belonging to the Dalradian Supergroup. In central- and north-western parts of Kintyre, these rocks possess a broad north-south trending strike, which is visible on digital surface models

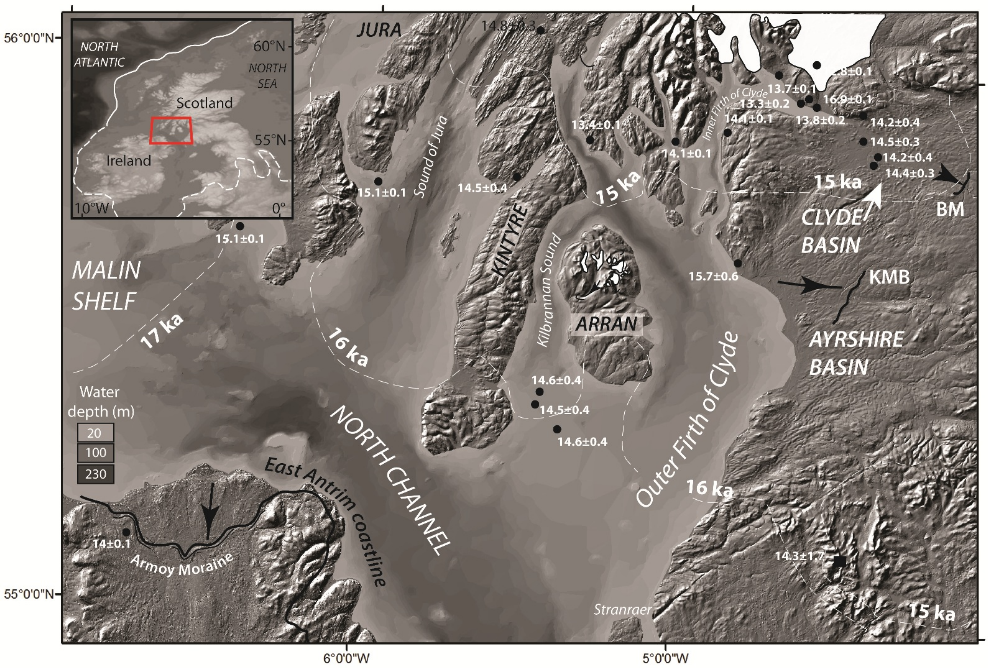

FIGURE 5.1: Location of the Kintyre Peninsula and Island of Arran, between the fjord coastline of western Scotland and the Malin Shelf to the west. KMB: Kilmarnock moraine belt; BM: Blantyreferme moraine. 15, 16 , and 17 ka ice retreat isochrones are taken from Clark et al. [2012]. Calibrated radiocarbon ages (black circles) from the database of Hughes et al. [2011a] and from Peacock et al. [2012]. Areas in white show maximum glacier extent during the Younger Dryas (12.9-11.7 ka BP), based on Clark et al. [2004] and Ballantyne [2007a]. Hill-shaded surface models built from Intermap Technologies NEXTMap Britain elevation data and Land and Property Services mapping data (Crown Copyright). Bathymetry from BGS Digbath-250 dataset. Inset: Location within a national context. The white line gives the approximate extent of the last BIIS, based on Bradwell et al. [2008b] (solid line) and Clark et al. [2012] (dashed line).

(Fig. 5.3). The central spine of the peninsula generally ranges between 100 m and 450 m above sea level (a.s.l.) in elevation. It is separated by a low-lying corridor, 10-50 m a.s.l., between Campbeltown Loch and Machrihanish Bay, where the underlying rocks consist of Carboniferous sandstones and lavas. Devonian conglomerate is present under the south-eastern corner of the peninsula and outcrops of Permian sandstone are present along parts of the western coastline, both resting unconformably on the underlying Dalradian rocks.

The Island of Arran (435 km^2) is separated from Kintyre by the Kilbrannan Sound and bounded to the east by the North-east Arran Trough (170 m b.s.l.)(Figs 5.1, 5.2). The northern half of the island is dominated by the Northern Granite Pluton, which was intruded into Dalradian metasediments and Devonian sandstones during the

Tertiary Period. The pluton comprises an outer coarse-grained granite and an inner fine-grained granite. It now forms an elevated massif, which is alpine in character with steep-sided corries, valleys and arêtes, and several summits that exceed 700 m – the highest being Goatfell (874 m). These northern hills are surrounded by a well-developed surface at approximately 300 m in elevation, known as the 'Thousand Foot Platform' [Tyrrell, 1928]. This surface, which crosses geological boundaries, possesses immature drainage, and is cut by glaciated valleys, has been suggested to be part of a preglacial, possibly Pliocene age, plateau [Gregory, 1926; Tyrrell, 1928]. The bedrock surface on the southern half of the island principally comprises Devonian, Permian and Triassic sandstones, with a smaller central granitic intrusion and numerous sill complexes. In the south of the island the relief rarely exceeds 400 m in elevation.

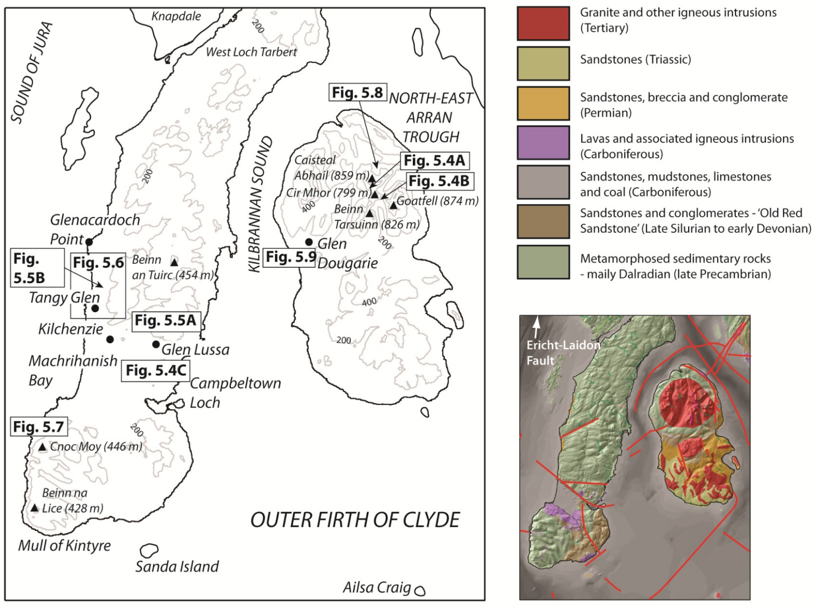

FIGURE 5.2: Topography and simplified bedrock geology of Kintyre and Arran. Red lines on the geology map indicate faults.

5.2.2 Glacial history

5.2.2.1 Pre-Main Late Devensian sediments and landforms

Sediments and landforms, which have been interpreted to pre-date the last major glacial cycle (the Main Late Devensian (MLD), Marine Isotope Stage 2, Greenland Stadial 5-1 [Lowe et al., 2008]) have been reported from Kintyre. Shell-bearing clays underlie till at three sites in and around Tangy Glen (Fig. 5.2) on the west coast

(Horne et al. 1896). These clays, found at elevations of between 40 and 60 m a.s.l., were reported to contain molluscs, ostracods and forams indicative of both arctic and warmer temperate environments, and were argued by Munthe [1897] to record a period of deposition spanning a glacial-interglacial-glacial transition. The shelly clays have subsequently been interpreted as being either *in situ* remnants of Middle Quaternary marine deposits from a period of significantly higher relative sea levels [Sutherland, 1981], or emplaced as a glacial raft by the advancing MLD ice sheet [Synge and Stephens, 1966]. A rock platform at 13 m a.s.l. also exists underneath till at Glenacardoch Point on the west coast [Sinclair, 1911; Gray, 1978]. The platform is one of the few sites in Scotland where low-level shore platforms pass beneath till, and it has been suggested to relate to an interglacial period pre-dating the last glacial cycle [Sissons, 1981; Gray, 1993].

Deposits containing both cold and warm water shells have also been discovered under and within till in the south of Arran, at elevations up to 55 m a.s.l. [Watson, 1864; Bryce, 1865]. Sutherland [1981] argued that the shell beds cannot have been transported glacially and are largely *in situ*, because they are present in an area where ice flow indicators on the land surface show that the last ice movement was towards, not from, the sea. However, an *in situ* interpretation is not consistent with the original descriptions of the sediments by Watson [1864], who wrote that, 'the layers of sand curve sharply upon themselves, as if they had been thrust forwards under a heavy weight from behind, and forced to over-ride one another'. Furthermore, recent MLD ice sheet reconstructions depict a stage of west-north-westward ice flow, presenting at least one possible mechanism for the transport of sediments from the sea across the southern edge of Arran [Salt and Evans, 2004; Finlayson et al., 2010; Livingstone et al., 2012a].

5.2.2.2 The Late Devensian glacial cycle (MIS 2, Greenland Stadial 5-1)

Early research on Kintyre used erratic dispersal patterns and glacial striae to recognise that the peninsula had been predominantly overridden by ice flowing westward towards the Malin Shelf during the MLD [Horne et al., 1896; Geological Survey of Scotland, 1913]. Synge and Stephens [1966] suggested that this westerly flow was preceded by an advance from the north, presumably directed along the deep rock basins of the Sound of Jura and Kilbrannan Sound, which had 'plugged' Tangy Glen with the shelly deposits. These authors also considered the final movement of ice on Kintyre to have been north to south, proposing that a former ice limit formed 'thick morainic accumulations' near Kilchenzie on the west coast. A general north to south pattern of ice movement through the Kilbrannan Sound and Firth of Clyde is also evident from striae on Arran, although this flow was diverted around the high ground where an independent ice dome was nourished during the MLD [Tyrrell, 1928; Gemmell, 1973].

There are no available dates from Kintyre to constrain the timing of deglaciation. However, dated samples obtained from sediment cores in surrounding marine waters indicate that postglacial sediment accumulation had begun by 13.1 - 12.7 ^{14}C (14.9 - 14.5 cal) ka BP [Peacock, 2008; Peacock et al., 2012] (Fig. 5.1). McCabe and Williams [2012] have recently proposed that deglaciation of the western central zone of the last BIIS was punctuated by a major 'North Channel Readvance', c. 15-15.5 cal ka BP, which they suggest formed coeval moraines in East Antrim, Stranraer, and the Ayrshire and Clyde basins (Fig. 5.1). These authors envisaged general westward or south-westward ice flow over Kintyre at that time. No subsequent glacier margin readvances or stillstands have been identified on Kintyre. However, two subsequent advances of locally-nourished glaciers took place on Arran, the latter during the Younger Dryas (12.9-11.5 ka BP) [Ballantyne, 2007a].

5.3 Methods

A combined remote sensing and field-based approach was employed to characterise the subglacial and ice marginal geomorphological assemblages on Kintyre and Arran. In order to refine the deglacial chronology in the area, ice marginal landform assemblages were sampled for cosmogenic dating.

5.3.1 Remote sensing evidence

Glacial landforms were mapped within a GIS environment, using a combination of hill-shaded surface models (DSMs) derived from the NEXTMap Britain elevation dataset, georeferenced 1:10,000 scale, colour aerial photographs, and offshore bathymetry from the BGS Digbath-250 dataset. The NEXTMap Britain DSM has a 1.5 m vertical and 5 m horizontal resolution and was viewed at scales ranging from 1:10,000 to 1:100,000. A sub-sampled version of the DSM, with a horizontal resolution of 50 m, was also used for investigation at scales of greater than 1:100,000. The DSMs were illuminated from both the north-west and north-east in an attempt to reduce the effects of azimuth biasing [Smith and Clark, 2005]. The landforms that were recorded during the remote sensing survey include: major rock basins and troughs, streamlined bedforms, eskers, meltwater channels, moraines, and deltas. The presence and general trend of bedrock structures at the land surface were also noted as a crude indicator for the presence of sediment cover, and for its orientation relation relative to streamlined bedforms.

5.3.2 Field evidence

Field mapping was carried out on Kintyre and parts of Arran in 2010 using a ruggedized tablet PC with a built-in GPS and ArcGIS software. The field mapping enabled verification of landforms identified during the remote sensing survey and helped identify smaller features that were not visible using the remote sensing datasets, such as tors, glacial erratics, and smaller moraines. Natural sections were also logged during the field investigation.

5.3.3 Compilation and utilisation of geomorphological data

All features observed during the remote sensing and field investigations were captured within a spatially attributed GIS database. Trommelen et al. [2012] highlighted the importance of integrating remotely-sensed and field-based geomorphological data in their Glacial Terrain Zone approach. This is particularly true when dealing with fragmented palaeoglaciological records, such as those found elsewhere in western Scotland [Salt and Evans, 2004; Finlayson et al., 2010]. The data were collectively used to infer different glaciological conditions based on established process-form relationships. This 'inversion' approach is a well-established tool in palaeoglaciological reconstruction [Kleman and Borgström, 1996; Kleman et al., 1997; Stokes et al., 2009]. Landforms and sediments that were produced, or survived, under the ice sheet allow inferences to be made about the action of the ice sheet on its bed. Consistently aligned clusters of streamlined bedforms may be grouped as 'flow sets' and used to infer episodes of warm-based ice sheet motion in a particular direction [Boulton and Clark, 1990; Kleman et al., 1997; Livingstone et al., 2009; Stokes et al., 2009]. Marginal landforms such as moraines and meltwater channels can be used to interpret patterns of ice margin retreat [Clark et al., 2012].

5.3.4 Cosmogenic nuclide analysis

A number of radiocarbon ages constrain the deglaciation chronology in the inner Firth of Clyde [Hughes et al., 2011a] (Fig. 5.1). However, fewer ages constrain the timing of deglaciation in the outer Firth of Clyde, and in particular, the decay of ice across the North Channel. In an attempt to improve chronological constraints on deglaciation, boulders from Glen Dougarie in western Arran and Glen Lussa in eastern Kintyre were sampled for cosmogenic nuclide analyses (Fig. 5.2). In Glen Dougarie, two granite erratics from the top of two linked low lateral moraines (50 m apart) at 45 m a.s.l. were sampled in order to date the formation of the moraines. Although a number of Arran granite erratics are present on Kintyre, difficulties were encountered finding suitable samples with a correct (ice marginal landform) context in areas not affected by anthropogenic activity. No single landform with granite erratic boulders on top was identified; as a result samples in Glen Lussa were taken from three granite erratics resting on gently undulating ground, within a wider area of deglacial features, comprising meltwater channels, boulder spreads and low ridges. Since the samples do not specifically relate to any ice marginal landform, they were collected to provide a minimum age for the ground becoming free of glacier ice. Skyline topography was measured in the field at 15 degree increments at all of the sample locations to allow calculation of topographic shielding.

The samples were prepared at the University of Glasgow Cosmogenic Isotope Laboratory at the Scottish Universities Environmental Research Centre (SUERC). Beryllium was extracted from Quartz, which was separated and purified following modified procedures adopted from Kohl and Nishiizumi [1992]. BeO targets were prepared for ^{10}Be/^{9}Be analysis using procedures modified from Child et al. [2000]. Between 215 and 219 μg Be was added as carrier and between 20 and 25 g of each sample was dissolved. The ^{10}Be/^{9}Be ratios were measured with the 5 MV accelerator mass spectrometer at SUERC [Xu et al., 2010]. ^{10}Be/^{9}Be ratios were normalised to NIST SRM 4325 with a ^{10}Be/^{9}Be ratio of $2.79 * 10^{-11}$ [in agreement with Nishiizumi et al., 2007]. Process blanks prepared with the samples yielded an average ^{10}Be/^{9}Be ratio of $4.1 * 10^{-15}$. Blank-corrected ^{10}Be/^{9}Be ratios of the samples ranged from 53 to $114 * 10^{-15}$. Total one-sigma uncertainties for the concentrations determined at the SUERC-AMS Laboratory include the one-sigma uncertainty of the AMS measurement and a 2% uncertainty as a realistic estimate for possible effects of the chemical sample preparation, which includes the uncertainty of the Be concentration of the carrier solution. Exposure ages were calculated using the CRONUS-Earth online calculator [Developmental version; Wrapper script 2.2, Main calculator 2.1, constants 2.2.1, muons 1.1; Balco et al., 2008] and calibrated using a locally derived ^{10}Be production rate based on ^{10}Be concentration in samples from erratic boulders on the terminal moraine of the Loch Lomond glacier advance [Fabel et al., 2012], approximately 75 km from the sites in this study. These sample ages are independently controlled by the radiocarbon ages of microfossils associated with a varve sequence deposited in a glacial lake at the time that the Loch Lomond moraine formed [Macleod et al., 2011]. The calculated ^{10}Be concentrations from the moraine boulders resulted in a reference ^{10}Be production rate of 3.92 ± 0.18 atoms $g^{-1}a^{-1}$. The exposure ages reported here (Table 5.1) are based on the time-dependent Lm scaling scheme of the CRONUS-Earth online calculator [Lal, 1991; Stone, 2000], and assumption of a sampling surface erosion rate of 0 mm ka^{-1}. For exposure ages <20 ka, the other scaling schemes (the St, Du, De and Li schemes) available via the online calculator produce ages that differ on average from the Lm scheme by less than 1% of sample age. Similarly, for ages <20 ka, assumption of an erosion rate of 1 mm ka^{-1} increases our calculated exposure ages by 1.1%.

Sample	Latitude (°N)	Longitude (°W)	Altitude	Thickness (mm)	Horizon correction	$[^{10}Be]^a$ (10^4 atoms g^{-1} SiO_2)	^{10}Be exposure age (ka)^b
GL1	55.4877	-5.6232	135	20	0.9988	6.439 ±0.352	13.77 ±0.98
GL2	55.4886	-5.6219	128	20	0.9997	6.046±0.241	13.00 ±0.79
GL3	55.49	-5.6269	130	20	0.9992	7.004 ± 0.281	15.05 ± 0.91
D1	55.592	-5.3369	47	20	0.9965	6.755 ± 0.263	15.86 ± 0.95
D2	55.592	-5.3368	49	20	0.9969	7.090 ± 0.255	16.60 ± 0.97

TABLE 5.1: Exposure ages from sampled granite erratic boulders. [a]Isotope ratios normalized to NIST SRM 4325 with a value of $2.79 * 10^{-11}$ [Nishiizumi et al., 2007]. Uncertainties are propagated at the 1σ level and include all known sources of analytical error (blank, carrier mass and counting statistics). A density of 2.65 g cm^{-3} is assumed for all samples. All samples are from the upper surfaces of glacially deposited boulders. [b]Calculated ages are scaled using zero erosion and the Lm scheme of the CRONUS online calculator (Balco et al., 2008), wrapper script version 2.2, main calculator version 2.1, constants version 2.2.1, muons version 1.1, with a ^{10}Be half life of $1.387 * 10^6 years$ [Chmeleff et al., 2010; Korschinek et al., 2010], and a local sea level high latitude production rate of 3.92 ± 0.18 atoms g^{-1} a^{-1}.

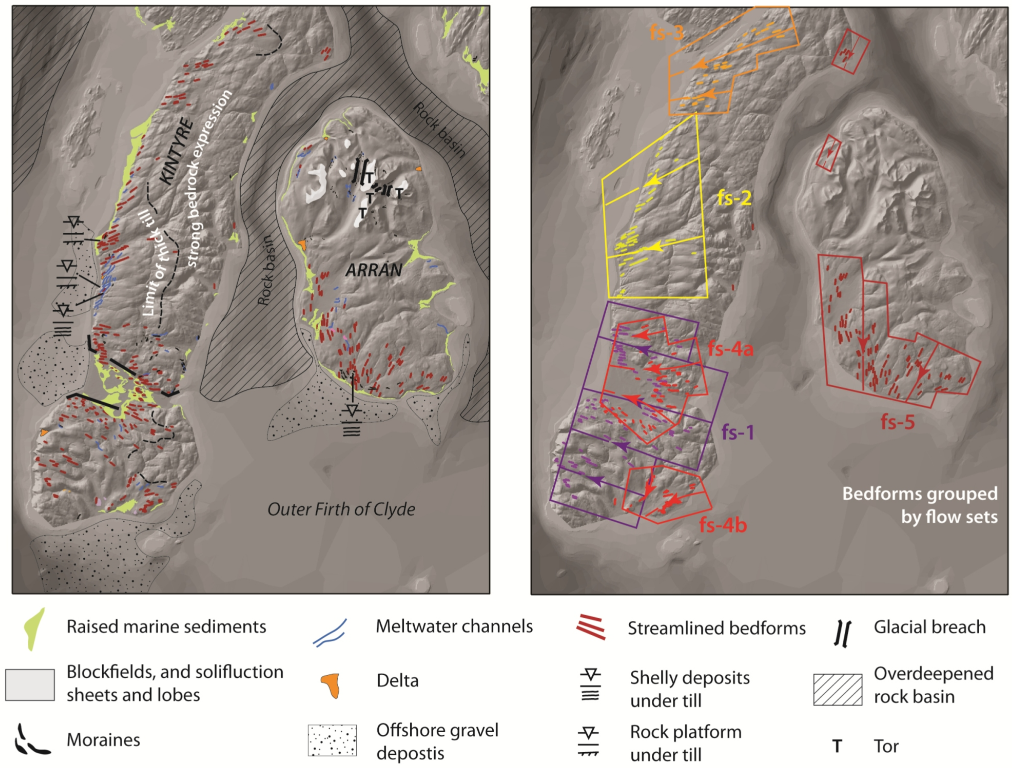

FIGURE 5.3: Glacial geomorphology of Kintyre and Arran. Distribution of raised marine sediments and offshore gravel deposits compiled from published BGS maps. Right hand panel shows streamlined bedforms grouped into flow sets (fs) (Table 5.2).

5.4 Geomorphology and sediments

Glacial geomorphological features are synthesised in Figure 5.3. The details of individual assemblages are described below.

5.4.1 Subglacial assemblages

5.4.1.1 Tors

Well developed granite tors are present on some of the highest summits on Arran, such as Caisteal Abhail (859 m a.s.l.), known as 'The Castles' (Fig. 5.4A), and Beinn Tarsuinn (826 m a.s.l.). These tors are high relief (up to 10 m), and possess delicately balanced blocks and deep joint sets. Large granite tors elsewhere in Scotland have been shown to develop over long periods (10^5-10^6 years), requiring preservation during the glacial cycles of the middle and late Quaternary [Phillips et al., 2006]. The tors on

Arran exist in close proximity to major, north-south aligned, erosional breaches on the island (see below). Glacially transported 'perched' granite boulders also exist on several of the highest summits of Arran, demonstrating that these peaks were overwhelmed by ice during maximum stages of past glaciations [Ballantyne, 2007a].

5.4.1.2 Erosional basins and breaches

Kintyre and Arran sit between three major north-south trending rock basins (Figs. 5.1, 5.3). The Sound of Jura is a basin that reaches a depth of 200 m below present sea level, closely follows the strike of the underlying Dalradian metamorphic rocks, and is located over the position of the Ericht-Laidon Fault [B.G.S., 1985]. The Kilbrannan Sound is a basin between Kintyre and Arran that reaches a depth of 120 m below present sea level, and is located in a zone where Permian and Triassic sandstones have most likely been down-faulted into the harder underlying Dalradian rocks. The basin of the Northeast Arran Trough reaches a depth of 160 m below present sea level, and is positioned over down-faulted Permo-triassic sandstones, bounded by the Sound of Bute Fault and the Brodick Bay Fault. Kintyre is also dissected by one major east-west breach between Campbeltown and Machrihanish Bay. Here the Dalradian metamorphic rocks, which form the bedrock surface for much of the peninsula, are replaced by unconformably overlying and down-faulted Carboniferous and Devonian sedimentary rocks and lavas. The contrast in land surface elevation is particularly pronounced along the Kilchenzie Fault, which marks the boundary between the Dalradian and younger rocks. In each of these cases the deepening or breach is located over fault zones, often associated with an increase in fracture density and weathering depth, or softer rocks relative to the surrounding lithologies. A series of alpine-style glacial breaches also exists on the Isle of Arran, within the mountains of the Northern Granite Pluton (Fig 5.4B). These breaches are relatively clear of weathered rock, and possess ice-moulded bedrock surfaces with perched boulders. Tyrrell [1928] noted that the main 'through' valleys tend to have an approximate north-south trend, which runs parallel to structural zones within the granite.

5.4.1.3 Streamlined bedforms

The streamlined bedforms observed on Kintyre and the south-western side of Arran comprise streamlined hills, crag-and-tails, and drumlins. These bedforms can be grouped into individual flowsets based on their alignment, geographical distribution

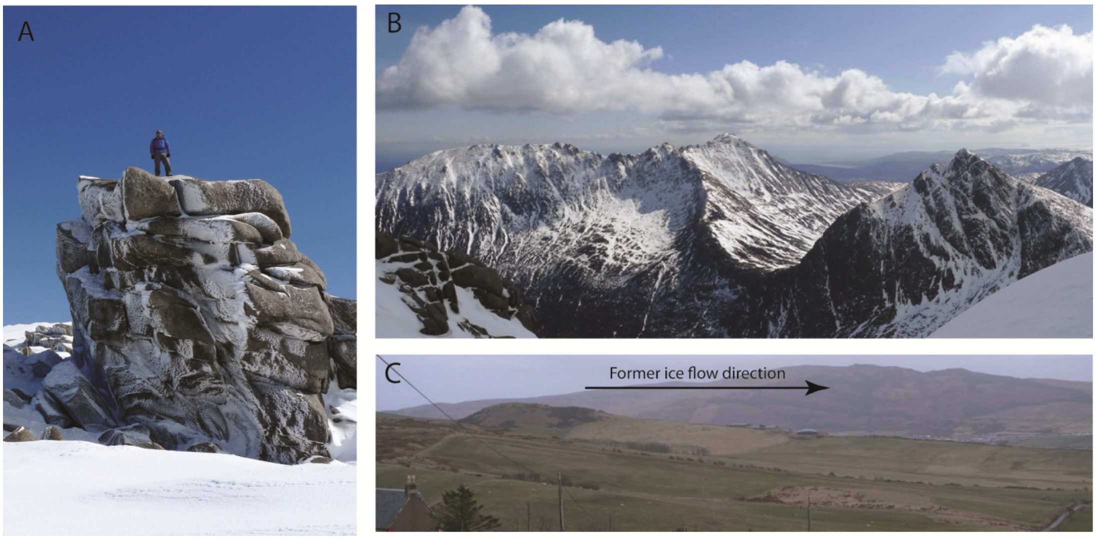

FIGURE 5.4: Examples of landforms that were preserved, modified or created under the last ice sheet. A: Tor on Caisteal Abhail, northern Arran. B: North-south directed glacial breach, northern Arran.C; Elongated crag-and-tail, southern Kintyre.

and relationship with topography (Fig. 5.3). Flow set statistics are shown in Table 5.2.

Flow set 1 Flow set 1 comprises west-north-westward aligned streamlined hills, crag-and-tails (Fig. 5.4C) and drumlins, which are present across the southern half of Kintyre. These bedforms maintain a similar alignment at all elevations on southern Kintyre, although they are absent on the far southern and south-eastern margins of the peninsula.

Flow set 2 Bedforms belonging to flow set 2 generally comprise west-south-westward aligned drumlins, streamlined hills and crag-and-tails, which are present over areas of thick till on the western central part of Kintyre. The eastward extent of these bedforms is marked by the transition from: (i) smooth, till-covered terrain on the western side of the central spine of the peninsula, to (ii) bedrock with little till cover in the east, where the north-south strike dominates morphology of the land surface. On the west coast of Kintyre, some of the flow set 2 bedforms are deeply incised by (sub)marginal meltwater channels (see below).

Flow set 3 Flow set 3 comprises west-south-westward aligned drumlins and crag-and-tails that occupy ground below 200 m a.s.l. around West Loch Tarbert. As observed for flow set 2, these bedforms are confined to the western dipping slopes to

Flow set	Elongation ratio			Centroid elevation (m)	
	Range	Mean	SD	Range	Median
1 (n = 86)	1.6-5.8	3.2	0.9	11-335	122
2 (n = 76)	2.1-7.7	3.4	1.3	31-363	81
3 (n = 52)	2.1-7.7	3.8	1.1	28-167	91
4a(n = 108)	1.2-4.9	2.7	0.9	17-154	81
4b(n = 62)	1.9-4.7	3.1	0.7	25-99	59
5 (n = 180)	1.6-5.4	3.4	0.7	15-312	85

TABLE 5.2: Streamlined bedform summary statistics. 'Centroid' refers to the middle point of each streamlined bedform.

the west of the central spine of the peninsula. Their trend is slightly oblique to the dominant south-west strike of the underlying metasedimentary bedrock.

Flow set 4 Flow set 4 comprises two subsets of crag-and-tails and drumlins on the southern half of Kintyre that are diverted around the high ground in the south-west. Flow set 4a displays a westward pattern of convergence towards Machrihanish Bay, while flow set 4b displays a south-westward convergence around the Mull of Kintyre.

Flow set 5 Flow set 5 comprises generally southward trending drumlins and crag-and-tails in south-western Arran, and sparse rock drumlins and crag-and-tails on eastern Kintyre and north-west Arran, which are locally oriented parallel to the metasedimentary bedrock strike. The drumlins and crag-and-tails on Arran show a weakly convergent pattern on the southern side of the island's southern hills.

5.4.1.4 Subglacial sediments

Thin, gravelly, shell-bearing tills have been identified locally on the eastern coast of Kintyre [Synge and Stephens, 1966]. Thick deposits of subglacial diamicton, which exceed 20 m in places, are generally only present in the west. The margins of the western distribution of thick sediment are clearly represented by the appearance of bedrock structures which can be seen at the land surface across eastern parts of the peninsula (Fig 5.3). Sediment exposures in western Kintyre generally reveal a firm to very stiff, red to dark reddish brown, massive to fissile, matrix supported, silty clay diamicton, containing predominantly sub-angular, striated and faceted clasts (Fig. 5.5A). Clast content is dominated by metasedimentary lithologies, although some volcanic and rare granitic clasts are also present. Locally the diamicton contains lenses or pods of sorted sands. In general, the thick diamicton observed in western Kintyre possesses the characteristics of a subglacial traction till [Evans et al., 2006].

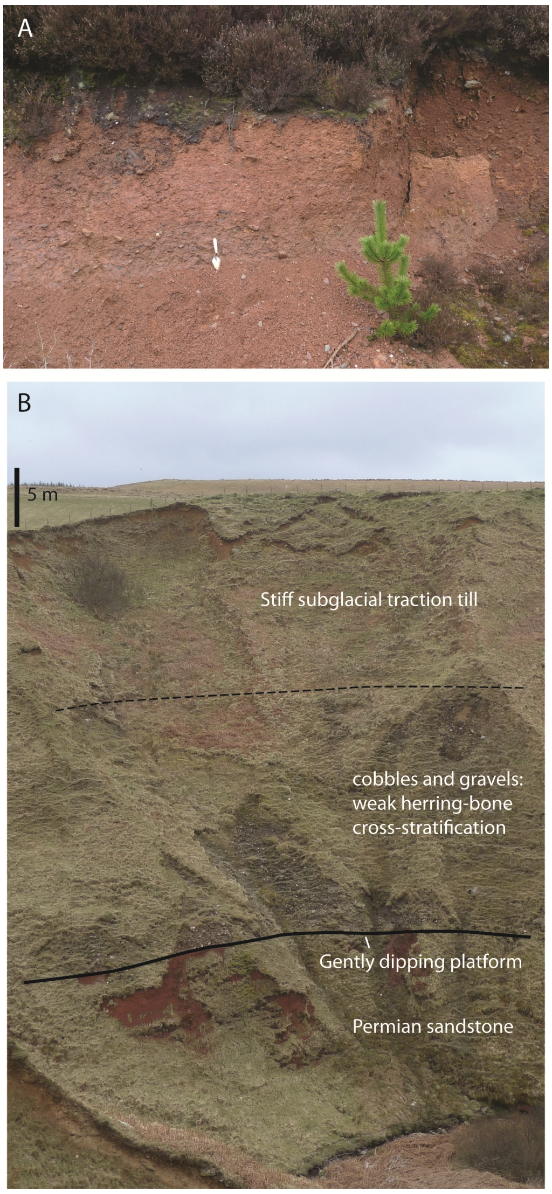

FIGURE 5.5: Subglacial sediments exposed on western Kintyre. A: stiff, red subglacial traction till, which forms thick sequences over the western central part of Kintyre. B: 15 m of subglacial traction till overlying weakly, herring bone cross-stratified gravels, interpreted as beach deposits. These rest on a platform cut into Permian sandstones at approximately 18 m a.s.l, only a few metres higher than the pre-last glacial cycle rock shore platform described by Gray [1978, 1993] at Glenacardoch Point to the north.

The three sites at Tangy Glen where shelly clays had been observed under till during the late 19th and early 20th Centuries were visited in 2010. At the time of field investigation, blue grey clays were exposed only at and below the water level of Tangy Burn. At Drumore Burn, 15 m of till was observed overlying 6-8 m of clast-supported, sub-rounded to sub-angular cobbles and gravels, with a sandy matrix (Fig. 5.5B). In places these moderately sorted gravels have a weakly developed herring-bone cross stratification. They are tentatively interpreted as beach gravels and overlie a clear platform cut into red Permian sandstone, which dips gently towards the coast. At this location, the platform surface lies at approximately 18 m a.s.l., only a few metres higher than the pre-last glacial cycle rock shore platform that was described by Gray [1978, 1993] at Glenacardoch Point to the north.

On Arran, Tyrrell [1928] noted that thick deposits of subglacial sediments are generally restricted to southern parts of the island, corresponding with the smooth, southward streamlined terrain observed on modern digital surface models (Fig. 5.3). The till in northern Arran is generally thinner and sandier than in the south. At a number of the valley mouths, pale brown to grey, granite dominated till crosses geological boundaries, indicating radial transport from the central granite complex to the coastline – an observation also made by Gemmell [1973].

5.4.2 Ice marginal assemblages

5.4.2.1 Meltwater channels

A well-preserved set of north-east to south-west trending marginal or sub-marginal meltwater channels is present over an 8 km stretch of the western coastline of Kintyre (Fig. 5.6). Individual channels are continuous for at least 3 km, their lower reaches having been erased by erosion of cliffs along the coastline. The channels are up to 150 m in width, and incise the surrounding till and the bedforms belonging to flowset 2, by up to 20 m. Isolated meltwater channels are present elsewhere on Kintyre, and Gemmell [1973] described a series of meltwater channels that descends along the western flanks Arran. In general, the meltwater channels on Kintyre and western Arran descend in an overall westward and southward direction.

5.4.2.2 Perched delta

A former delta, which is open to the North Channel, exists at an elevation of 130 m a.s.l at Innean Glen in south-west Kintyre (Fig. 5.7). It consists of 20 m of

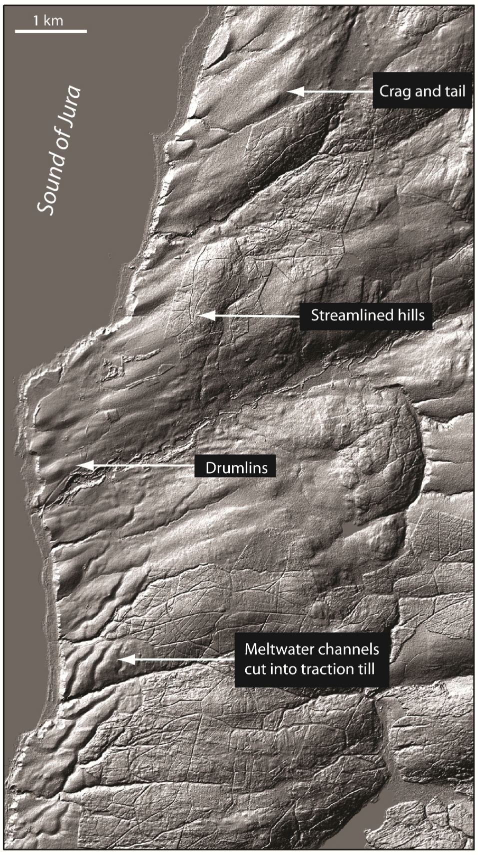

Figure 5.6: Meltwater channels (bottom left of image) dissecting west-south-west streamlined bedforms formed in subglacial till, western Kintyre. Hill-shaded surface models built from Intermap Technologies NEXTMap Britain elevation data.

westward dipping, stratified sands and imbricated gravels and cobbles, which overlie a stiff, red, matrix supported, sandy clay diamicton. The diamicton contains isolated, striated and faceted, subangular clasts, and is interpreted as a subglacial till. At 130 m a.s.l., the delta surface lies far above any lateglacial or postglacial relative sea level high stand [Synge and Stephens, 1966]. It must therefore relate to subaerial drainage ponding against a low-profile ice sheet margin that was grounded offshore, the local water depth being insufficient for floatation of ice that was at least 170 m thick (height of delta surface minus sea bed surface) at that time.

5.4.2.3 Moraines

Prominent moraines are rare on Kintyre. The 'thick morainic accumulations' near Kilchenzie, described by Synge and Stephens [1966], are interpreted here as drumlins

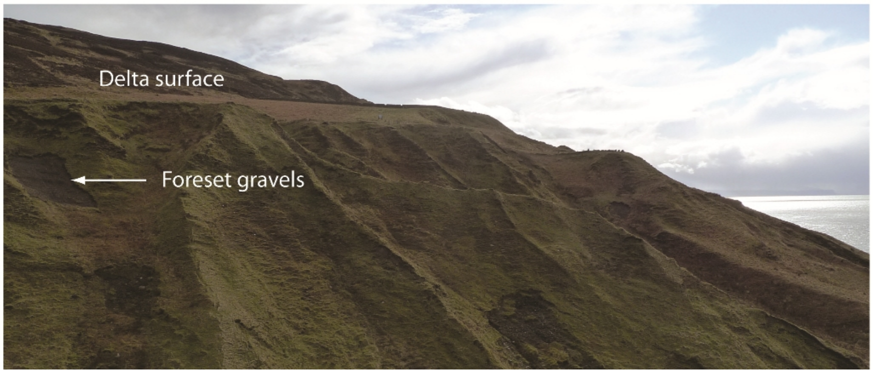

Figure 5.7: Perched delta at an elevation of 130 m a.s.l. on south-western Kintyre. The delta formed as water ponded against an outlet glacier flowing along the low ground offshore.

and thick undulating till deposits, which have been deeply incised by meltwater channels (Fig. 5.6). This reinterpretation is supported by exposures of stiff, subglacial traction till within these features. Some isolated moraines are, however, present on Kintyre. Subdued mounds with boulders scattered on their surfaces exist in Glen Lussa; they occur in association with westward descending meltwater channels. Three erratic boulders of Arran granite, having been transported at least 20 km across the Kilbrannan Sound, were selected from the Glen Lussa landform assemblage for cosmogenic nuclide analyses, to put a minimum constraint on the time since deglaciation.

Suites of moraines on Arran have been described by previous workers [Gemmell, 1973; Ballantyne, 2007a]. In the north of the island, a number of valleys and corries possess an inner suite of clear, boulder moraines (Fig. 5.8). These were previously interpreted by Gemmell [1973] as evidence for a late stillstand or readvance during the final stages of the Younger Dryas, and subsequently reinterpreted by Ballantyne [2007a] as the maximum limits of glacier advance during the Younger Dryas, based on the mutually exclusive relationship with Lateglacial periglacial features. Both workers also recognised sets of more subdued outer moraines close to the coast at the valley mouths. Gemmell [1973] suggested that these outer moraines represented three separate stages during deglaciation (the innermost of the three he attributed to the Younger Dryas), while Ballantyne [2007a] concluded that they pointed towards a pre-Younger Dryas (re)advance

Figure 5.8: Clear boulder moraine at the head of north Glen Sannox, Arran. This moraine probably formed during a Younger Dryas glacier advance.

A series of exposures reveal the stratigraphy in the vicinity of a set of 'outer' moraines at Dougarie, between 0.1 km and 0.7 km up the valley from where a prominent delta surface exists at 30-32 m a.s.l. (Fig. 5.9A). At the time of field investigation, four lithofacies were recognised in sections.

LFA 1 consists of stiff, thinly laminated, very pale brown, grey and white silts and clays, which show varying degrees of folding and attenuation (Figs. 5.9B,C). In places, the laminations are clearly graded. These silts and clays contain rare, isolated, subangular gravel- and cobble-sized clasts. Sedimentary structures around the clasts include wrapped foliation and asymmetrical inclined folds indicative of an east to west sense of shear. Locally, the silts and clays are cut by sand-filled hydrofractures, which appear to have exploited detachments within the silts and clays. Small rafts of attenuated and folded silts and clays are contained within the sand. The base of LFA1 was not exposed. The upper contact with LFA2 is erosional (Fig. 5.9D). LFA 1 is interpreted as a glacitectonite. It represents a period of proglacial deposition in a subaqeous environment, followed by phases of deformation associated with a local glacier advance from the east.

LFA 2 varies in thickness between 0 and 1.5 m. It comprises a dense, grey to pale brown, generally massive to locally stratified, matrix-supported diamicton, containing subangular clasts. The clasts are faceted and consist predominantly of granite (erratics) and metasedimentary lithologies. No primary bedding was observed in LFA 2. The upper contact with LFA 3 is gradational. LFA 2 is interpreted as a subglacial till, deposited by the overriding glacier

LFA 3 comprises a variably loose to dense, poorly sorted, clast-supported bouldery diamicton with coarse sandy matrix and infrequent lenses of sorted, bedded sands (Fig. 5.9E). LFA 3 is dominated by granite erratics, which are sourced from farther up the valley, and rare metasedimentary clasts. This lithofacies forms the topographic expression of the set of moraines, which vary in elevation from 25 - 40 m a.s.l. in the valley centre. These moraines were deposited during local glacier retreat, following its advance.

LFA 4 is sporadically present between moraines, and consists of loose, westward dipping, upward coarsening, stratified sands and gravels, which form delta foresets (Fig. 5.9F). LFA 4 probably represents deposition into ponds formed in proglacial depressions, during glacier retreat.

Collectively, these sediments support the views of both Gemmell [1973] and Ballantyne [2007a], that glacier oscillations took place at the lower end of some valleys in Arran, during overall deglaciation. Many of the moraine (LFA 3) surfaces are lower than the surface elevation of the delta farther down the valley (Fig. 5.9A). Therefore their deposition during overall retreat is likely to have occurred after sea level had fallen from the highpoint marked by the delta surface at 32 m a.s.l. No clear surface boulders exist on the moraines where the sections were exposed. However, two boulders from low lateral moraine fragments, approximately 500 m farther up the valley, were sampled for cosmogenic nuclide analyses in an attempt to constrain the timing of moraine deposition.

5.5 Chronology results

Exposure ages for the sampled boulders in Glen Dougarie, Arran and Glen Lussa, Kintyre are shown in Table 5.1. The samples from Glen Dougarie on Arran yielded overlapping exposure ages with a mean of 16.23 ± 0.969 ka. The Dougarie ages predate, and are therefore consistent with, dated sediment accumulation in the Firth of Clyde [Peacock et al., 2012]. They are only slightly older than the 16 ka ice margin isochrone, which was placed just 20 km to the south by Clark et al. [2012], lending support to the framework ice sheet retreat chronology proposed by these authors. These ages also support previous suggestions by Gemmell [1973] and Ballantyne [2007a] that these lowermost moraines on Arran pre-date the Younger Dryas.

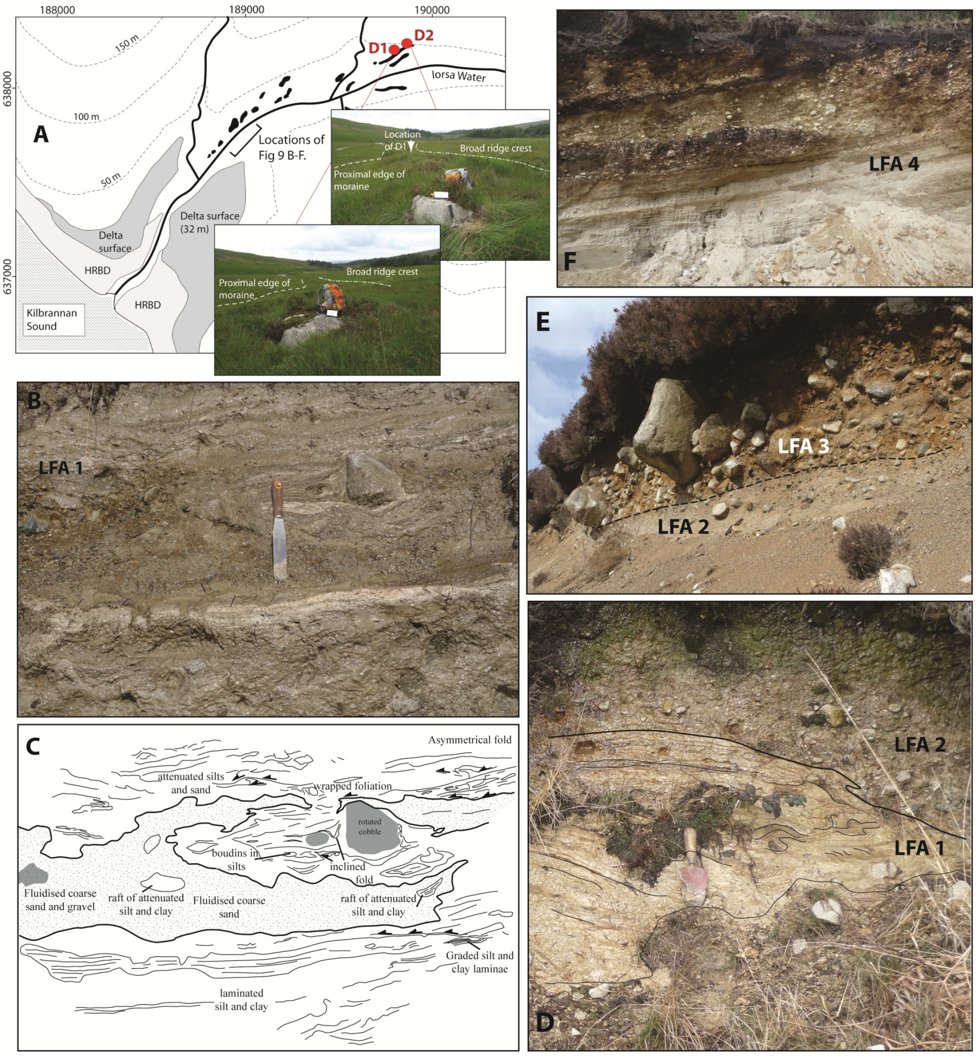

FIGURE 5.9: Sediment exposures at the mouth of Glen Dougarie, Arran. A: Geomorphological context. Filled black polygons indicate the position of moraines. The locations of samples D1 and D2 are shown. HRBD: Holocene raised beach deposits. B: Photograph of lithofacies association 1 (glacitectonite). C: Line drawing highlighting deformation structures in lithofacies association 1. D. Section revealing the contact between lithofacies association 1 and lithofacies association 2 (subglacial till). E: Lithofacies association 3 (moraine). F lithofacies association 4 (delta foresets).

Given their sampling context (discussed above), the Glen Lussa ages represent only a minimum period of time since deglaciation. This is confirmed since: (i) GL1 and GL2 are younger than calibrated radiocarbon ages and fauna assemblages obtained from sediment cores at the southern end of the Kilbranan Sound [Peacock et al., 2012]; (ii) the ages are younger than those from Arran, contrary to the geomorphological evidence for the pattern of north-westward ice retreat (see below); and (iii) the ages are internally inconsistent, with the youngest sample (GL2, 13.0 ± 0.8 ka) and oldest sample (GL3, 15.0 ± 0.9 ka) not sharing overlapping uncertainties. Nonetheless the oldest sample, GL3, together with the Glen Dougarie samples, provide additional independent support to the contention by Peacock et al. [2012] that the outer Firth of Clyde was deglaciated *before* the opening of the Lateglacial Interstadial (Greenland Interstadial-1, 14.7 ka BP).

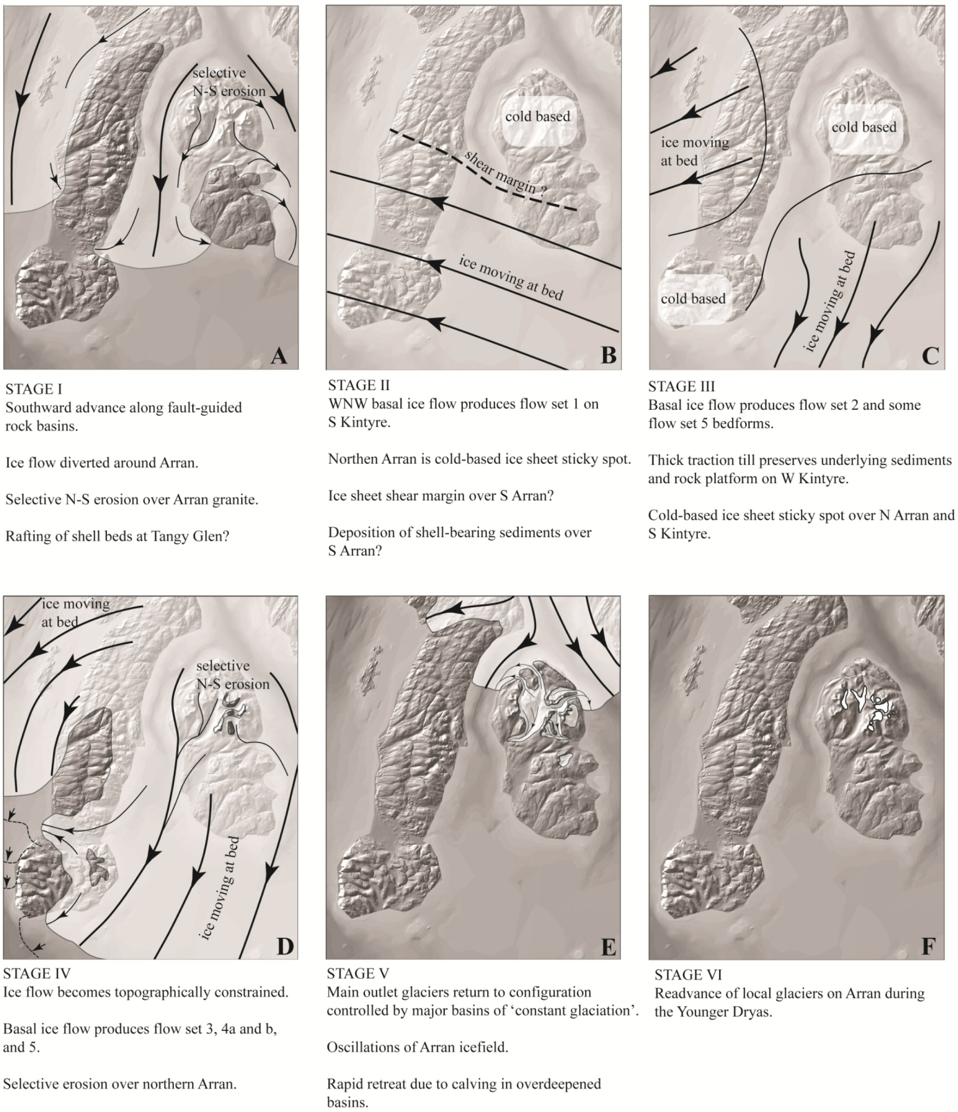

FIGURE 5.10: Interpretation of ice sheet stages that affected the landscape of Kintyre and Arran.

5.6 Ice sheet evolution over Kintyre and Arran

The simplest interpretation of the growth and decay of the last BIIS over Kintyre and Arran, based on the geomorphological evidence reviewed above, is shown in Figure 5.10.

5.6.1 Stage I: Southward ice sheet advance (Fig. 5.10A)

Synge and Stephens [1966] interpreted the shell beds at Tangy Glen as glacial rafts and similar interpretations have been advanced for high-level shell beds and shelly tills elsewhere in Scotland [Merritt, 1992; Peacock and Merritt, 1997b; Phillips and Merritt, 2008]. If a rafting origin is correct, an advancing outlet glacier from the north is the most likely mechanism to have glacitectonically deposited the shelly clays on the eastern Kintyre coastline. A northern sourced advance is supported by the southerly transport of Glen Fyne granite erratics onto Arran [Tyrrell, 1928; Sissons, 1967a], and by the north-south oriented over-deepened basins around Arran and Kintyre (Figs. 5.1,5.3). These geologically controlled, glacially carved fjords are too deep to have been cut during a single glacial cycle [Kessler et al., 2008], and the preservation of pre-MLD rock shore platforms and sediments at the margin of the Sound of Jura are illustrative of an area where bedrock erosion during the last glacial cycle was limited. The over-deepened basins may therefore be considered products of 'average glacial conditions' through the Quaternary [Porter, 1989; Clapperton, 1997; Golledge et al., 2009]. They determined the flow of the advancing, mostly land-based, MLD ice sheet before it expanded onto the Malin Shelf – a configuration that is replicated in numerical simulations of ice sheet flow during the build up phase [Hubbard et al., 2009].

5.6.2 Stage II: non-topographically constrained west-north-westward ice flow onto the Malin Shelf (Fig. 5.10B)

Bedforms belonging to flow set 1 were formed under west-north-westward directed ice movement. At that time ice flow was no longer topographically confined and warm-based ice movement occurred over southern Kintyre at all elevations (Table 5.2). West-north-westerly flow to the south of Arran, and across southern Kintyre is also supported by dispersal patterns of erratics from Ailsa Craig and Loch Doon, SW Scotland [Sissons, 1967a]. The pattern of ice flow could have transported shelly deposits from offshore to onshore over southern Arran [Watson, 1864]. An ice sheet shear zone is inferred across southern Arran and central Kintyre separating southern warm-based ice that flowed towards the Malin Shelf, from northern cold-based, internally deforming ice. The cold-based ice to the north is suggested by: (i) the absence of west-north-westerly aligned bedforms over northern Arran and northern Kintyre; (ii) the preservation of delicate tors on some summits of northern Arran; and (iii) the absence of west-north-westward transported erratics of Arran granite on northern Kintyre [Horne et al., 1896; Eyles et al., 1949].

5.6.3 Stage III: non-topographically constrained south-westward ice flow into the North Channel and flow divergence over southern Kintyre (Fig. 5.10C)

Flow set 2 bedforms and some of the flow set 5 bedforms developed under warm-based ice moving towards the west-south-west and south-south-west, into the North Channel. West-south-westward ice motion occurred easily over the smooth terrain of western central Kintyre, where bedforms developed in the thick traction till that must have protected the underlying pockets of shelly clays, beach gravels, and the rock platform. South-south-westward ice motion occurred over southern Arran, where bedforms are preserved on the present land surface. The absence of streamlined bedforms and the preservation of tors on northern Arran (Fig. 4A) suggests that it remained largely overlain by cold-based ice at that time. However, some warm-based ice flow through the north-south oriented glacial breaches, which possess ice-moulded rock surfaces, could have fed the south-south-westward directed ice movement. The high ground of southern Kintyre, where no south-westward oriented bedforms exist, may have been cold-based at that time.

5.6.4 Stage IV: progressively topographically constrained south-westward ice flow and glacier retreat (Fig. 5.10D)

Bedforms belonging to flow sets 3, 4a and 4b, and 5 were forming under warm-based ice as glacier flow became topographically confined in the outer Firth of Clyde and Sound of Jura during deglaciation. The high ground of southern Kintyre deglaciated first, as indicated by the presence of the perched delta which fed into a lake that was ponded against a grounded glacier offshore (Fig. 5.7). On western Kintyre, ice marginal / sub marginal meltwater subsequently cut deep channels across thick deposits of traction till, dissecting bedforms belonging to flow set 2 (Fig. 5.6). During this phase of events ice flow in the outer Firth of Clyde was directed through the fault-controlled gap between Campbeltown and Machrihanish Bay, demonstrated by the convergent pattern of flow set 4a, which generally occurs at a lower elevation than, and is superimposed on, flow set 1. Southward-flowing ice in the Firth of Clyde was diverted around the high ground of northern Arran, although some basal ice motion may have taken place through the southward oriented valleys and glacial breaches transporting sub-rounded granite boulders to the south and south-west. The spreads of sand and gravel offshore around Kintyre (Fig. 5.3) probably accumulated as ice-proximal subaqeous fans during this overall phase of events.

5.6.5 Stage V: fjord glacier retreat and oscillations of Arran icefield (Fig. 5.10E)

The distribution and orientation of ice marginal meltwater channels show that the major pathways of glacier retreat were along corridors of low lying ground, and principally through the over-deepened, fault controlled, glacially carved basins of the Kilbrannan Sound and North East Arran Trough. The pattern of deglaciation suggested here supports that deduced earlier by Gemmell [1973]. Rapid glacier retreat in the main basins would have been aided by calving as the ice margins thinned and pulled back into deeper water. The sediments and geomorphology at Dougarie, on western Arran, indicate that an advance of a locally sourced glacier took place following separation from the main outlet glacier in the Kilbrannan Sound. Retreat from this local advance took place $\sim$ 16.2 ka, and post-dated a fall in relative sea level from the high-stand that produced the main delta at 32 m (Fig. 5.9A) and other high lateglacial shorelines that are only present on the southern half of the island [Gemmell, 1973]. This timing supports relative sea level simulations for the area, where a falling relative sea level is modelled between $\sim$16.5 and $\sim$15 ka BP [Shennan et al., 2006]. Glaciers are inferred to have advanced or oscillated at similar positions in other valleys on Arran at that time [Gemmell, 1973; Ballantyne, 2007a]. This may reflect internal adjustments of the Arran ice field as it responded to either: (i) the retreat of larger confining glaciers in the surrounding Kilbrannan Sound and North-east Arran Trough, or (ii) enhanced snowfall over the high ground of northern Arran, following the deglaciation of offshore areas farther to the west. The overall configuration proposed at this stage is very similar to that envisaged by Gemmell [1973]. The general timing proposed here is broadly similar to the timing of retreat proposed by Clark et al. [2012], and supports simulations of large marine-based ice losses in the North Channel region and outer Firth of Clyde between 17 ka and 16 ka BP [Hubbard et al., 2009].

5.6.6 Stage VI: Advance of Arran glaciers during the Younger Dryas (Fig. 5.10F)

The suites of clearly defined, sharp-crested moraines that exist in the upper reaches of the valleys of northern Arran (Figs. 5.3, 5.8) point towards an episode of alpine glaciation when small corrie glaciers grew. These moraines have been recognised by several previous authors [e.g. Tyrrell, 1928; Gemmell, 1973; Ballantyne, 2007a]. Detailed mapping of the moraine limits and their mutually exclusive relationship with periglacial features led Ballantyne [2007a] to conclude that this last phase of glaciation took place during the Younger Dryas (12.9-11.5 ka BP). This view is supported by the observations made during this study.

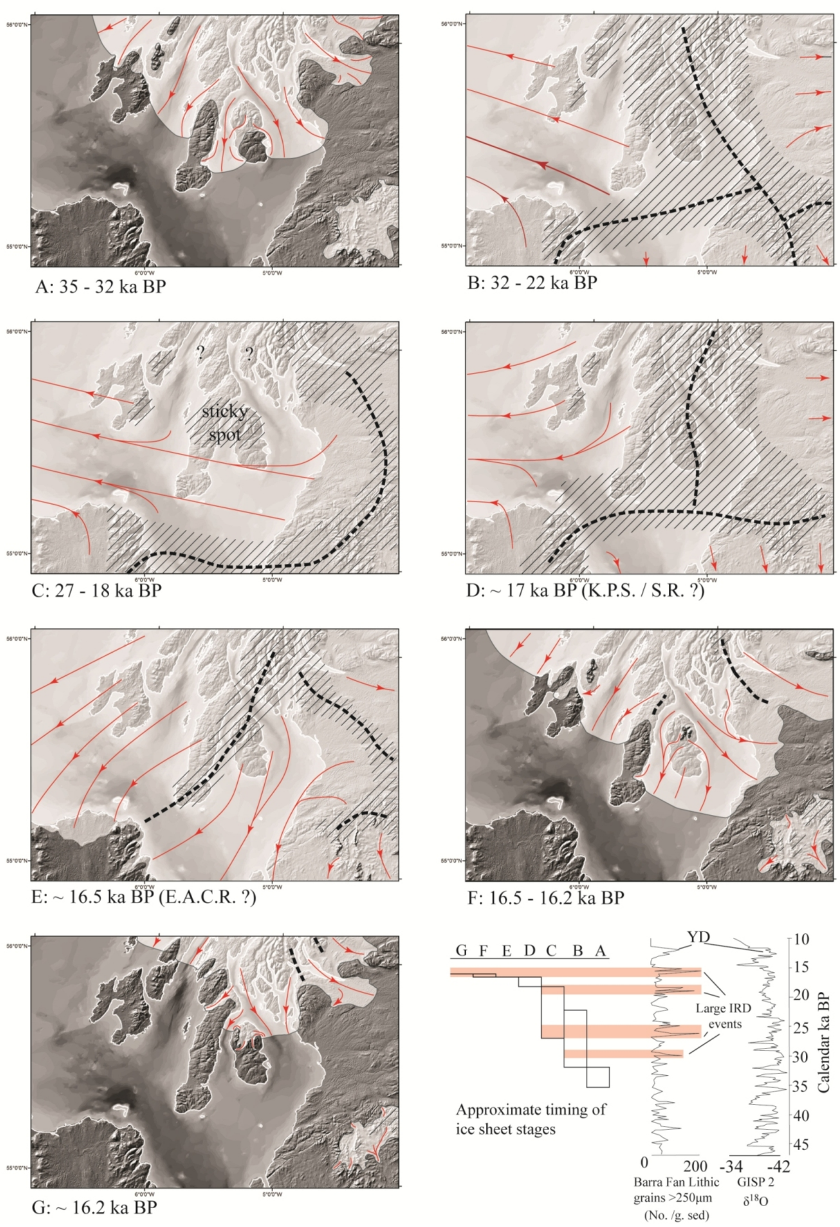

FIGURE 5.11: Growth and decay of the last BIIS over western Scotland, the North Channel, and north-east Ireland. This reconstruction is synthesised from work presented here and existing published research [Salt and Evans, 2004; Dunlop et al., 2010; Finlayson et al., 2010; Livingstone et al., 2012a; Clark et al., 2012; McCabe and Williams, 2012].Diagonal shading indicates probable cold-based ice. Dashed line denotes suggested ice divides. K.P.S.: Killard Point Stadial; S.R: Scottish Readvance; E.A.C.R.: East Antrim Coastal Readvance. Lower right: Lithic grains observed in core MD95-2006 (Barra Fan) and GISP 2 Oxygen isotope record, from Knutz et al. [2001]

5.7 Regional ice sheet evolution

Combining our reconstructed sequence of events with recently published interpretations from south-west Scotland [Salt and Evans, 2004], west-central Scotland [Finlayson et al., 2010, this thesis' Chapter 4], northern England [Livingstone et al., 2012a], north-east Ireland [Greenwood and Clark, 2009; McCabe and Williams, 2012], and the Malin Shelf [Dunlop et al., 2010] allows us to attempt to synthesise the overall growth and decay of the western zone (Clyde-North Channel-Malin Shelf) of the last BIIS (Figs. 5.11 A-G, 5.12).

Published dates from interstadial deposits that underlie till indicate that ice advance into the Clyde and Ayrshire basins occurred after ∼35 ka BP [Bos et al., 2004; Brown et al., 2007; Jacobi et al., 2009]. Prior to that, a more restricted ice cap, which intermittently terminated at the marine limit, existed over the western Scottish Highlands from ∼45 ka BP [Knutz et al., 2001; Scourse et al., 2009]. The advancing outlet lobes of the ice cap encountered reverse slopes in the Clyde and Ayrshire basins, and in the north-east Arran Trough, the Kilbrannan Sound, and Sound of Jura (Figs 5.10A, 5.11A). These topographic settings, combined with the presence of water at the ice margins provided favourable conditions for glacitectonic deformation [Aber et al., 1989], and glacitectonic structures have been recognised in sediments in the Clyde basin [McMillan and Browne, 1983; Browne and McMillan, 1989b].

The Western Highlands ice cap joined with a smaller ice cap centred over the Southern Uplands, prior to a major expansion of the BIIS, which occurred after 29 ka BP [Scourse et al., 2009]. This phase was marked by the western advance (average rate of ∼ 30 m a^{-1}) of marine-based ice sheet sectors over the Malin Shelf [Dunlop et al., 2010], and elsewhere on the western British-Irish continental shelf [Clark et al., 2012; Ó Cofaigh et al., 2012; Everest et al., 2013; Howe et al., 2012]. An ice divide had developed over Arran, most of Kintyre, and the adjacent marine areas at that time, acting as a link to the ice dome over the western Highlands (Fig. 5.11B). Eastward ice flow occurred over west central Scotland [Finlayson et al., 2010], and through topographic corridors in northern England [Livingstone et al., 2012a]. Slow moving ice in the vicinity of the ice divide did not significantly modify the landscape of Kintyre and Arran. An ice ridge had also developed over the North Channel, bridging the British and Irish ice centres [Greenwood and Clark, 2009].

The ice divide that was positioned over Arran and Kintyre migrated ∼ 60 km to the east during a phase, or phases, of enhanced drawdown to the western marine margins of the ice sheet, drained by the large Barra-Donegal Fan / Hebrides Ice Stream

[Dunlop et al., 2010; Ó Cofaigh et al., 2012; Howe et al., 2012] (Figs. 5.10B, 5.11C). This was associated with the development of west-north-west oriented streamlined bedforms at all elevations over southern Kintyre, and possibly also transport of shelly till onto southern Arran (Fig 5.10, stage II). Ice flowing over southern Arran and Kintyre merged with powerful north-westerly flowing ice which overwhelmed the topography of Islay [Cousins, 2012]. However, delicate landforms on northern Arran were preserved beneath a cold-based ice sheet sticky spot, which existed within an overall area of accelerating ice flow. Recent analysis of geochronological data from the Irish Sea Basin show that the retreat of the Irish Sea Ice Stream slowed between ~23 and ~20 ka BP, as the margin entered the constriction between Ireland and Wales [Chiverrell et al., 2013]. Slowing of the Irish Sea Ice Stream, combined with drawdown to the Barra-Donegal Fan / Hebrides Ice Stream could have driven the North Channel ice divide to the south-east over the northern Irish Sea. Such a migration is captured in both the geomorphological reconstruction by Greenwood and Clark [2009] and numerical simulations by Hubbard et al. [2009]. Peaks in IRD concentrations observed in core MD95-2006, from the Barra Fan, suggest that distinct pulses of iceberg discharge took place from ~ 27 ka BP (Fig. 5.11). These pulses may relate to earlier ice stream drawdown and iceberg discharge events, possibly documenting interplay of the Barra-Donegal Fan / Hebrides Ice Stream and the Irish Sea Ice Stream as the BIIS altered between configurations approximating those presented in Figures 5.11B and 5.11C.

A significant iceberg discharge event at the Barra Fan, which may have been associated with large ice losses over the Malin Shelf, ceased ~ 18.5 ka BP (Fig. 5.11) [Knutz et al., 2001]. Following this, the BIIS is suggested to have thickened again over north-east Ireland, advancing at its margins during the Killard Point Stadial, at or soon after 17 ka BP [McCabe et al., 1998; McCabe, 2008](Fig. 5.11D). Livingstone et al. [2012a] summarised the evidence for a readvance of Scottish-sourced ice into northern England at a similar time, although they note that chronological constraints are insufficient to conclusively link the two events. The ice sheet may also have thickened over Arran, most of Kintyre, and the North Channel at this stage, under which little landscape modification took place (Fig. 5.11D), although some south-westward ice flow may have begun to occur over westernmost parts of Kintyre.

McCabe and Williams [2012] provided strong evidence for a later advance of Scottish-sourced ice onto the East Antrim coast of Northern Ireland (the East Antrim Coastal Readvance). We suggest that the East Antrim Coastal Readvance was caused by the delayed response of Scottish-sourced ice to warming at the end of the Killard Point Stadial (17-16.5 ka BP)(Fig. 5.12). The Irish Ice Sheet is reconstructed to have been only ~500 m thick during the Killard Point Stadial, and therefore extremely sensitive to any rise in equilibrium line altitude [Clark et al., 2009b], while the Scottish sector was larger and thicker, with its core positioned over the western Scottish Highlands (Fig. 5.12A). In addition, initial ice sheet break up over the Malin Shelf and the opening of a marine embayment may have allowed more precipitation to reach Scottish source areas, as suggested by McCabe and Williams [2012]. As a result, rapid wasting of the Irish Ice Sheet meant that it no longer obstructed Scottish-sourced ice. The North Channel ice divide collapsed and the Scottish Ice Sheet margin was allowed to temporarily advance over the East Antrim coast, before it too rapidly retreated (Figs. 5.10C, 5.11E, 5.12B), reaching the inner Firth of Clyde in $\sim$ 500 years or less – requiring retreat rates in the order of 10^2 ma^{-1} (Figs. 5.10D, E and 5.11F, G). These retreat rates are likely to have been influenced by reverse slopes in the subglacial topography – a basal condition also observed under parts of the modern-day WAIS [e.g. Ross et al., 2012]. Minor ice front readvances or stillstands occurred during that time, possibly as local outlet glaciers responded to the retreat of larger confining ice masses, or as the wasting ice sheet allowed precipitation to be focused elsewhere. We suggest that this overall phase of rapid thinning and retreat of the Scottish Ice Sheet (south-west sector) may be associated with a peak in iceberg calving, identified in the Barra Fan IRD record at $\sim$16 ka BP [Knutz et al., 2001] (Fig. 5.11).

Our scenario differs somewhat to the proposal by McCabe and Williams [2012] that the East Antrim Coastal Readvance was part of a larger 'North Channel Readvance' approximately 15-15.5 ka BP, with contemporary ice margins across the East Antrim Plateau ($\sim$ 300 m), at the Kilmarnock moraine (100 m a.s.l.) in the Ayshire basin [Finlayson et al., 2010] and Blantyreferme moraine (50 m a.s.l.) in the Clyde basin [Browne and McMillan, 1989b] (Fig. 5.1). We find it difficult to support the overall configuration and timing of the 'North Channel Readvance', proposed by McCabe and Williams [2012] for two reasons. First, linking the East Antrim Coastal Readvance with glacier limits in the Ayrshire and Clyde basins would require ice surface slopes along eastward flow lines to be $\sim$5 times steeper that those flowing onto the north-east Irish coastline. The unusual ice surface topography would necessitate much higher basal shear stresses along eastern flow lines, which is difficult to reconcile with the soft sediment (marine) bed in the outer Firth of Clyde, and the presence of streamlined eastward directed bedforms (mean elongation ratio: 4.3) in Ayrshire [Finlayson et al., 2010]. Furthermore, the thickness of ice required to over top the Antrim Plateau (300 m a.s.l.) means that it would have been grounded in the North Channel at the time of the advance, ruling out the existence of a very low gradient ice shelf as a potential solution to the reconstruction by McCabe and Williams [2012]. Second, McCabe and Williams' proposed timing of 15-15.5 ka BP is within error of radiocarbon ages from molluscs

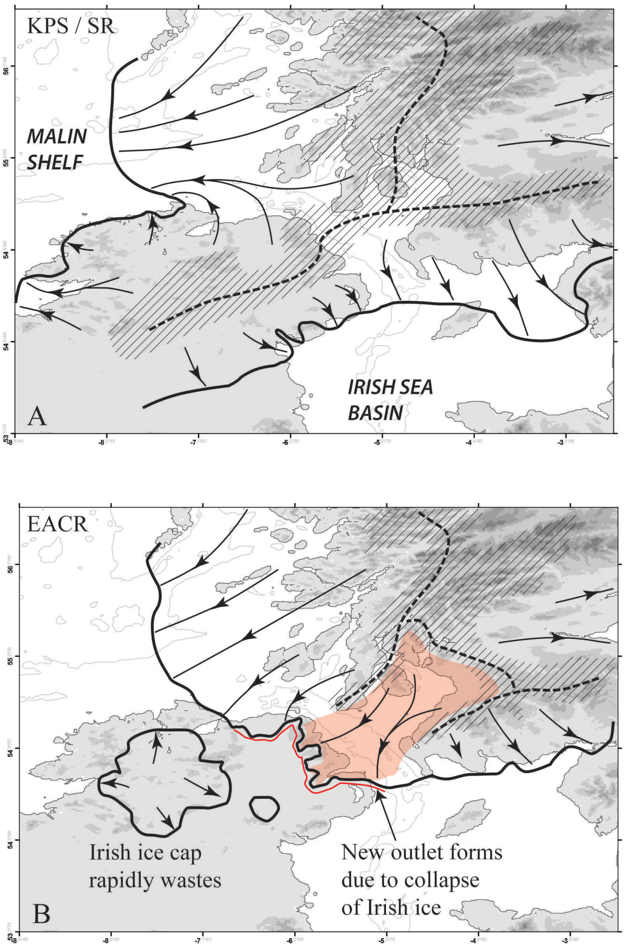

FIGURE 5.12: Interpretation of ice sheet / ice cap configuration prior to and during the East Antrim Coastal Readvance. KPS: Killard Point Stadial; SR: Scottish Readvance; EACR: East Antrim Coastal Readvance. Diagonal shading indicates probable cold-based ice. Dashed line denotes suggested ice divides.

in sediment cores, suggesting that glaciomarine conditions existed around Islay and in the outer Firth of Clyde at that time [Peacock et al., 2012]. The exposure ages from moraines at Dougarie on Arran also suggest that the Kilbrannan Sound and outer Firth of Clyde were ice free by ~16.2 ka BP, and therefore that the East Antrim Coastal Readvance must have occurred slightly earlier than this. The scenario presented here also differs from part of the reconstruction of Finlayson et al. [2010] (their Fig. 17B), who considered ice marginal oscillations in East Antrim and the outer Firth of Clyde (though not necessarily contemporaneous) to be of the same overall phase of events at the GS-2 to GI-1 transition. These events were probably earlier, with the ice sheet having retreated from much of the outer Firth of Clyde by ~16 ka BP, supporting the more recent reconstruction of Clark et al. [2012].

5.8 Ice sheet evolution and the glacial landscape

Our results and reconstruction based on the geomorphological record concur with the prevailing view of a dynamic former BIIS [e.g. Bradwell et al., 2008b; Greenwood and Clark, 2009; Livingstone et al., 2012a]. The ice sheet expanded from a mountain ice cap with tidewater margins to the continental shelf edge in ~7 ka or less. The addition of the marine sector to the ice sheet was accompanied by a marked change in ice-flow directions in the vicinity of Arran and Kintyre. Initially, ice flow had been directed through the geologically influenced north-south oriented fjord basins. These over-deepened glacial troughs probably represent a position that was often reached by restricted, marine-proximal mountain ice sheets during the middle and late Quaternary. Ice flow along these corridors was then abandoned once the extensive Malin Shelf sector became established, with powerful ice sheet drawdown towards the continental shelf forcing ice to flow at right angles to the initial flow direction.

The marine terminating phase of ice sheet glaciation was strongly influenced by episodes of ice divide migration, possibly linked to ice streaming and large calving events. Importantly, however, *stable* ice sheet configurations were also a feature of the marine-influenced phase. For example, while the main west-east ice divide migrated by up to 60 km over low relief areas in the outer Firth of Clyde and Clyde and Ayrshire basins, it remained a relatively stable, stationary feature over the western Scottish Highlands. Similarly, the zone of cold based ice (ice sheet sticky spot?) over northern Arran was probably a permanent and stationary feature through the whole marine phase of the ice sheet cycle. These stable features in the BIIS provide some support to recent suggestions of long term stability (over 10^4 years or more), influenced by subglacial topography, for parts of the West Antarctic Ice Sheet [Ross et al., 2011].

The North Channel ice divide linked an ice ridge over the Southern Uplands in Scotland with the higher ground of north-east Ireland. Although it migrated over time due to the interplay between the Barra-Donegal Fan / Hebrides Sea Ice Stream and the Irish Sea Ice Stream, it remained a constant feature of the marine BIIS until the Irish Ice Sheet rapidly decayed on land, after 17 ka BP (Fig 5.12). Collapse of the North Channel ice divide allowed the Scottish Ice Sheet to temporarily advance over north-east Ireland, before it too retreated back into the coastal fjords, at rates in the order of 10^2 m a^{-1}, and readopted the restricted north-south, fjord-aligned ice flow pattern. This represents a relatively rapid phase of ice sheet decay, exceeding the overall average retreat rate from the shelf edge, which was in the order of 10^1 m a^{-1}, similar to the rates identified by Clark et al. [2012].

The landscape of Kintyre and Arran lay under both a small land-based ice sheet with tidewater margins and a larger ice sheet with significant marine sectors. These different ice sheet configurations and the variability in conditions at the ice sheet bed are highlighted by the composite landscape that is now preserved; it includes: (i) tors of probable middle Quaternary age; (ii) breaches and rock basins that are hundreds of metres in depth; (iii) an (interglacial?) rock shore platform, which was cut prior to the last glacial cycle; (iv) preserved pre-Late Devensian marine sediments, which may have been emplaced by glacitectonic rafting at the start of the last glacial cycle; (v) streamlined bedrock and soft sediment bedforms that were developed during the maximum phases, and subsequent retreat phases of the last BIIS; and (vi) ice marginal assemblages formed during a readvance of alpine-style glaciers during the Younger Dryas.

The first order components of the glacial landscape are the deep, geologically controlled, north-south aligned rock basins, used by Clayton [1974] in his 'relatively high glacial erosion' (Zone III) classification of the landscape. We have demonstrated that these features do not relate to the most recent period of extensive marine-terminating ice sheet glaciation. The scales (10^2 m vertical, and 10^3-10^4 m horizontal) of the rock basins indicate that they have been cut over repeated glacial cycles [Kessler et al., 2008]. The rock basins extend ~50-100 km from lines of maximum glacial erosion modelled in the Scottish Younger Dryas ice cap by Golledge et al. [2009], suggesting western Scotland has often supported a mountain ice sheet with tidewater margins, slightly larger than the Younger Dryas ice configuration. This 'restricted, mountain ice sheet with tidewater outlets' configuration is suggested to have been the dominant glacial mode in Britain for large parts of the Quaternary, and particularly prior to 1.1 Ma BP [Lee et al., 2012]. Similar patterns in the Quaternary glacial landscape have been recognised in Fennoscandia, where parts of the landscape were shaped exclusively during restricted mountain ice sheet phases, which dominated the early and middle Quaternary [Fredin, 2013; Kleman et al., 2008]. These findings have implications for studies on present ice sheets, where modern geophysical techniques are now being used to map the glacial landscape under the ice [e.g. Smith et al., 2007; King et al., 2009]. At the margins of the Ellsworth Subglacial Highlands, for example, erosional basins at 10^2-10^3 vertical and 10^4 horizontal scales have been suggested to have formed under an early marine-proximal, mountain ice sheet, and do not relate to flow of the present marine WAIS [Ross et al., 2014]. Similarly, parts of the present-day bed of the East Antarctic Ice Sheet are now recognised to have been sculpted by a succession of ice sheet configurations that were substantially different from today's [Young et al., 2011].

These suggestions are supported by our reconstruction of the BIIS and its relationship with the glacial landscape of western Scotland.

5.9 Conclusions

The following conclusions can be drawn by synthesising the new findings from Arran and Kintyre with published work from the wider area.

- The glacial landscapes of the Kintyre peninsula and the island of Arran preserve a record of both restricted, marine-proximal mountain glaciation and shelf-edge glaciation. The diverse, composite landscape has enabled the evolution of the western marine margin of the last BIIS to be reconstructed.

- Ice advance was initially directed through north-south aligned, geologically-controlled basins that have been carved during successive glacial cycles. These basins record a restricted, marine-proximal mountain ice sheet configuration, slightly larger than the Younger Dryas glacial extent, which probably existed for large parts of the middle and late Quaternary.

- Published dates indicate that ice advanced to the shelf edge after ~35 ka BP, at an average rate of ~ 30 m a^{-1}. The development of a marine sector was marked by a 90° shift in ice flow direction over Arran, Kintyre and the adjacent marine areas. The marine phase of the western BIIS margin saw ice divide migration by up to 60 km, possibly linked to ice streaming and calving events. However, stable ice sheet features also persisted over subglacial topographic highs.

- A significant calving event at the western margin of the BIIS was followed by ice sheet regrowth during the Killard Point Stadial (KPS). The KPS ended ~16.5 ka BP with rapid wasting of the Irish Ice Sheet on land. The North Channel ice divide collapsed as a result, allowing grounded Scottish ice to advance over north-eastern Ireland (the East Antrim Coastal Readvance).

- Subsequent retreat of Scottish ice to the inner fjords was rapid, in the order of 10^2 m a^{-1}. Overall ice retreat was accompanied by oscillations of the Arran ice field, possibly due to removal of confining fjord glaciers, or refocusing of precipitation.

- The ‘restricted’ and ‘extensive’ ice sheets had very different flow regimes over Arran, Kintyre and the surrounding area. First order features in the glacial landscape relate to the former. Similar first order features, relating to restricted glacial conditions, may be identified in geophysical surveys used to map subglacial highland landscapes under interior parts of modern ice sheets.

Part III

Palaeoglaciology, geological modelling and ice sheet beds

Chapter 6

Ice dynamics and sediment movement: last glacial cycle, Clyde basin, Scotland

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Abstract

The nature and behaviour of sediment beneath glaciers influences how they flow and respond to changing environmental conditions. The difficulty of accessing the bed of current glaciers is a key constraint to studying the processes involved. This paper explores an alternative approach in which accessible sediments under the beds of former mid-latitude ice sheets are examined and related to changing ice behaviour during a glacial cycle. The paper focuses on the partly marine-based Pleistocene British Ice Sheet in the Clyde basin. A three-dimensional computation of subsurface glacial sediment distribution is derived from 1260 borehole logs. Sediment distribution is linked to an empirically-based reconstruction of ice sheet evolution, permitting identification of distinctive phases of sedimentation. Maximum sediment mobilisation and till deposition (<0.04 m a^{-1}) occurred during ice advance into the basin from adjacent uplands. Subglacial processes were influenced locally by the relative stiffness of pre-existing sediments, the permeability of the sub-till lithology, and topography; the resulting mean till thickness is 7.7 m with a high standard deviation of 7.0 m. In places, focused till deposition sealed pre-existing permeable substrates, promoting lower effective pressures. Sediment remobilisation by meltwater was a key process as ice margins retreated back through the basin, upon deglaciation.

6.1 Introduction

Patterns of sediment movement beneath soft-bedded glaciers are poorly understood, despite proposed links to glacier dynamics [Alley, 1991; Boulton, 1996; Alley et al., 1997; Clarke, 2005]. The original distribution and subsequent redistribution of sediment not only have implications for glacier motion, but also affect the landscape that glaciers override [Nolan et al., 1995; Motyka et al., 2006], producing feedbacks that can influence glacier flow. Despite the success of some recent investigations on present glaciers [Smith et al., 2007], quantifying the volume and style of subglacial sediment mobilisation remains a challenge, largely due to inaccessibility of the bed.

The geological record in formerly glaciated landscapes offers an alternative approach. Here landforms and sediments contain a cumulative signature of recent ice sheet cycles, which enable reconstructions of palaeo-ice sheet evolution to be made [Boulton and Clark, 1990; Kleman et al., 1997; Stokes et al., 2009]. Where formerly glaciated areas have densely-spaced borehole datasets, these can be used to compute a three-dimensional geological model of glacigenic deposits, allowing their distribution and volumes to be calculated. Combining these geological models with time-transgressive ice sheet reconstructions makes it possible to reconstruct patterns of sediment mobilisation during a glacial cycle – a key aid in understanding glacial sediment transport and deposition mechanisms [Alley, 1991; Boulton, 1996; Alley et al., 1997; Piotrowski et al., 2001; Thomason and Iverson, 2009].

In the Clyde basin, western Scotland (Fig. 6.1), a well-preserved landform and sediment record documents the build up and decay of the last, Main Late Devensian, British Ice Sheet (BIS) [Geikie, 1863; Price, 1975; Rose and Smith, 2008]. An extensive database of borehole logs exists for the area [Menzies, 1976, 1981]. Consequently, the basin provides an ideal case study to examine both former ice sheet evolution and sediment mobilisation. Finlayson et al. [2010, this thesis' Chapter 4] presented a synthesis of geomorphological and stratigraphical evidence from west central Scotland (including the western part of the Clyde basin) (Fig. 6.1), to derive a conceptual model of ice sheet advance and decay during the last glacial cycle. The aim of the work reported here is to elucidate the pattern and volume of sediment that was mobilised during the glacial cycle, and examine the implications for the style of sediment transport during different phases of glaciation.

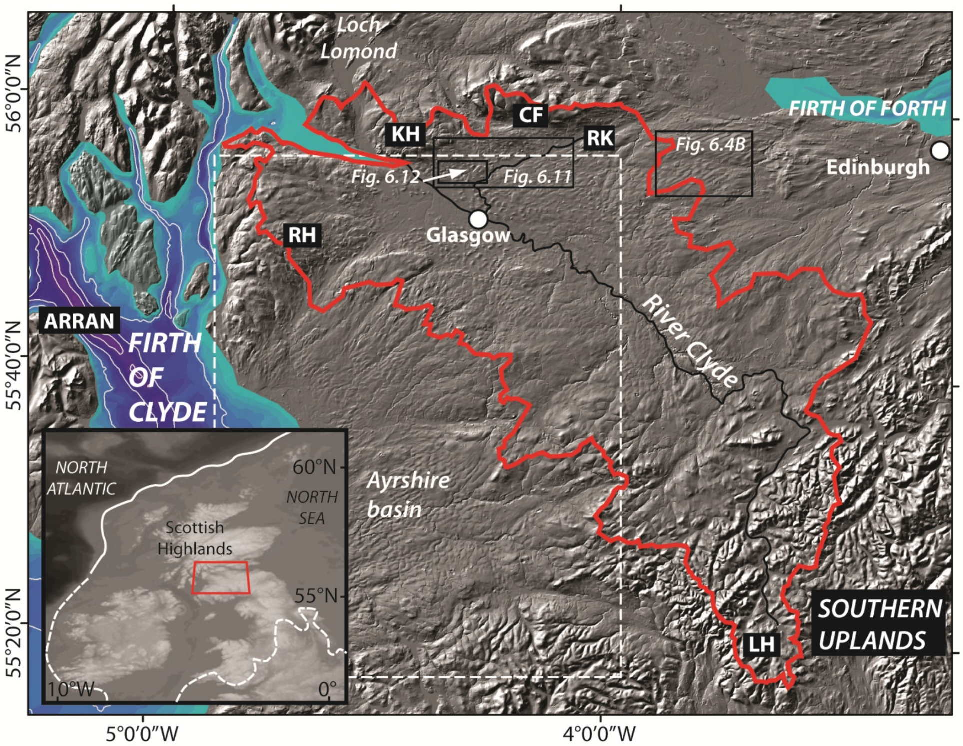

FIGURE 6.1: The boundary of the Clyde basin is shown by the red outline. Dashed white rectangle shows area studied by Finlayson et al. [2010]. Abbreviations: LH Lowther Hills; RH Renfrew Hills; KH Kilpatrick Hills; CF Campsie Fells; RK River Kelvin. Locations of subsequent figures are shown. Hill-shaded digital surface model built from Intermap Technologies NEXTMap Britain topographic data (NW illumination). Offshore bathymetry from BGS DigBath contours. Inset: Location of Clyde basin within a UK context. White line gives approximate extent of last British and Irish Ice Sheet, based on Bradwell et al. [2008b] (solid line) and Clark et al. [2012] (dashed line).

6.2 Physiographic and palaeoglaciological setting

The Clyde basin (3173 km^2) is a predominantly lowland, formerly glaciated landscape (Fig. 6.1). It is drained by the River Clyde, which flows in a north-westerly direction for ~120 km, from its source in the Lowther Hills, through the Glasgow conurbation, to reach the coast at the Inner Firth of Clyde. Much of the eastern and central basin is underlain by cyclically deposited sedimentary rocks (Scottish Coal Measures Group and Clackmannan Group). These are bounded to the north and west by lavas forming the Kilpatrick Hills, Campsie Fells, and Renfrew Hills, and to the south by turbidite sequences and lavas, which underlie the fringes of the Southern Uplands. Most of the ground surface is less than 250 m above sea level (a.s.l.), with the surrounding hills rising to <700 m a.s.l.

The Clyde basin is first known to have been glaciated ∼0.45 Ma BP; however, earlier episodes of glacier ice advance into the basin may also have occurred [Lee et al., 2012]. Since then, a further four continental-scale ice sheet glaciations are thought to have affected the British Isles. The most recent ice sheet overrode the Clyde basin sometime after 35 ka BP [Brown et al., 2007; Jacobi et al., 2009]. Initial glacier advance was from the northwest, sourced from an ice cap over the Scottish Highlands; ice flow was then diverted eastwards following the coalescence of Highland and Southern Upland ice masses [Price, 1975; Sutherland and Gordon, 1993]. During or soon after maximum glaciation, an ice divide migrated eastward over the Clyde basin, at a time of enhanced drawdown towards western marine outlets of the BIS [Eyles and McCabe, 1989; Finlayson et al., 2010]. Final deglaciation of west central Scotland was characterised, once more, by south-eastward and eastward ice flow (a configuration similar to initial ice sheet advance), prior to complete glacier stagnation. The Clyde basin is thought to have become ice free by ∼15 ka BP, at which time relative sea level locally approached 40 m a.s.l.[Peacock, 2003; Rose, 2003].

The regional glacial stratigraphy for the Clyde basin has been well documented [Rose, 1981; Menzies, 1981; Browne and McMillan, 1989b; Finlayson et al., 2010]. For context, key units relevant to this study are summarised in Table 6.1. A buried till (older than 35 ka), the Ballieston Till Formation, has been recorded in parts of northern Glasgow. It is overlain by glaciofluvial gravelly sands derived from a glacier margin positioned near the entrance to the basin (the Cadder Sand Formation), which have yielded woolly rhinoceros bones dated to 35 ka cal BP [Jacobi et al., 2009]. Isolated pockets of glaciolacustrine sediments (the Broomhill Clay Formation) also locally overlie the Ballieston Till Formation. All these deposits predate the last Main Late Devensian glacier advance into the basin and are predominantly found filling a concealed bedrock valley that trends west-southwest to east-northeast under the floodplain of the River Kelvin (Fig. 6.1), or in bedrock hollows beneath the River Clyde. Above rests the regional till of the area, the Wilderness Till Formation. In places, the Wilderness Till Formation is overlain by the Broomhouse Sand and Gravel, Bellshill Clay, and Ross Sand formations; these being glaciofluvial, glaciolacustrine and glaciodeltaic deposits respectively. Raised glaciomarine deposits of the Clyde Clay Formation are found at, or close to, the ground surface across western parts of the basin that lie below 40 m a.s.l.

Stratigraphic unit	Geological model unit	Main lithologies
Clyde Clay Formation	Raised marine deposits	Clay and silt
Ross Sand Formation	Glaciodeltaic deposits	Sand
Bellshill Clay Formation	Glaciolacustrine deposits	Clay
Broomhouse Sand and Gravel Formation	Glaciofluvial deposits	Sand and gravel
Wilderness Till Formation	Till	Diamict
Broomhill Clay Formation	Buried glaciolacustrine deposits	Clay
Cadder Sand Formation	Buried glaciofluvial deposits	Sand and gravel
Ballieston Till Formation	Buried till	Diamict
Pre-Quaternary Bedrock	Pre-Quaternary bedrock	Rock

Table 6.1: Stratigraphy of glacigenic sediments in the Clyde basin

6.3 Methods

To determine the pattern of sediment mobilisation through the last glacial cycle in the Clyde basin, two sets of information were required: (i) an event stratigraphy, based on geomorphological investigation; and (ii) a 3D geological model that reveals the volume and distribution of glacigenic sediments laid down during particular events or stages.

6.3.1 Geomorphological investigation

As noted, Finlayson et al. [2010] have synthesised evidence from west central Scotland, including the western half of the Clyde basin. In this study, the remainder of the Clyde basin was observed to produce a basin-wide geomorphological dataset to complement the 3D geological modelling (described below). The glacial geomorphology was investigated using digital surface models (DSMs) and georectified 1:10 000 monoscopic aerial photographs, interrogated within ESRI Arcmap 9.3. The DSMs, built from NEXTMap Britain topographic data (1.5 m vertical and 5 m horizontal resolution) were illuminated from the north-west and north-east in order to limit bias that can be introduced by relief shading [Smith and Clark, 2005] and enable the capture of landforms with different alignments. The DSMs were viewed at several scales, ranging from 1:10,000 to 1:200,000, and sub-sampled at progressively lower horizontal resolutions, from 5 m to 50 m, in order to capture both small- and large-scale features. Mapped distributions of glaciomarine, glaciodeltaic, glaciolacustrine, and glaciofluvial sediments were based upon a digital 1:50,000-scale geological map of the whole area (DiGMapGB 50).

6.3.2 3-D geological modelling

The dimensions and distribution of subsurface glacigenic sediments in the Clyde basin were established using the geological modelling software: GSI3D (©Insight GmbH) and GOCAD™. In this study, GSI3D [Kessler et al., 2009] was used primarily for development of the model, while GOCAD was used for model interrogation. Input datasets required for the geological modelling process are summarised in Figure 6.2. Using GSI3D, 85 digital cross-sections (total length = 1860 km) were created at regular (2 to 4 km) intervals. These cross-sections were constrained by 1260 borehole logs, the 1:50,000-scale digital superficial geology map, a digital surface model, and a bedrock elevation model. In the Clyde basin, the bedrock elevation model is based on an extensive borehole dataset with 44753 proven bedrock elevation records; it is also 'influenced' by a further 7028 total depth (bedrock not reached) records [Lawley and Garcia-Bajo, 2011]. Collectively, the 85 cross-sections formed a 'fence diagram', which was combined with the digital geological map of the basin to create envelopes, representing the lateral (surface and buried) extent of model units (Table 1). Triangulated Irregular Networks (TINs) were then computed for the surface and base of each unit in GSI3D, by interpolating between regularly-spaced x,y,z nodes on the cross sections and envelopes using an incremental Delaunay triangulation algorithm [Green and Sibson, 1978].

The borehole log and location files, used in the cross-sections, were imported from a nation-wide database, which includes geotechnical site investigation logs, water well records, sand and gravel assessment reports and coalfield investigations. Because the original focus of each data subset was different, the detail concerning the glacigenic deposits is variable. Consequently, during selection of the 1260 boreholes used for the geological modelling, preference was given to those that described the glacigenic deposits in most detail and those that penetrated the full thickness of the glacigenic deposits. Data handling limitations and variable borehole record quality precluded use of the full 50,000+ borehole dataset for the geological modelling. Care was taken to ensure that selected boreholes were from sites evenly distributed throughout the model area. The only 'holes' in the borehole data are where bedrock is present at the land surface. However, since these sites lie outside the envelopes created for model units, they did not affect the calculation.

A requirement of the GSI3D program is that the geological model adheres to an assigned stratigraphy. The stratigraphy used here (Table 6.1) is a simplification of the stratigraphy developed by Browne and McMillan [1989b], and is consistent with that given by Menzies [1981]. Due to the scale of the study area, 'till' was treated as a

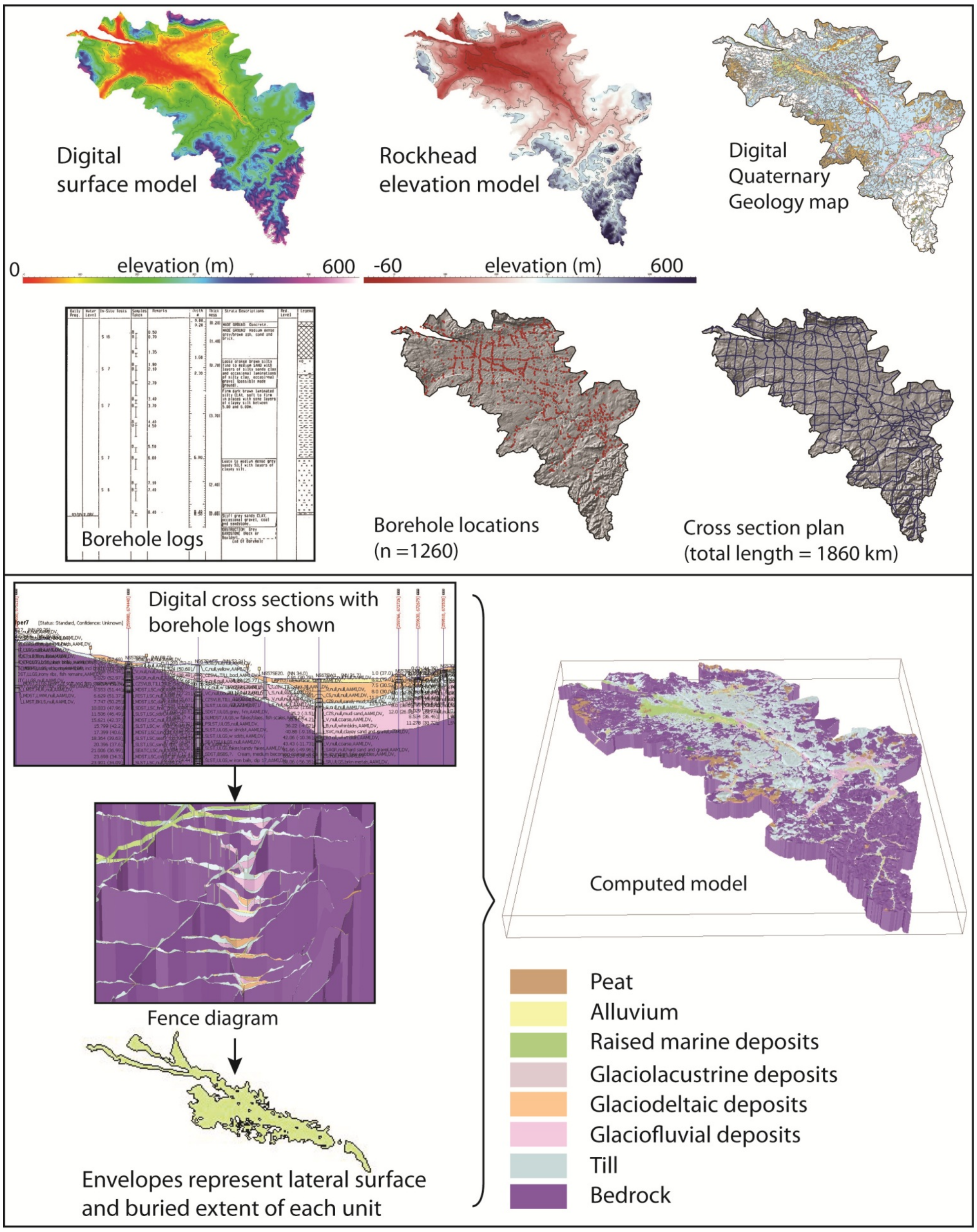

Figure 6.2: Upper box: input datasets required for the 3D geological modelling. Lower box: summary of workflow.

homogeneous unit. The Wilderness Till Formation has been described as a deformation till [Rose et al., 1988], and predominantly comprises a massive matrix supported diamict. However, lenses of sand and gravel, tectonised sorted sediments and till stratification have also been observed [Menzies, 1981; Browne and McMillan, 1989b]. These till heterogeneities were too small-scale to be included in the present basin-wide study, although they are a focus of ongoing work within the Glasgow conurbation using densely-spaced geotechnical data and geostatistical analyses [Kearsey et al., in review].

6.4 Results and discussion

6.4.1 The geomorphology of the Clyde basin

A prerequisite for interpretation of the 3D geological modelling results is an event stratigraphy, based here on geomorphological data. Results of the geomorphological investigation (Fig. 6.3) and the interpreted event stratigraphy are given below. Note that data for the western part of the basin have been published previously [Finlayson et al., 2010])(Fig. 6.1), and that new mapping encompasses the remainder of the basin. However, for clarity the basin as a whole is summarized.

6.4.1.1 Description

The north-western part of the basin is dominated by a landsystem of drumlins and ribbed moraine over the low ground (underlain by sedimentary rocks). There is little sediment cover over the volcanic rocks forming the higher ground, where numerous striations have been recorded [Paterson et al., 1998]. The drumlinised landsystem is overlain by glaciodeltaic sequences in parts of the Kelvin Valley, and by raised glaciomarine sediments in the Glasgow area. The south-eastern part of the Clyde basin is characterised by glaciofluvial deposits (including eskers) and by meltwater channels.

Drumlins in the Clyde basin possess differing alignments (Fig. 6.3), as previously noted by Rose and Letzer [1977] and Rose and Smith [2008]. Individual flow sets [e.g. Boulton and Clark, 1990; Kleman et al., 2006; Stokes et al., 2009], inferred from drumlin alignment are shown for the Clyde basin in Figure 6.4A. The relative age of each flow set is based on landform overprinting (e.g. Fig. 6.4B) [Hughes et al., 2010], and stratigraphical evidence synthesised from sites across west central Scotland [Finlayson et al., 2010]. As expected, drumlin morphological characteristics are consistent with

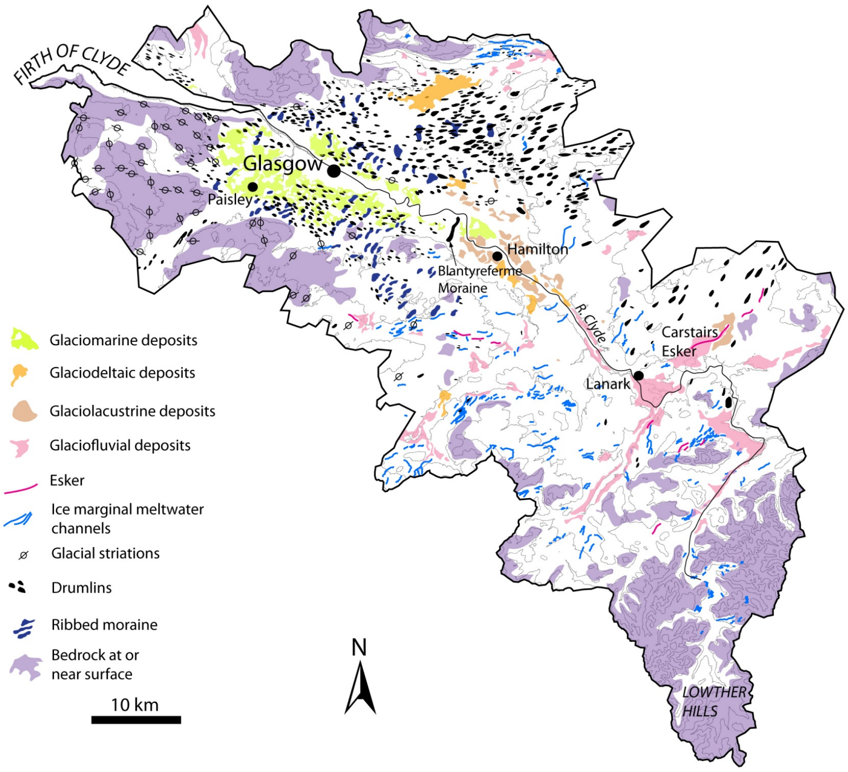

FIGURE 6.3: Geomorphology and glacial geology of the Clyde basin. Glacial striations taken from Paterson et al. [1998].

those observed for British drumlins by Clark et al. [2009a]. The range of drumlin lengths, widths and elongation, and their relationships, are shown as a plot of co-variation in Figure 6.5, following the approach of Clark et al. [2009a]).

In the Clyde basin (and Ayshire basin to the south) ribbed moraines, where present, always appear to be overprinted by drumlins – a pattern also apparent in the subglacial bedform map of Hughes et al. [2010]. Finlayson et al. [2010] suggested that some of these ribbed moraines may have initiated as ice marginal sediment ridges, generated by folding and thrusting of thick sequences of proglacial sediments, during ice sheet advance into the basin from the north-west. There is sedimentological evidence that supports this, where till overlies a low ridge of glaciotectonically thrusted silts, sands and gravels [McMillan and Browne, 1983]. These landforms tend to be present only in the lower Clyde basin, where a number of the factors considered important for genesis of glaciotectonic phenomena [Aber et al., 1989] were likely to have operated (e.g. ice advance against topography, damming of proglacial lakes).

The ice marginal landforms, including meltwater channels, glaciodeltaic deposits, glacio-lacustrine deposits and moraines, reveal stages of ice margin retreat [Clark et al., 2012], shown in Figure 6.6. A zone of ice cap separation is inferred across the Southern part

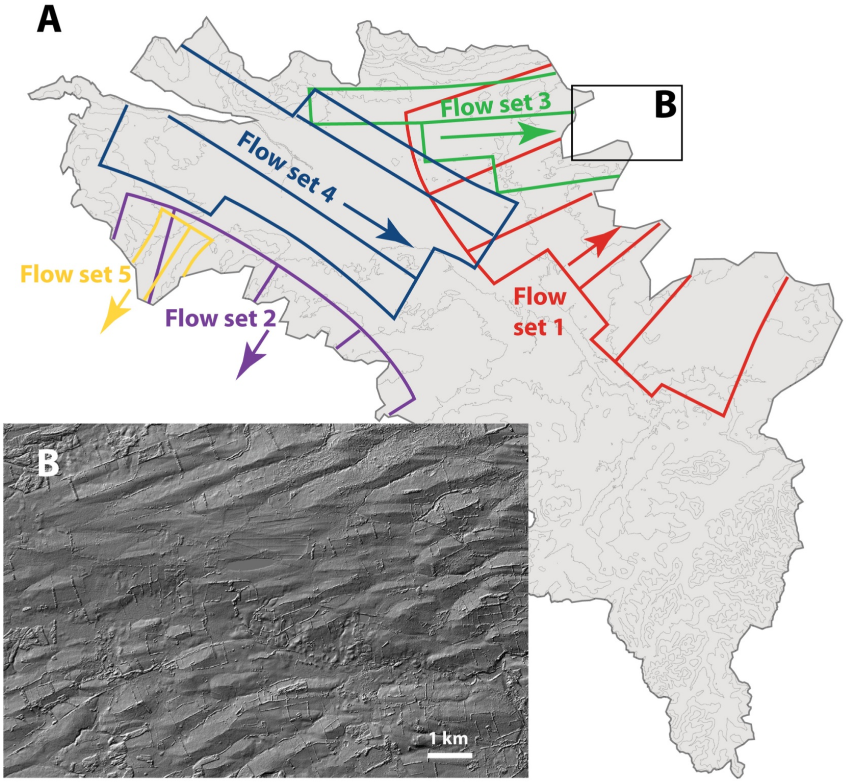

FIGURE 6.4: A: Flow sets inferred from drumlin alignment in the Clyde basin. B: Interference pattern developed immediately to the east of the Clyde basin. Here, north-easterly oriented drumlins of Flow set 1 are overprinted by easterly oriented drumlins of Flow set 3.

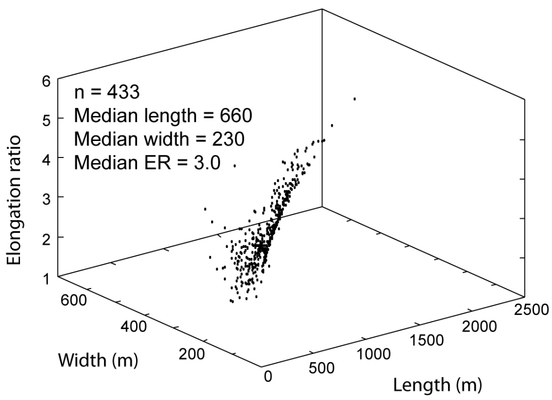

FIGURE 6.5: Co-variation plot of Clyde basin drumlin characteristics. The scale dependent elongation limit, first recognised for drumlins by Clark et al. [2009a] can be seen.

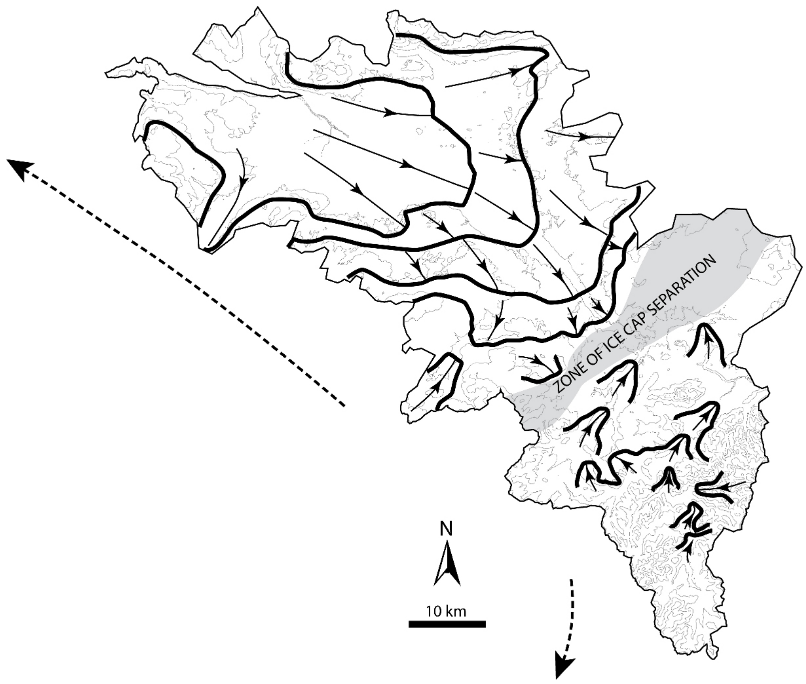

Figure 6.6: Suggested pattern of ice cap separation and subsequent retreat, interpreted from ice marginal landforms (e.g. meltwater channels, ice-dammed lake deposits, moraines). Black arrows denote ice flow direction. Dashed line shows retreat direction.

of the basin in the vicinity of the Carstairs esker (Fig. 6.3). This supports the suggestion of Thomas and Montague [1997] that the esker system developed in an interlobate sediment sink, during uncoupling of northern- and southern-sourced ice caps. Similar persistent subglacial conduits have been observed in modern glacial environments, separating the behaviour of confluent ice masses [Benn et al., 2009]. Aside from minor differences in ice margin detail, the overall pattern of glacier retreat inferred here supports that presented by Clark et al. [2012].

6.4.1.2 Interpretation: an event stratigraphy

Combining the geomorphological evidence with recently published reconstructions for adjacent parts of the last BIS allows an event stratigraphy to be proposed for the Clyde basin (Fig. 6.7). Initial ice sheet advance into the Clyde basin was from the northwest (Fig. 6.7A) sometime after 35 ka BP [Jacobi et al., 2009]. The configuration (Fig. 6.7A) would have permitted the build up of an ice-dammed lake by blocking the River Clyde [Price, 1975]. The buried glaciolacustrine (Broomhill Clay Formation) sediments (Table 6.1) may be remnant deposits from this lake. In the Kelvin Valley, outwash sediments of the Cadder Sand and Gravel Formation were overridden during glacier advance.

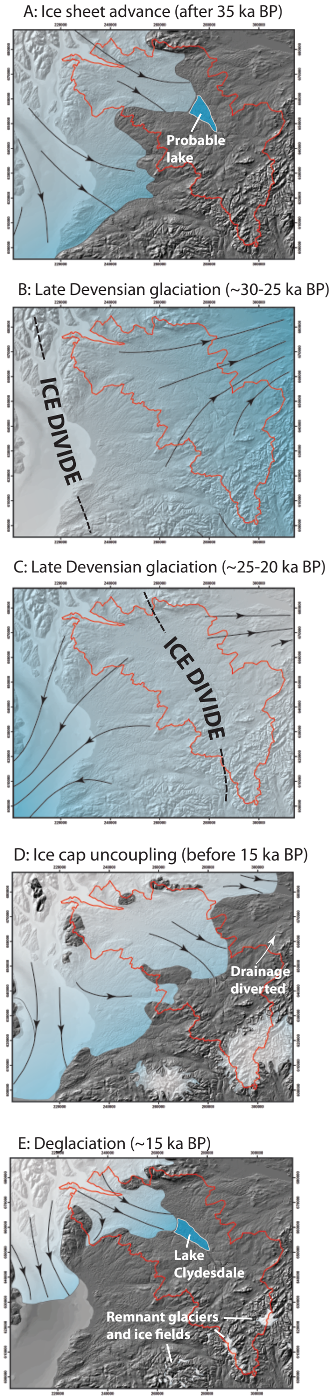

FIGURE 6.7: Reconstructed stages showing the evolution of the last BIIS in the Clyde basin. (a) Advance of outlet glacier into the Clyde basin, accompanied by lake ponding. (b) Development of ice divide to the west of the Clyde basin, accompanied by ice flow to the northeast. (c) Migration of ice divide over the Clyde basin. (d) Ice-divide migration to northwest, ice-sheet decay and separation into ice caps. (e) Final glacier retreat in Clyde basin, accompanied by the ponding of 'Lake Clydesdale'. Hill-shaded digital surface model built from Intermap Technologies NEXTMap Britain topographic data (NW illumination).

An ice dome developed over the Southern Uplands, and coalesced with Highland-sourced ice to form an ice divide to the west of the Clyde basin, forcing flow towards the northeast (Fig. 6.7B). The drumlins of flow set 1 probably initiated at that time. This configuration may have broadly persisted until the BIS approached its maximum extent, around 27 ka BP [Clark et al., 2012]. Enhanced drawdown towards western marine outlets [Eyles and McCabe, 1989; Scourse et al., 2009; Dunlop et al., 2010] then forced the ice divide to migrate over the Clyde basin (Fig. 6.7C). Flow set 2 may have begun to develop during, or following, this phase; it is not possible to establish if flow set 3 also began to form at that time, or later. In general the geomorphological signature of this stage (Fig. 6.7C) is limited in the Clyde basin. However, westerly and south-westerly flow left a strong imprint in the Ayrshire basin to the south-west [Finlayson et al., 2010, 2014, see this thesis' Chapter 4 and Chapter 5]. The presence of an ice divide over the Clyde basin, with associated low or zero velocities, favoured preservation of subglacial bedforms left during the earlier part of the glacial cycle [Clark, 1993].

During ice sheet decay (Fig. 6.7D), ice flow in the lower Clyde basin was from the north-west, documented by flow set 4. A zone of separation developed between the northern-sourced ice and ice caps centred over the Southern Uplands (Fig. 6.6) [Thomas and Montague, 1997]. Glaciofluvial sand and gravel deposits, and glaciolacustrine deposits, were focused along this corridor (Fig. 6.3). The latter stages of deglaciation in the Clyde basin (Fig. 6.7E) were accompanied by the deposition of lacustrine and deltaic sediment into 'Lake Clydesdale', which was dammed by an ice margin positioned at the Blantyreferme moraine [Bell, 1874] (Fig. 6.3). Drumlins of flow set 4 probably continued to form at this stage, as well as those belonging to flow set 5. Final retreat of Highland-sourced ice in the Clyde basin was accompanied by invasion of the contemporary sea, in which glaciomarine sediments were deposited up to altitudes 40 m a.s.l. [Browne and McMillan, 1989b].

6.4.2 3-D geological modelling results

The computed model of glacigenic deposits (total volume = 9.75 km^3) in the Clyde basin is shown in Figure 6.8. Volumes of individual units are given in Table 6.2 and modelled thicknesses of example units are shown in Figure 6.9. The buried till and buried glaciofluvial deposits are generally restricted to the concealed bedrock valley under the floodplain of the modern River Kelvin. Here, they exceed 40 m in thickness (Fig. 6.9), representing an important aquifer. Buried glaciofluvial deposits are also

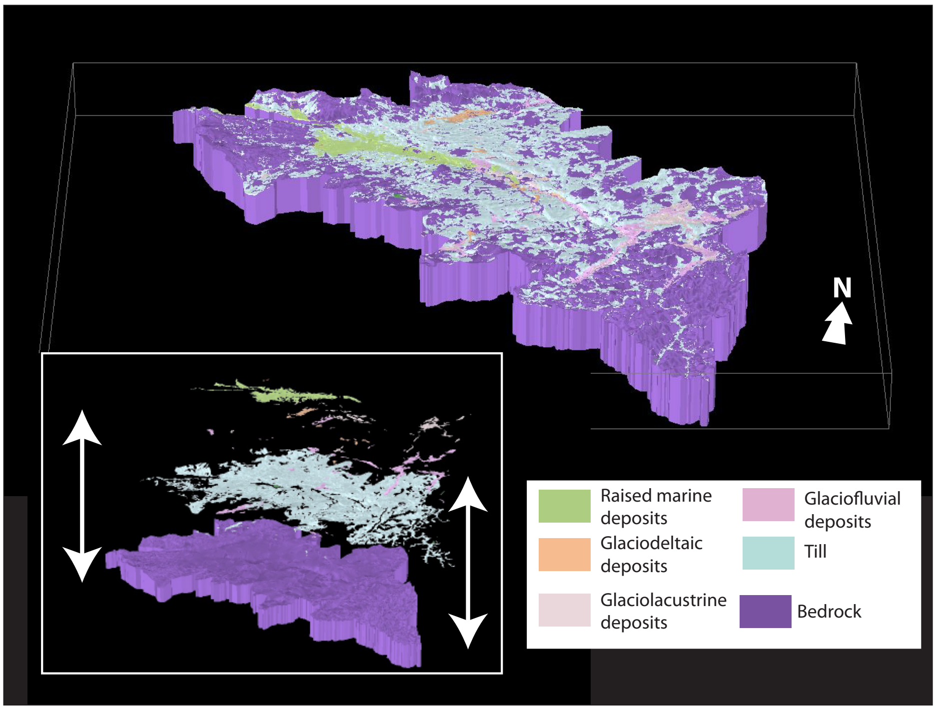

FIGURE 6.8: Computed geological model showing glacigenic deposits in the Clyde basin. Inset: geological model is vertically separated to show individual sediment packages.

present in concealed channels in the upper Clyde basin. Buried glaciolacustrine deposits are restricted in extent, only occurring as isolated pockets in deepenings under the River Clyde.

There is only one bulk till unit across much of the basin; it accounts for 70% of the total sediment in the basin. Variations in till thickness are apparent, ranging from 0.01 m to more than 40 m (Fig. 6.9). Median and mean modelled till thickness is 5.8 and 7.7 m respectively (standard deviation = 7.0). These values are consistent with the 6 m mean thickness calculated by Menzies [1981] for till in central Glasgow. Till is generally thin (or absent) around the margin of the basin (Fig. 6.9). The thickest till sequences are developed over the buried glaciofluvial deposits (Fig. 6.10), which are more permeable than the surrounding sedimentary rocks. The presence of thick till over these glaciofluvial gravelly sands supports suggestions elsewhere that efficient sub-till drainage can cause dewatering and stiffening of subglacially mobilised sediments, thereby promoting deposition [Boulton et al., 2001; Meriano and Eyles, 2009]. However, the distribution of thick till probably also partially reflects infilling of the pre-existing topography [Boyce and Eyles, 2000], since the buried glaciofluvial deposits tend to be preserved in bedrock depressions. Assuming the buried glaciofluvial

Geological model unit	Area (km^2)	Volume (km^3)
Raised marine deposits	160.86	1.29
Glaciodeltaic deposits	22.74	0.09
Glaciolacustrine deposits	72.39	0.32
Glaciofluvial deposits	140	0.67
Till	2259	7.2
Buried glaciolacustrine deposits	1.84	0.0038
Buried glaciofluvial deposits	32.7	0.15
Buried till	8.67	0.03
Total		*9.7538*

TABLE 6.2: Computed volumes of glacigenic sediments in the Clyde basin

deposits are correlative with the Cadder Sands and Gravel Formation (Table 6.1), the overlying till package in those areas can be considered wholly a product of the last glacial cycle. Figure 6.11 shows an example of one such till package, which overlies the largest deposit of buried glaciofluvial sediments in the Kelvin valley. Here a net volume of 0.62 km^3 was deposited over 32 km^2. Thick till is also present in an area immediately down ice from zones likely to have been occupied, initially, by soft deformable sediment during ice sheet advance (see below).

Glaciofluvial, glaciodeltaic and glaciolacustrine deposits are widespread alongside drainage pathways in the Clyde basin where they reach up to 20 m thickness, in places (Fig. 6.9). Collectively, these deposits represent 11% of the total sediment volume in the basin. The raised marine deposits in the lower Clyde basin provide the second largest contribution (13%) to the total basin sediment volume; these sediments exceed 20 m thickness along pre-existing topographic lows.

6.4.3 Sediment mobilisation during the last glacial cycle

Using the event stratigraphy (Fig. 6.7) and the geological modelling results, progress can be made towards a basin-scale reconstruction of sediment mobilisation during the last glacial cycle.

6.4.3.1 Ice sheet advance (Fig 6.7A)

It has been theorised that ice sheet advance is a key phase of till deposition [Boulton, 1996]. The preservation of sediments and bedforms interpreted to have begun formation during, or following, this initial stage of the glacial cycle (ribbed moraine and

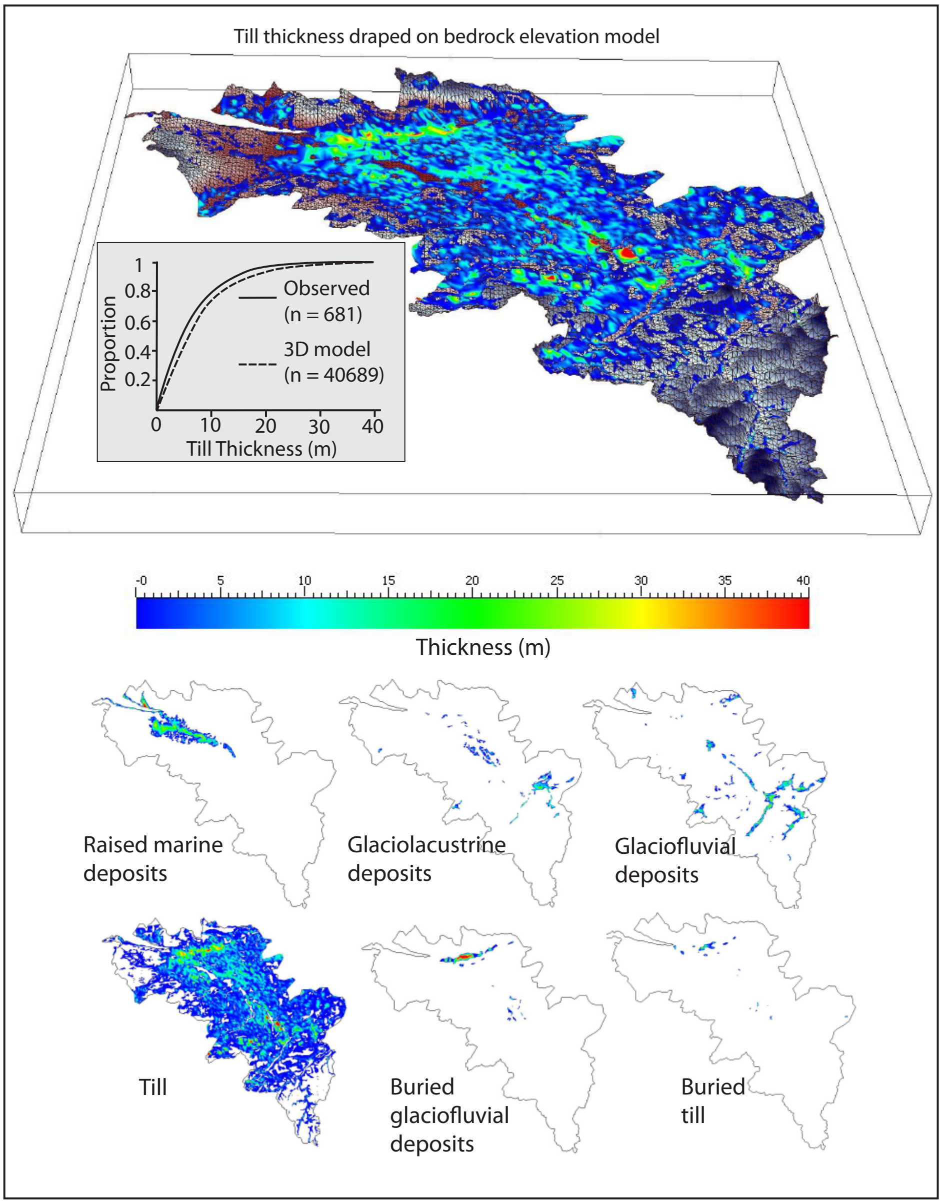

FIGURE 6.9: Thicknesses of selected units. Upper panel shows till thickness draped over bedrock elevation model. Cumulative plots are shown for till thickness revealed from borehole observations, and for till thickness represented in the geological model. The overall thickness distribution in the geological model is consistent with the population of till thickness records from boreholes, demonstrating that the model remains faithful to the observed data.

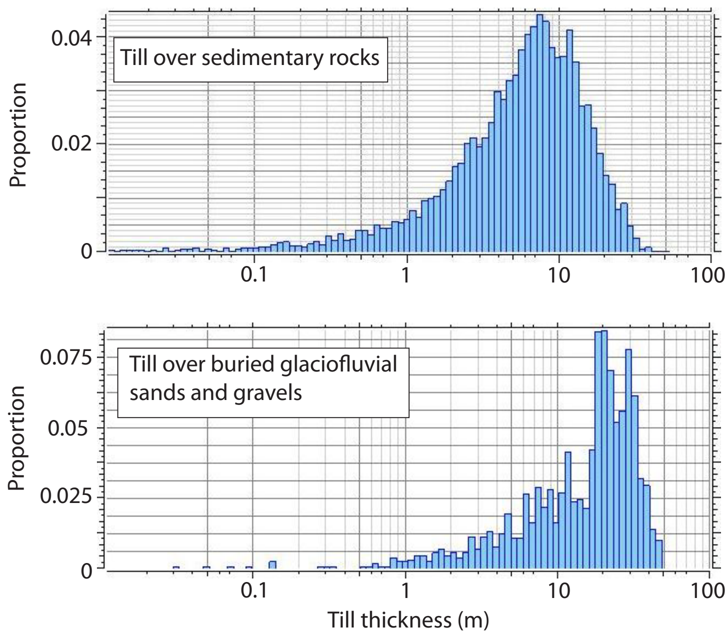

FIGURE 6.10: Histograms of modelled till thickness, in areas where: till rests on Carboniferous sedimentary rocks (upper plot); and till rests on glaciofluvial sands and gravels (lower plot).

flow set 1 drumlins), does indeed suggest that a significant volume of till had been deposited during the ice sheet advance phase. Sediment within drumlins is often found to pre-date the drumlin forming event [Knight and McCabe, 1997a; Stokes et al., 2011]. If this were also true for the Clyde basin drumlins, it also supports a case for an early phase of till deposition. The till package that rests on the buried fluvial deposits (Fig. 6.11) in the Kelvin valley is considered to have been deposited during the last glacial cycle (see above). Here 0.62 km^3 of till has been deposited over 31 km^2, which equates to a uniform net deposition rate through the glacial cycle ($\sim$35 ka BP to $\sim$15 ka BP) of 0.001 m a^{-1} (or m^3 m^{-2} a^{-1}). However, if much of that till package was deposited during advance through the basin (this could have occurred over as little as 0.5 ka, according to the recent BIS simulation of Hubbard et al. [2009]), which is a prerequisite for drumlins to have formed in those sediments, reconstructed net deposition increases to $\sim$0.04 m a^{-1}. These till deposition rates are only slightly higher than those proposed for ice marginal zones by Boulton [1996] and are of a similar order of magnitude to rates of subglacial till deposition considered by Sugden and John [1976], but they assume no latter erosion. This assumption can be partially justified here on two accounts: the Clyde basin was subsequently positioned close to, or under, an ice divide during maximum glaciation (Figs. 6.7B, 6.7C; Finlayson et al. [2010]), where minimal erosion would have been expected [Boulton, 1996]; and bedforms interpreted to have formed during early ice flow stages are preserved. It should be noted that these till accumulation rates cannot be used to estimate bedrock erosion, since a significant fraction was likely to have derived from remobilised sediment (discussed below).

It is clear, however, that till deposition during advance was not uniform across the basin; the 20 m thickness of till described above, in the Kelvin Valley, is not reproduced everywhere (Fig. 6.9). Pre-conditioning factors may therefore have influenced till accumulation. It is notable that no marine deposits and little glaciolacustrine deposits are revealed below till in the model. However, conditions prior to ice sheet build up indicate that such deposits were likely to have been locally extensive: global sea level was at least 6.6 m higher during the last interglacial [marine isotope stage (MIS) 5e; Kopp et al., 2009] and relative sea level was probably higher during the prior continental-scale ice sheet deglaciation (MIS 6); ice-dammed lakes are also likely to have accompanied ice sheet build up by blocking the River Clyde [Price, 1975]. Reworking of these deposits probably formed the bulk of till that accumulated during ice sheet advance [e.g. Ó Cofaigh et al., 2011]; indeed lenses of folded laminated sediments have been observed in till in the Glasgow area [Menzies, 1981]. The presence of the thickest till packages in the vicinity, or immediately down ice, from modern marine and glaciolacustrine sediments (Figs. 6.3, 6.9) lends support to the idea that similar sediments were previously a key source for till deposited during advance. Where ice sheet advance was into relatively soft, water-saturated sediments (e.g. glaciomarine, glaciolacustrine, or deltaic deposits), deformation probably ensued as plug flow [Leysinger Vieli and Gudmundsson, 2010]. Sediment removal up glacier may have been in excess of 20 m, based on the thickness of raised marine deposits presently occupying the Clyde basin. Subsequent deposition of thick till then occurred when glaciotectonic stress (lateral stress + shear stress, Aber et al. [1989]) fell below sediment strength, possibly through sediment dewatering as it moved over more permeable (buried glaciofluvial deposits) material, or as the ice margin passed over leading to a reduction in glaciotectonic stress. The result is that overall transport distances were relatively short, consistent with the limited down ice distribution of thick till sequences and the preservation of some sedimentary structures locally within the till. Where ice advanced over stiffer, pre-existing till, overriding or mixed flow [Leysinger Vieli and Gudmundsson, 2010] was more likely, with reduced basal sediment mobilisation and reduced accumulation of 'new' till. A switch to these conditions would also have taken place in zones previously characterised by plug flow, once the supply of relatively soft sediments became exhausted, otherwise the thick till wedge would have migrated down glacier with the advancing ice front. Thus, spatial variations in sediment mobilisation by deformation were probably governed by the relative stiffness of pre-existing substrate, with

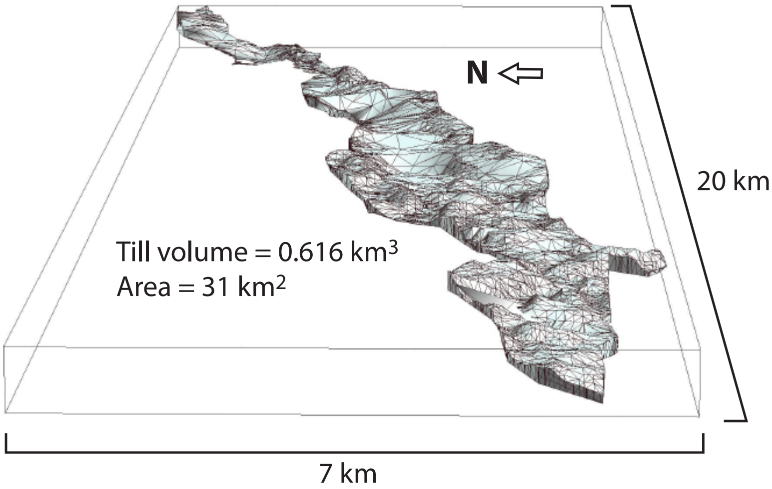

FIGURE 6.11: Till package overlying buried glaciofluvial deposits. This till package it likely to be wholly a product of the Late Devensian glacial cycle.

implications for the glacier flow mechanism.

A result of the geological modelling, and previous stratigraphical work in the Glasgow area [Menzies, 1981; Browne and McMillan, 1989b], is that, with the exception of those restricted locations where buried till packages are identified (Fig. 6.9), only one main till layer (median thickness = 5.6 m) is present in the basin. It is extremely unlikely that the ~7 km^3 of till forming that layer is wholly a deposit from the last glacial cycle. To generate that volume of till again would require ~3 times the volume of all other sediments presently available in the basin for reworking. Given the topographically confined (by Clyde Plateau Volcanic Formation) north-western entrance to the basin, which has little or no till cover (Fig. 6.9), transport of the remaining volume of till into the basin via a continuous deforming bed is not probable. It is more likely that till from previous glaciations remained in the basin and was, to an extent, reused by the next.

6.4.3.2 Ice sheet established; ice divide migration (Figs. 6.7B, 6.7C)

Coalescence of Highland and Southern Upland ice masses allowed an ice divide to develop in the vicinity of the Clyde basin. The transition of the Clyde basin, from being positioned down ice from the equilibrium line, to being positioned close to the ice divide (Figs. 6.7A, 6.7B) meant that overall sediment erosion was limited, since maximum erosion is thought to occur just up ice of the equilibrium line [Boulton, 1996]. The initial position of the ice divide ∼30 km to the west facilitated eastward flow associated with drumlins of flow set 1; these probably formed by erosion and deposition by a mobile till layer [Boyce and Eyles, 1991]. It is not possible to quantify the volumes involved; however, the presence of much thinner, and in places absence of, till at the eastern boundary of the basin (Fig. 6.9), suggest that overall sediment loss through continuous bed deformation would have been restricted. Furthermore, preservation of both sediments and landforms from early parts of the glacial cycle [Browne and McMillan, 1989b; Finlayson et al., 2010] indicates that basal motion was probably concentrated at or near the ice-sediment interface, limiting large-scale sediment transport by deformation. Subsequent migration of the ice divide to a position over the Clyde basin (Fig. 6.7C) would have further limited any sediment mobilisation.

Eastward ice flow (Fig. 6.7C) became more established towards the end of this phase, as deglaciation commenced. It is interesting to note that there is no difference in characteristics of the eastward oriented drumlins between zones of buried glaciofluvial deposits and adjacent areas (Fig. 6.12). Lengths (L) and elongation ratios (ER) for drumlins overlying buried glaciofluvial deposits (n = 46, median L = 645 m, median ER = 2.6) and drumlins within a 2 km zone beyond the margin of the buried glaciofluvial deposits (n = 81, median L = 660 , median ER = 2.9) were statistically indistinguishable (Mann Whitney: $z = 0.9$, $p = 0.37$ for L; $z = 0.91$, $p = 0.36$ for ER). Drumlin elongation has been suggested to be a proxy for relative basal ice flow velocity [Stokes and Clark, 2001], while basal motion is linked to subglacial water pressure [Paterson, 1994]. It is therefore likely that enhanced drainage through the buried glaciofluvial gravelly sands (which were at the land surface prior to advance), was precluded by the thick till preferentially deposited on top of these deposits (Fig. 6.10). This would have equalised effective pressures between zones underlain by buried fluvial deposits and adjacent areas. Thus, by focusing thick till sequences over more permeable substrates during advance, the ice sheet may have essentially regulated its basal conditions, facilitating smoother overall basal motion.

6.4.3.3 Ice cap uncoupling and deglaciation (Figs. 6.7D, 6.7E)

During and following ice cap uncoupling, glacier flow in the lower Clyde basin was generally towards the southeast, documented by flow set 4. The drumlins formed during this stage are some of the best preserved in the basin and indicate that till erosion and down-ice deposition was locally occurring at this time. However, for reasons outlined above, it is unlikely that significant volumes of till were lost from the basin by continuous deformation.

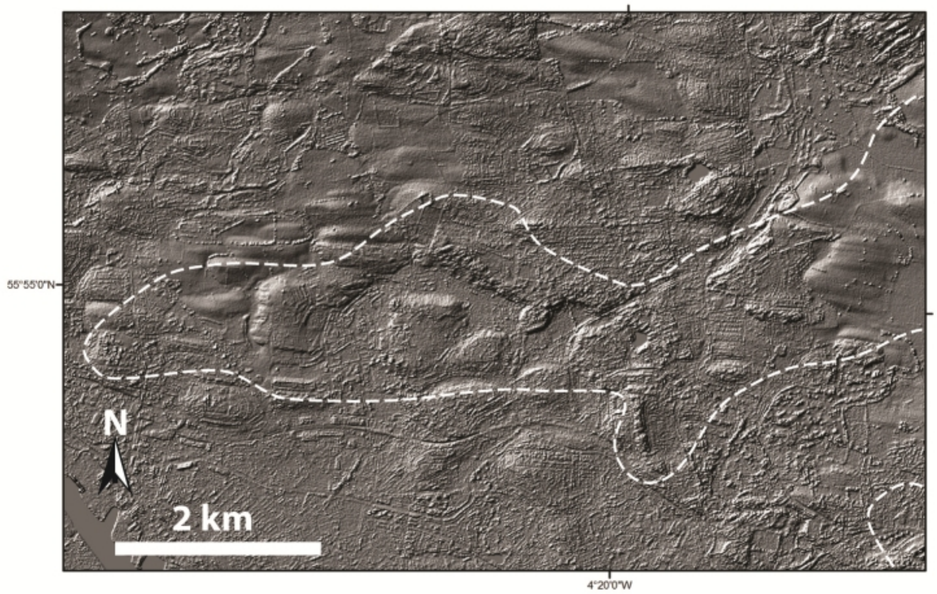

Figure 6.12: Drumlinised terrain in north Glasgow. The zone of buried glaciofluvial deposits revealed by the geomodel is delimited by the white dashed line. The extent of streamlining is statistically indistinguishable inside and outside this zone of buried glaciofluvial deposits.

The combined volume of the glaciofluvial, glaciolacustrine, glaciodeltaic deposits is 1.08 km^3, giving an approximation for the minimum volume of sediment remobilised (predominantly by sub-/pro- glacial fluvial processes) during ice sheet withdrawal. The approximation is a minimum since drainage was diverted eastward (via successive overflow cols towards the Firth of Forth) for a period following ice cap decoupling, which would have resulted in some unaccounted sediment loss, and sediment loss to the Firth of Clyde is not included. These sediments make up 11% of the total sediment volume in the basin. The duration over which they were deposited only represents the final 2.5% of the last glacial cycle in the area ($\sim$0.5 ka maximum, based on a synthesis of dates and ice margin retreat isochrones published in Clark et al. [2012]), indicating that transport and deposition by glacial meltwater formed a disproportionally high contribution to the basin sediment budget as the ice margin retreated back though the basin. This supports the suggestion by Alley et al. [1997] that high sediment transport capacities are achieved by subglacial streams in ice marginal areas, where surface water accessing the bed, promotes high water discharges forced by steep head gradients.

6.5 Wider implications

The work presented here has attempted to characterise the volume and style of sediment mobilisation through the last glacial cycle in the Clyde basin. During that time the basin was occupied by an advancing outlet glacier; it was then positioned under a migrating ice divide, and then once more by an outlet glacier during ice sheet uncoupling and deglaciation. The data presented here may therefore provide insight into sediment mobilisation in such environments. Although the Clyde basin may have temporarily been located close to the onset zone of fast glacier flow, it was not occupied by an ice stream and is thus not representative of ice streaming conditions.

Ice margin advance is shown to have been associated with highly variable spatial and temporal pattern of sediment mobilisation. Where ice advanced into relatively soft glaciomarine and glaciolacustrine deposits, sediment fluxes in the marginal area were high. Only rare opportunities exist to study modern glaciers advancing into soft sediments, but where observed, high fluxes are apparent. For example, advance of the Taku Glacier, Alaska, which is a good modern analogue for the Clyde glacier in its advance stage, has displaced >100 m of marine sediment from its bed since 1890 [Nolan et al., 1995]. In the case of the Clyde glacier, this high rate of sediment mobilisation and redeposition was spatially, and probably temporally, restricted. Reduced basal sediment mobilisation took place where relatively soft sediments were absent, or following the exhaustion of soft sediments.

The preferential deposition of thick till over glaciofluvial sediment appears to have acted as a seal, negating the increased drainage capacity of the latter. This suggests that glacier advance and till deposition can alter basal conditions by promoting lower effective pressures. As a result, motion is focused at the ice-till interface, and ice infiltration into the bed by regelation inhibited [Iverson, 2010]. In the case of the Clyde basin, this would have favoured preservation of both sediments [Browne and McMillan, 1989b] and bedforms [Finlayson et al., 2010] that were produced early in the glacial cycle. However, localised erosion and deposition by a relatively shallow layer of flowing sediment [Boyce and Eyles, 1991; Clark, 2010] must also have occurred in order to carve bedforms during subsequent phases of the glacial cycle.

The Clyde basin is a relatively small area at the ice-sheet scale. Thus, it highlights the spatially variable nature of some bed properties over small areas. The highly variable modelled till thickness (Fig. 9) (mean = 7.7 m, standard deviation = 7.0 m) demonstrates that uniform till sheets [Alley, 1991] do not necessarily occur in all lowland areas.

A considerable volume of sediment in the Clyde basin was remobilised by glacial meltwater during ice sheet withdrawal, highlighting its role as a geomorphological agent. While the subglacial imprint (e.g. drumlins) often produces the most striking landscape features, ice margin advance and deglaciation are suggested, in the case of the Clyde basin, to have been key (possibly more important) factors influencing overall sediment mobilisation during the glacial cycle.

Chapter 7

Digital surface models do not always represent former glacier beds: palaeoglaciological and geomorphological implications

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Abstract

Quantitative palaeoglaciological studies that use digital surface models (DSMs) may be subject to error because former glacier beds are not always accurately represented. This is because the Earth's surface may have changed significantly since deglaciation. This paper evaluates potential errors caused by postglacial sedimentation, by comparing the results of physical palaeoglaciological reconstructions and bedform morphometric analyses in parts of Scotland, using both the modern land surface and interpolated former glacier beds derived from borehole data. For a former terrestrial outlet glacier, removal of postglacial sediments increases the modelled ice surface elevation and ice thickness by 0.7% and 5%, respectively, over a 27-km flow line. For a former tidewater glacier, the reconstructed steady state ice flux is increased by 250% when the modern land/seabed surface is replaced with an interpolated former glacier bed. In a classical drumlinised landscape, removal of postglacial sediments affects bedform morphometrics, with an increase in measured drumlin length, width, relief, and volume. The cases presented in this paper are from environments known to have experienced postglacial sedimentation. They provide situational examples of the degree of error that can be introduced when the modern land surface is used to represent former glacier beds in these environments. In some regions, sufficient subsurface data exists over large areas to create improved topographic

representations of former glacier beds; these could form important inputs to the next generation of palaeo-ice sheet and palaeoglacier simulations.

7.1 Introduction

Digital surface models (DSMs) can provide high resolution geomorphological information about the Earth's surface. They are used to represent past glacier beds for the reconstruction of former ice sheets [Lidmar-Bergström et al., 1991; Ó Cofaigh et al., 2009; Trommelen and Ross, 2010], numerical palaeoglacier simulations [Plummer and Phillips, 2003; Golledge et al., 2008] and statistical analyses of glacier bedform morphometrics [Dunlop and Clark, 2006; Clark et al., 2009a; Hess and Briner, 2009]. When using DSMs, geomorphologists have to assess the risk of any land surface change, in the time between glacier ice occupation and capture of elevation data, having affected the geomorphic expression of the former glacier bed. Commonly, these changes are too small to introduce significant error to conceptual palaeoglaciological reconstructions. However, the importance of bed topography to numerical simulations and quantitative morphometric assessments could make them prone to errors if the land surface has been considerably lowered by erosion or raised by sediment deposition after glacier retreat. In lowland and coastal areas, high rates of sedimentation have been shown to accompany, and immediately follow, deglaciation [Eyles et al., 1990; Cowan and Powell, 1991; Leventer et al., 2006], and some workers have recognised that sediment laid down after ice margin retreat (hereafter referred to as postglacial sediment) may lead to errors in quantitative studies [Piotrowski and Tulaczyk, 1999; Golledge et al., 2012; Spagnolo et al., 2012]. The goal of this paper is to test how the results of simple quantitative palaeoglaciological investigations differ when the topographic expression of postglacial sediments is removed from DSMs, thereby providing some indication of the error for a given set of examples.

Three case studies from parts of Scotland that were deglaciated ca. 15 ka BP (Fig. 7.1A) are presented. Reconstructed glacier characteristics and bedform morphometric analyses obtained using the modern land surface are compared with those derived from interpolated former glacier beds based on densely spaced borehole data. The first case considers differences in valley shape and the effects on a glacier surface profile calculated using an iterative flowline model in the Clyde basin, west-central Scotland (Fig. 7.1B). The second case examines the differences in reconstructed calving rates and hypothetical ice fluxes at a former tidewater glacier margin in the Cromarty Firth, northeast Scotland (Fig. 7.1C). The third case compares three-dimensional morphometric measurements for a small sample of drumlins in southwest Glasgow (Fig. 7.1D).

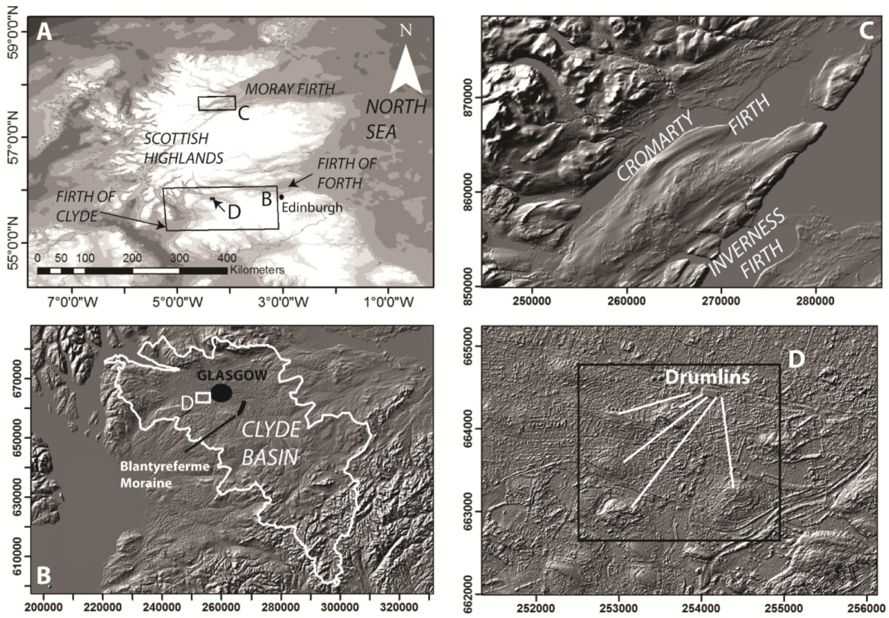

FIGURE 7.1: (A) Location of case study sites within national context. (B) Clyde basin, west-central Scotland. White line marks the area included in the geological model described in Finlayson [2012]. (C) Cromarty Firth, northeast Scotland. (D) Drumlinised terrain in southwest Glasgow. Images derived from ETOPO1 Global Relief Model (A) and Intermap Technologies NEXTMap Britain elevation data (B,C,D). Coordinates in B, C, and D in British National Grid.

The examples are chosen from near-coastal areas known to have been subjected to postglacial sedimentation. Such environments are often included in palaeoglaciological studies, and coastal margins are recognised as key dynamic zones of past ice sheets. Therefore a requirement exists to quantitatively evaluate potential errors that may be introduced into palaeoglaciological models and bedform measurements by unrepresentative DSMs in these areas.

7.2 Study area and methods

7.2.1 Clyde basin: valley shape and former glacier profile

The first study focuses on valley shape and reconstructed glacier surface profile in the Clyde basin, west-central Scotland (Fig. 7.1B) at the time when the Blantyreferme moraine was formed. During overall ice sheet retreat, the Blantyreferme moraine was constructed in the lower part of the Clyde basin by an outlet glacier, sourced from an ice cap centred over the Scottish Highlands [Price, 1975; Finlayson et al., 2010, this thesis' Chapter 4]. Glacier flow at that time was toward the southeast. Final glacier decay in the lower Clyde basin was accompanied by relative sea level rise to almost 40 m above present when thick sequences of glaciomarine silts and clays were laid down, partially masking the former glacier bed [Browne and McMillan, 1989b; Peacock, 2003].

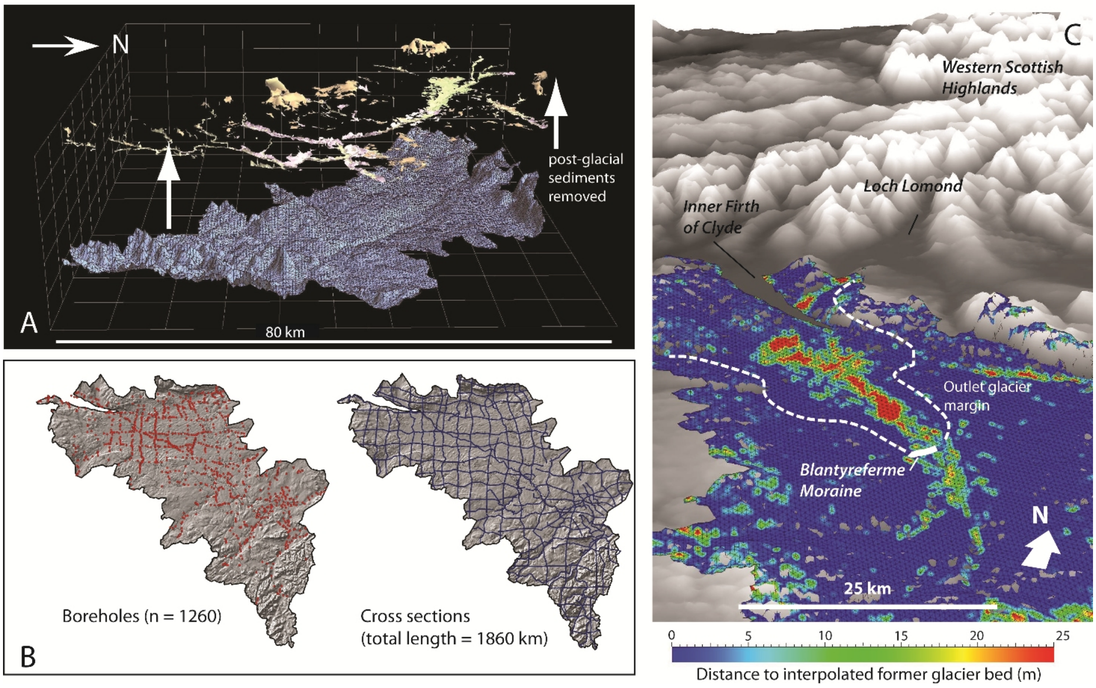

FIGURE 7.2: (A) Removal of 2.37 km^2 of postglacial sediments in the Clyde basin reveals interpolated former glacier bed. (B) Location of boreholes and cross section lines used to control interpolations in the model described by Finlayson [2012]. (C) Distance from modern land surface to the former glacier bed. Dashed white line shows interpreted margin of outlet glacier at the time when the Blantyreferme moraine was formed.

In the Clyde basin, the former glacier bed was extracted from the three-dimensional geological model described by Finlayson [2012, this thesis' Chapter 6]. The model (Fig. 7.2A) adheres to surface sediment distribution shown on 1:50,000-scale geological maps and subsurface data derived from 1260 borehole logs (Fig. 7.2B). It comprises a series of surfaces, representing the tops and bases of lithostratigraphic units, derived through triangulation of regularly spaced x,y,z nodes along cross sections (total length 1860 km) and 'envelopes,' which represent the lateral (surface and buried) extent of lithostratigraphic units. The model was calculated at a 500-m grid resolution. Postglacial lithostratigraphic units, representing 2.37 km^3 of sediment, were removed to obtain an interpolated glacier bed, more closely representing bed topography at the time when the Blantyreferme moraine was formed (Fig. 7.2C). The interpolated former glacier bed is overdeepened and lies below the postglacial and modern sea level. The depth of the interpolated bed is therefore unlikely to have been enhanced by postglacial fluvial erosion.

The influence of postglacial sediments on glacial valley shape was examined in the lower part of the Clyde basin, using valley shape factor (f). Shape factor is used to account for the part of a glacier's weight that is supported by the valley sidewalls; it defines the proportion of driving stress (τ_D) that is transferred to basal shear stress (τ_B) at the valley centre, so that $\tau_B = f\tau_D$. Driving stress is calculated from

$$\tau_D = \rho_I g H \, tan\alpha \tag{7.1}$$

where ρ_I is the density of glacier ice ($\sim$ 900 kg m^{-3}), g is gravitational acceleration (9.81 m s^{-2}), H is glacier thickness (m) and α is glacier surface slope. For a flat bed (which presents no side drag), $f = 1$, and for a semi-ellipse-shaped valley with a half-width equal to centre-line ice thickness $f = 0.5$ [Paterson, 1994]. Shape factor can be calculated from

$$f = \frac{A}{HP} \tag{7.2}$$

where A is the cross-sectional area of the valley that is filled with glacier ice, and P is the cross-sectional perimeter that is in contact with glacier ice. The approach adopted here is that used by Benn and Hulton [2010] in which A is calculated along the cross section from

$$A = \sum_{i=1}^{n} \frac{((B_{MAX} - B_i) + (B_{MAX} - B_{i+1}))\,\Delta y}{2} \tag{7.3}$$

where B is the glacier bed elevation, and Δy is the horizontal step size across the valley. The value for P is obtained from

$$P = \sum_{i=1}^{n} \sqrt{(B_{i+1} - B_i)^2 + (\Delta y)^2} \tag{7.4}$$

To test the effect that removing the postglacial infill has on the reconstructed glacier profile, an iterative valley centre flowline model was applied:

$$h_{i+1}^2 - h_{i+1}(B_i + B_{i+1}) + h_i(B_{i+1} - H_i) - \frac{2\Delta x(\tau_B/f)}{\rho_I g} = 0 \tag{7.5}$$

where h is ice surface elevation, and x is the horizontal coordinate along the valley centre line. The solution to Eq. (7.5) is usefully described by Benn and Hulton [2010] who provide an accompanying spreadsheet program.

7.2.2 Cromarty Firth: former calving speed and ice flux

The Cromarty Firth (Fig. 7.1C) is a long marine inlet in northeast Scotland that was a tributary to the former Moray Firth Ice Stream (MFIS) [Merritt et al., 1995]. Former ice flow in the area was toward the northeast, broadly parallel to the alignment of the Cromarty Firth. During deglaciation, calving ice fronts of the MFIS retreated into the Cromarty and Inverness Firths when relative sea level was at least ~ 30 m above present [Peacock, 1974; Firth, 1990; Merritt et al., 1995; Turner et al., 2012]. Following glacier withdrawal, deposits of glaciomarine and marine sediments exceeding 60 m in thickness were laid down in the Cromarty Firth [Peacock, 1974].

The former glacier bed was manually interpolated along a single cross section using borehole logs and descriptions from Peacock [1974]. The section line is approximately normal to former glacier flow direction as ice retreated into the Cromarty Firth during deglaciation [Firth, 1990; Merritt et al., 1995] (Fig. 7.3A and 7.3B).

To examine the effect that removal of postglacial sediments has on former calving speed (U_C) in the Cromarty Firth, an empirically derived water depth (D_W) relation was used [Brown et al., 1983; Pelto and Warren, 1991]:

$$U_C = 70 + 8.33 D_W \tag{7.6}$$

Although more robust, physically based calving laws now exist [Benn et al., 2007; Nick et al., 2010], they require input variables not easily obtained from the palaeorecord. The water depth relation used here describes real calving rates in many instances [Hooke, 2005], and its simplicity allows it to be employed in numerical models with minimal computation [Golledge et al., 2008].

Ice thickness at the calving palaeoglacier margin (H_C) is calculated based on the floatation criterion used by Vieli et al. [2001, 2002], in which the calving margin is located where the glacier approaches floatation thickness. H_C can be calculated from

$$H_C = (1+q)\frac{\rho_{SW}}{\rho_I} D_W \tag{7.7}$$

where q is a fraction representing the height of the ice front above buoyancy at the calving margin, and ρ_{SW} is the density of sea water (1030 kg m^{-1}). A value of 0.15 is adopted for q [Vieli et al., 2001]. Application of Eq. (7.7) enables cross-sectional area (A_C) of the calving margin to be calculated (Fig. 7.3C). Depth-averaged velocity (U)

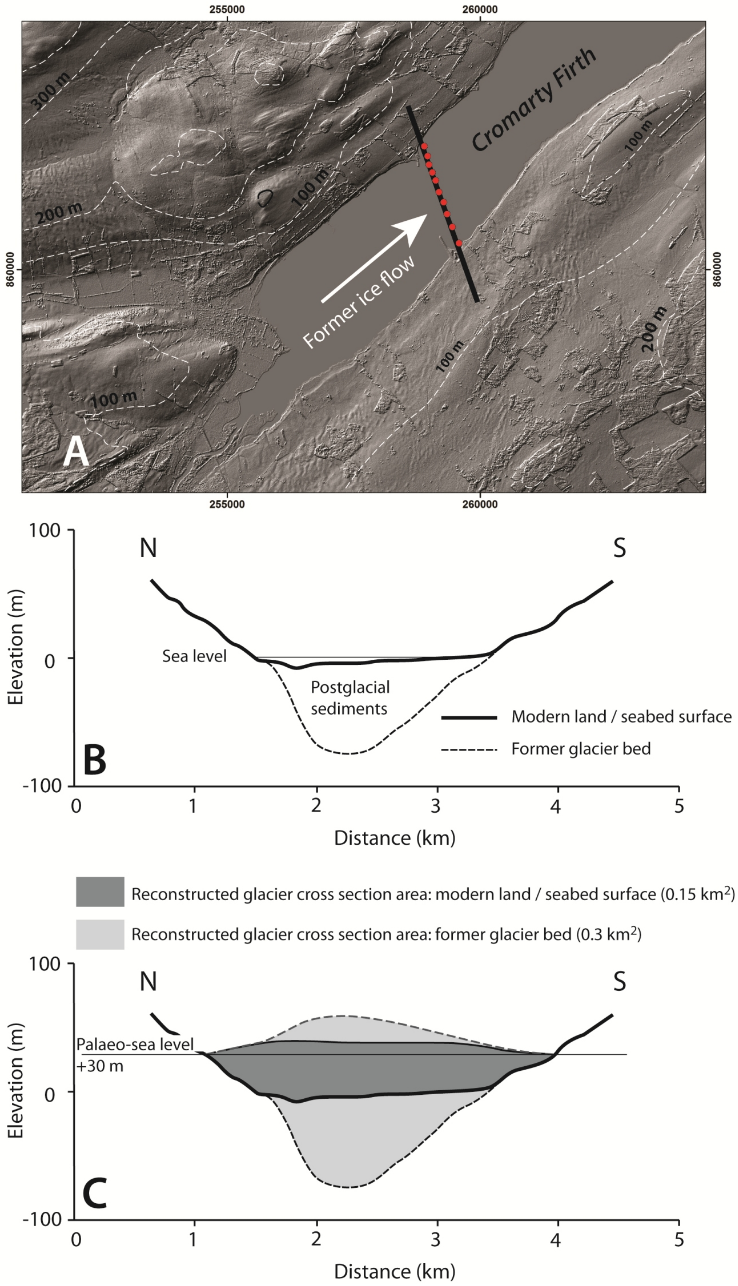

FIGURE 7.3: (A) Location of boreholes (red) and position of cross section (black line) in the Cromarty Firth, northeast Scotland. Hill-shaded digital surface model built from Intermap Technologies NEXTMap Britain elevation data. (B) Cross sectional line showing the thickness of postglacial sediments. x 10 vertical exaggeration. (C) Reconstructed cross section areas for a calving glacier in the Cromarty Firth, based on the modern seabed surface and the interpolated former glacier bed.

at the ice front is given by the equation

$$U = U_C - Q_M + \frac{\Delta L}{\Delta t} \tag{7.8}$$

where Q_M is ice loss caused by melting, L is glacier length, and t is time. Under steady state conditions $\Delta L/\Delta t = 0$. Assuming negligible melt, ice flux (Q_I) at the calving margin can be calculated from

$$Q_I = U_C \, A_C \tag{7.9}$$

7.2.3 Southwest Glasgow: drumlin morphometric analyses

The third example focuses on morphometric measurements of drumlins in southwest Glasgow (Fig. 7.1D). The lower Clyde basin is well known for its drumlins, which have been included in several regional and national morphometric data sets [Menzies, 1996; Clark et al., 2009a; Finlayson et al., 2010; Spagnolo et al., 2012]. Many of these drumlins occur in areas inundated by the Lateglacial sea following deglaciation of the Glasgow area and are partially overlain by glaciomarine deposits [Peacock, 2003; Finlayson et al., 2010].

For a 6.1-km^2 area in southwest Glasgow, the glacier bed was interpolated at a 25-m grid spacing using ordinary kriging, based on bed surface points interpreted from 144 borehole records (Fig. 7.4). The surface of the former glacier bed was readily distinguished in borehole logs from contrasting sedimentary and geotechnical properties between glacial till or bedrock and the overlying, generally soft, raised marine silts and clays. One hundred regularly spaced additional points were added where glacial till is shown at the modern land surface on digital geological maps (1:10,000 scale DiGMapGB-10) prior to kriging in order to achieve a uniform spatial distribution of control points. These additional points were sampled from the Intermap Technologies NEXTMap Britain elevation data set (5-m horizontal resolution subsampled to 25 m). Areas where till is shown at the modern surface in geological maps were then removed from the surface produced by kriging and replaced with extracts from the NEXTMap data set. The result is a single DSM that better represents the former glacier bed (Fig. 7.4).

To test the extent to which removal of postglacial sediments affects the results of glacier bedform analysis, drumlin morphometrics were examined using the modern land surface and interpolated glacier bed in southwest Glasgow (Fig. 7.4). Drumlin length, width, relief, area, and volumes were measured. Measurement of drumlin relief followed the method of Spagnolo et al. [2012], whereby a planar drumlin base was interpolated from its outline (identified by break in slope). The maximum vertical difference (a) between drumlin surface and the planar base (dipping at slope angle, θ) can be used to define drumlin relief (r) where

$$r = a \, sin(90 - \theta) \tag{7.10}$$

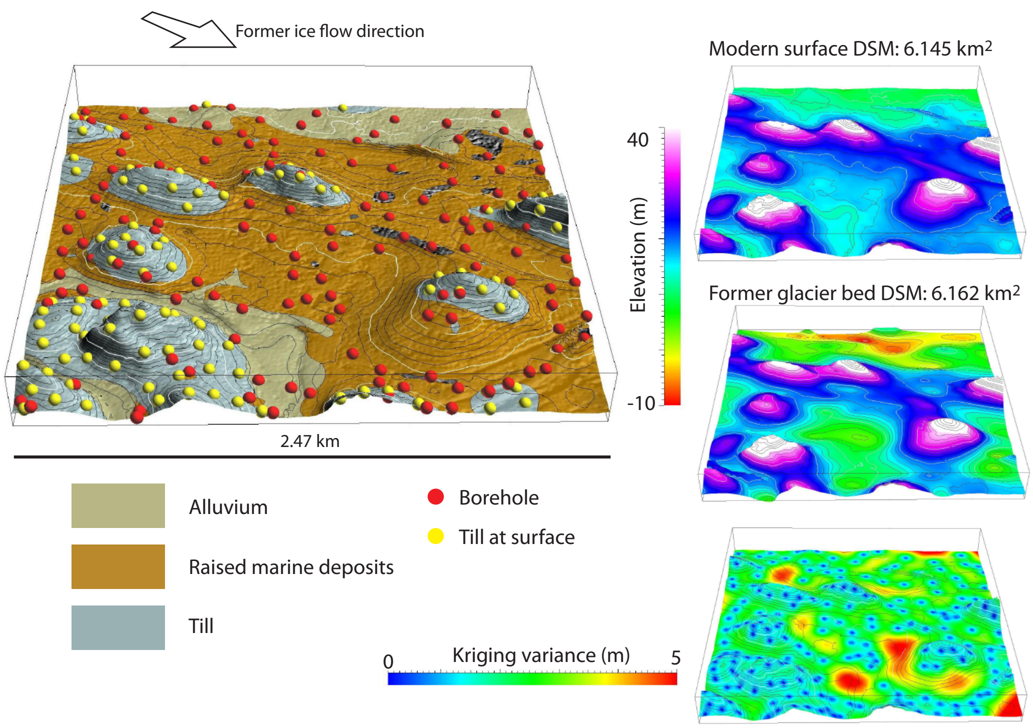

FIGURE 7.4: Modern DSM, draped with surface geology, showing location of boreholes and surface data points used for ordinary kriging, southwest Glasgow. The modern land surface and interpolated former glacier bed are shown on the right-hand side. Kriging variance gives an indication of potential error that may be expected for the interpolated surface and is largely influenced by distance to nearest data point.

Drumlin volume was calculated by combining the triangulated drumlin surfaces and drumlin bases to form solid objects comprising numerous tetrahedrons whose volumes were summed.

7.3 Results

7.3.1 Valley shape and reconstructed ice surface profile in the Clyde basin

Valley shape factor and ice surface profiles were calculated for an outlet glacier in the lower Clyde basin at the position of the Blantyreferme moraine (Fig. 7.5). Shape factors calculated using both the modern land surface and the interpolated former glacier bed are shown in Fig. 7.5A. Removal of postglacial sediments causes a deepening of the cross valley profile and an increase in length of the cross valley perimeter. As a result the calculated shape factors for six investigated cross profiles are reduced by 2%–18%. The reconstructed ice surface profiles and ice thickness are shown in Figures

7.5B and C. For the purposes of the reconstruction, a constant basal shear stress of 25 kPa was assumed; this is consistent with inferences of relatively low effective pressures during deglaciation in the lower Clyde basin [Finlayson, 2012]. The overall effect of removing postglacial sediments in this example is relatively small, and the reconstructed ice surfaces closely follow each other (Fig. 7.5B). The reconstructed ice surface slope derived from the interpolated glacier bed, however, is slightly steeper, compensating for the greater proportion of driving stress supported by valley sides (shown by the shape factor calculations). This increase is largely offset by the lower bed elevation of the interpolated surface. Best fit lines over the whole 27-km flow line show that the reconstructed ice surface elevation is 0.7% greater and the reconstructed ice thickness is 5% greater when modelled using the interpolated glacier bed (Figs. 7.5D, E).

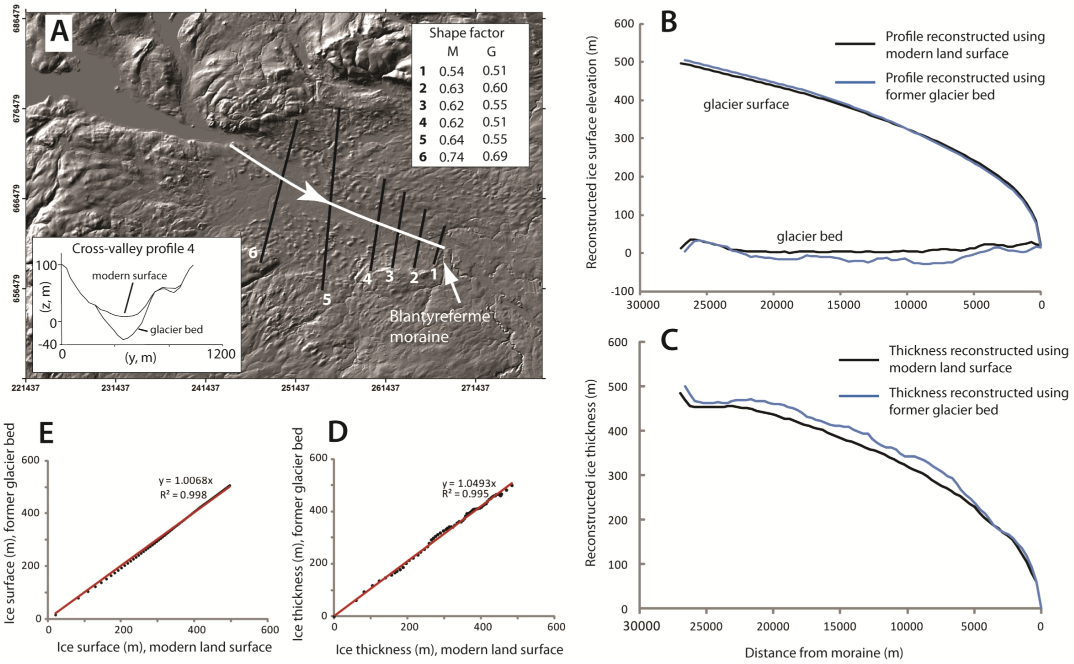

FIGURE 7.5: (A) Shape factors calculated for six cross-valley profiles in the lower Clyde basin, using the modern land surface (M) and the interpolated former glacier bed (G). White arrow denotes the ice flowline shown in B, C, D, and E. Hill-shaded digital surface model built from Intermap Technologies NEXTMap Britain elevation data. Inset: example of both surfaces across profile 4. (B) Reconstructed ice surface elevations using the modern land surface and the interpolated former glacier bed. (C) Reconstructed ice thickness, using the modern land surface and the interpolated former glacier bed. (D) Comparison of reconstructed ice thickness, calculated using the modern land surface and the interpolated former glacier bed. (E) Comparison of reconstructed ice surface elevations, calculated using the modern land surface and the interpolated former glacier bed.

7.3.2 Reconstructed calving rate and ice flux in the Cromarty Firth

Figure 7.3C illustrates the reconstructed glacier cross sections at the hypothetical calving margin in the Cromarty Firth, based on the modern land / seabed surface and the interpolated glacier bed. Removal of postglacial sediment effectively doubles the cross-sectional area of the glacier. Furthermore, based on the water–depth relation, calculated width-averaged calving speed for the glacier increases by 74%, resulting in an $\sim 250\%$ increase in steady state ice flux (Table 7.1). The removal of postglacial sediments also exposes more of the valley sides (the shape factor reduces from 0.7 to 0.5), which would increase drag and have implications for the necessary driving stresses.

Surface	**Width-averaged calving speed (m a^{-1})**	**Cross sectional area (km^2)**	**Steady-state flux (km^3 a^{-1})**
Modern	283	0.15	0.043
Glacier bed	494	0.3	0.148

Table 7.1: Characteristics of calving margin, reconstructed from the modern land surface and from the interpolated former glacier bed

7.3.3 Drumlin morphometrics in southwest Glasgow

The morphometric characteristics for five sample drumlins, derived from the modern land surface and from the interpolated glacier bed, are given in Table 7.2. Drumlins were delimited by their breaks in slope (Fig. 7.6). Using the modern surface model, a clear break in slope was apparent allowing straightforward, objective identification of each drumlin perimeter. However, breaks of slope are not as clear for the interpolated glacier bed (Fig. 7.6). This may partially result from the kriging procedure, which like many interpolation techniques has a tendency to underestimate highs and overestimate lows. It may also be that some of these particular drumlins have wave-like, rather than blister-like long profiles [Spagnolo et al., 2012] (far larger sample populations would be required to investigate this further). As a result, delimiting the perimeter of drumlins from the interpolated glacial land surface is slightly more subjective.

For drumlins 1, 2, 3, and 4, all morphometric characteristics increase following removal of postglacial sediments, resulting in volume increases of 37%–119%. Drumlin 5 is not affected because its interpreted perimeter lies beyond the extent of any modelled postglacial sediment infill.

The drumlins in this study have relatively low elongation ratios (ERs) (<1.7). However, their morphometric characteristics are within the range identified by Clark et al. [2009a]

Drumlin	Length (m)	Width (m)	Elongation ratio	Height (m)	Area (m^2)	Volume (m^3)
1: M	700	431	1.62	27.1	235856	1778684
1: G	730	477	1.53	32.7	291131	2719984
2: M	734	460	1.60	22.6	287120	1826651
2: G	895	600	1.49	29	442955	4002971
3: M	637	482	1.32	20.1	239769	1518055
3: G	900	568	1.58	20.4	390160	2096181
4: M	775	657	1.18	25.2	404264	2905079
4: G	1085	661	1.64	30	538266	4002242
5: M	625	450	1.39	37	227835	2541572
5: G	625	450	1.39	37	227835	2541572

TABLE 7.2: Characteristics of drumlins in southwest Glasgow, measured using the modern land surface (M) and the interpolated former glacier bed (G).

for drumlins in Britain. The undulating topography of the inter-drumlin area on the interpolated glacier bed reveals a subtle ridge linking drumlins 1 and 4 (Fig. 7.6). Some of these drumlins may represent the upper parts of larger, ribbed-moraine-like features, which have been identified elsewhere in the Clyde basin [Finlayson et al., 2010], and this could perhaps explain the relatively low ERs. However, a far larger sample area and data set would be required to test if this is the case.

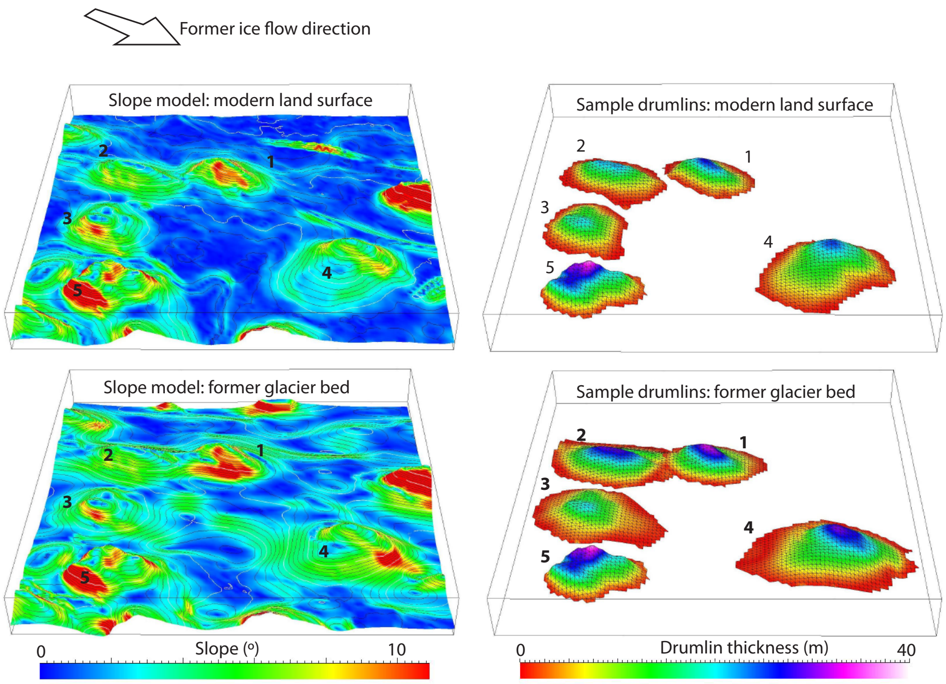

FIGURE 7.6: Slope models used to aid identification of drumlin perimeters for the modern land surface and for the interpolated former glacier bed, southwest Glasgow. The drumlins are shown as individual solid objects on the right hand side. Note that only those drumlins with their entire perimeter falling within the study area were included.

7.4 Discussion

The examples given in this paper focus exclusively on the influence of postglacial sedimentation upon a set of palaeoglaciological reconstructions and bedform morphometric measurements, which are based on analysis of the land surface. However, no account is made for postglacial erosion of parts of the bed. While the influence of erosion is expected to have been minimal in these largely depositional lowland settings, it is noted as an unknown source of error that is not included in this study.

The results demonstrate that removal of postglacial sediments from the land surface can affect the outcomes of investigations by varying amounts. In the Clyde basin, despite the thickness of postglacial sediments (up to 35 m), ice surface elevations reconstructed using the flowline model are only slightly affected (0.7% difference). Similar flowline models are often used in palaeoglacier and palaeo-icecap reconstructions [Locke, 1995; Rea and Evans, 2007; Hughes et al., 2011b; Finlayson et al., 2011, this thesis' Chapter 3], and the results presented here demonstrate one case where error introduced by postglacial sediment fill is relatively small. However, ice thickness variations resulting from postglacial sediment removal ($\sim$ 5% over the 25-km flowline in the lower Clyde basin) may have a significant bearing on ice volume estimates, particularly where numerous in-filled valleys are included within a study.

Of the examples presented here, the effects of postglacial sediment removal on the reconstructed calving glacier in the Cromarty Firth are perhaps the most significant, with large increases in both reconstructed calving speed and the required steady state ice flux. The thick sequences of deglacial and postglacial sediments in the Cromarty Firth are not unique. In fact, they may be the norm. For example, around the Scottish coastline postglacial sediments reach 47 m in the inner Moray Firth [Andrews et al., 1990], $>$ 50 m in parts of the Firth of Forth [B.G.S., 1987b], 50–70 m in outer Loch Broom [Stoker et al., 2006], and $>$ 100 m in parts of the outer Firth of Clyde [B.G.S., 1985]. Removal of thick postglacial sediments from near-shore former ice-marginal environments could have an influence on the behaviour of numerical palaeoglacier models, significantly increasing rates of simulated ice loss. This is one possible explanation for mismatches that sometimes occur when comparing simulated glaciers with empirical data. However, such influences may be restricted to near-shore environments as postglacial sediments tend to reduce in thickness farther offshore [Andrews et al., 1990], as indicated by the expression of glacial landforms on the seabed surface [Bradwell et al., 2008b; Bjarnadóttir et al., 2013].

The drumlin examples from southwest Glasgow comprise only a very small sample population — similar high resolution geostatistical interpolations over a larger area would be time consuming and require adequate, well-distributed subsurface data sets. However, the results are of note because they illustrate real examples where true drumlin morphology probably differs from that indicated by modern DSMs. Approximately 170 km^2 of the drumlinised lower Clyde basin is mantled by postglacial raised marine deposits, suggesting that a much larger number of drumlin measurements there could be affected. Geological maps of the UK indicate several coastal and lowland areas where glacier bedforms and mapped sequences of postglacial sediments (e.g., raised marine deposits, glaciolacustrine deposits, glaciofluvial deposits) occur together, with the possibly that summary statistics of drumlin morphometry based on national data sets [Clark et al., 2009a; Spagnolo et al., 2012] could be influenced. Outside those areas, however, the morphometries of glacial bedforms are less likely to differ, as shown by the unaffected characteristics of drumlin 5 (Table 7.2), at the margin of the modelled postglacial infill, although other processes not considered here (e.g. colluvial activity) may still have affected drumlin shape.

A further point to note from the Glasgow example is the increase in overall land surface relief when postglacial sediments are removed. The present day *true* land surface area (based on a triangulated land surface at 25-m resolution) in the southwest Glasgow study area is 6.144 km^2. Removal of postglacial sediments results in an increased former glacier bed area of 6.162 km^2, highlighting the smoothing effect that postglacial sediments have on the landscape. While the increase in area is small in this example, it is a useful illustration of how parameters such as glacier bed roughness, which is linked to glacier velocity, can be affected by removal of postglacial sediments in palaeoglaciological studies.

This study has demonstrated how palaeoglaciological reconstructions and bedform morphometric analyses can vary if postglacial sediments are removed. The examples use high resolution surface analysis at local to regional scales. Whether accounting for postglacial sediment infill would significantly affect the results of larger scale (and perhaps lower resolution) studies is difficult to ascertain. High resolution DSMs are becoming increasingly available, and numerical ice sheet simulations can now be performed using anisotropic meshes with high spatial resolution at dynamic zones [Seddik et al., 2012]. Therefore, the influence of (and potential error caused by) postglacial sediments in palaeoglaciological analyses is likely to become more important. The British palaeo-ice sheet has been described as a 'conceptual playground for glaciologists' [Boulton, 2012]. In mainland and coastal Britain, large subsurface data sets exist ($>$ 600,000 borehole records), providing the potential to create surfaces that more closely represent former glacier bed topography. This raises the question: should the next generation of palaeo-ice sheet simulations that include lowland and coastal areas be performed using the modern land surface topography or interpolated glacier bed topography? If the former is chosen, some inherent error will exist.

7.5 Conclusions

- Borehole data sets were used to interpolate former glacier beds in three lowland and coastal areas of Scotland. These former glacier beds differ in elevation and relief from the modern land surface, which includes the topographic expression of postglacial sediments.

- In the lower Clyde basin, removal of postglacial sediments results in a deepening of the valley and reduction in valley shape factor. The effects on a reconstructed 27-km-long glacier profile are that the ice surface slope is steepened, surface elevations are increased by 0.7%, and glacier thickness is increased by 5%.

- In the Cromarty Firth, removal of postglacial shallow seabed sediments doubles the reconstructed cross-sectional area at a former calving glacier margin. Reconstructed width-averaged calving speed is increased by 74%, resulting in a 250% increase in the required steady-state ice flux.

- In the Glasgow area, removal of postglacial sediments results in an increase in measured drumlin length, width, and relief, causing an increase in drumlin volume of between 37% and 119%.

- The examples presented in this paper were chosen from lowland and coastal areas where thick postglacial sediments were known to be present. These environments form significant components of formerly glaciated terrains, and the examples presented here demonstrate how physical palaeoglaciological reconstructions and statistical analyses of glacier bedform morphometrics can be influenced by postglacial sediments. Thick sequences of postglacial sediments may provide one explanation for mismatches between simulated glaciers and empirical data in such areas. In some regions sufficient subsurface data sets exist over large areas to provide improved topographic representations of former glacier beds. These could be of great benefit to the next generation of palaeo-ice-sheet simulations.

Part IV

Synthesis

Chapter 8

Synthesis, further work and conclusions

8.1 Synthesis and further work

The five papers presented in this thesis sought to address the research questions posed in Section 1.2 (and listed below). The overall goal was to examine how parts of the last BIIS evolved through the last glacial cycle, and how these evolving ice masses interacted with the underlying landscape. In this section, key findings relating to each of the research questions are highlighted, together with suggestions about how the work may be taken forward.

8.1.1 What were the regional patterns of ice mass growth and decay, and what changes in ice mass organisation occurred during their evolution?

Key findings. In all three palaeoglaciogical settings (a mountain ice cap, an ice sheet core, and an ice sheet periphery), ice mass geometry and flow changed during the course of the glacial (stadial) cycle (Chapters 3, 4, and 5). At the scale of a small mountain ice cap, the locations of glacier inception were not necessarily the same as the main source areas during ice retreat (Chapter 3). This was probably at least partially related to catchment size and elevation. These local variations were masked at the ice sheet scale, where the western Scottish Highlands as a whole acted as the dominant source area during both ice sheet build up and retreat (Chapters 4 and 5).

The topography of the subglacial landscape was linked to the long term sensitivity / stability of parts of the BIIS. Relatively stable ice divides and zones of cold-based ice were associated with subglacial topographic highs, while ice divide migration and changes in basal thermal regime were focused through corridors of low relief subglacial topography (Chapters 4 and 5). The main east-west ice divide over the Clyde and Ayrshire lowlands migrated as much as 60 km (approximately 10% of the ice sheet width) through one such corridor, probably in response to enhanced phases of marine drawdown during the glacial cycle (Chapters 4 and 5).

The growth of the marine-based, Malin Shelf sector of the ice sheet had a profound effect on ice flow, drawing it westward – cross-cutting the well-established, geologically controlled, topographic corridors where earlier flow had been focused (Chapter 5). Upon retreat of the Malin Shelf sector, ice flow through the main fjords of south-west Scotland resumed. Based on previously published chronological information, and new ages reported here, average rates of marine margin retreat during deglaciation of the North Channel and Firth of Clyde were in the order of 10^2 m a^{-1} (Chapter 5). This represents a relatively rapid phase of ice sheet decay, exceeding overall average rates of ice margin retreat which were probably in the order of 10^1 m a^{-1} [Clark et al., 2012].

Future research An obvious extension of this research is the incorporation of numerical modelling in order to simulate the patterns of ice build up, flow and retreat, reconstructed here. The techniques are complementary: we will have more confidence in models that capture changes reconstructed from the geomorphological record (assuming they are correctly interpreted); and modelling can help highlight where our interpretations of the geomorphological record may be physically flawed. One of the challenges in future modelling experiments lies in simulating different processes at different scales. For example, the numerical model discussed in Chapter 3 successfully replicated geomorphologically reconstructed ice masses across Scotland during the Younger Dryas at a broad scale by implementing a strong west-east precipitation gradient. However, at the smaller scale of individual ice masses, wind redistribution of snow probably acted in the opposite direction, locally reversing this spatial trend in mass input (Chapter 3).

8.1.2 Were particular phases of ice mass evolution dominant in their effect on the landscape?

Key findings The landscape in each of the three settings considered in this thesis contains palimpsest landform and sediment records (Chapters 3, 4, 5). Instances exist in each case where landforms and sediments that pre-date, or formed early during the glacial cycle, were not subsequently destroyed. At the terrestrial to marine transition in south-western Scotland, the dominant erosional landscape features (breaches, overdeepened basins) were not found to relate to ice flow during the period of maximum ice sheet extent (Chapter 5). Rather, these features recorded glacier flow during repeated, restricted glaciations by a marine-proximal mountain ice sheet. This is thought to have been the dominant glacial mode in Britain and Ireland throughout the Quaternary.

Subglacial bedforms in unlithified sediment evolved by varying amounts during warm-based flow phases of the last glacial cycle (Chapter 4). In some cases the broad shapes (and internal structures) of initial bedforms were retained through the whole glacial cycle, despite being subjected to different subsequent flow regimes (Chapter 4). However, elsewhere the bedform morphology appears to almost entirely relate to warm-based flow during deglaciation (Chapter 4). Stratigraphic evidence confirms that bedforms in the central part of the Clyde basin relate to the last glacial cycle. Much of the sediment is likely to have been emplaced during initial ice sheet advance (Chapter 6), and progressively reshaped by localised sediment erosion and deposition during later stages of the ice sheet cycle. In this sense, much of the subglacial landscape may have been conditioned prior to, or during early stages of the glacial cycle. Subsequent re-shaping and streamlining was spatially and temporally variable, but generally focused under relatively fast-flowing outlets. Given the rapid rates of ice sheet retreat, the genesis of constructional ice marginal and submarginal landforms during deglaciation was probably restricted, due to reduced forward motion associated with glacier thinning and falling driving stresses. This has been demonstrated at a modern temperate, maritime glacial environment, where construction of ice marginal landforms has ceased during rapid ice front retreat Bradwell et al. [2013].

The key issue to highlight from this work is the relative importance of earlier glaciations and the build up phase of the last ice sheet in shaping, or conditioning, the overall glacial geomorphological signature that we see in the present landscape.

Future research Through the course of this research, I have often found myself asking how much of the glacial landscape we see today actually relates to the last ice sheet? We often assume in our analysis of the landform (and in particular subglacial bedform) record, that it documents flow regimes of the most recent ice sheet cycle. However, the research here has documented numerous other facets of the glacial landscape that are much older, and that landforms composed of unconsolidated sediment have survived from earlier deglaciations (Chapter 3). The well-constrained stratigraphy of the central Clyde basin confirms that at least some of the bedforms in the basin were wholly formed during the last glacial cycle. However, this level of stratigraphical investigation is rare, relative to the extent of geomorphological mapping in glaciated terrains across the world. The size of some subglacial bedforms (e.g. ribbed moraines in Ireland and mega-scale transverse bedforms in Canada [Clark and Meehan, 2001; Greenwood and Kleman, 2010]) makes the possibility of their survival through repeated glacial-interglacial-glacial cycles intriguing. In this sense, a greater understanding of bedform genesis could be gained by establishing a robust glacial sediment stratigraphy in many other areas [e.g. Ó Cofaigh et al., 2013].

8.1.3 How did the soft sediment ice sheet bed evolve in the Clyde basin during the last ice sheet cycle? What were the patterns and volumes of sediment moved, and how does this compare with proposed mechanisms of subglacial ice/sediment motion?

Key findings Much of the ~ 7 km^3 of 'till' in the Clyde basin was probably present prior to the last glacial cycle (Chapter 7). At the start of the glacial cycle, ice advanced into a landscape that had been partially conditioned by earlier deglacial and interglacial marine sedimentation. These soft sediments were glacitectonically deformed and mobilised upon glacier advance, with relatively short overall transport distances. Average rates of net till deposition at that time may have been up to ~ 0.04 m a^{-1}. Till deposition was focused over pre-existing permeable substrates, where it may have acted to seal the glacier bed, decrease effective pressures and facilitate basal sliding. Although the landscape was subsequently moulded by focused subglacial erosion and deposition at the ice bed, widespread, continuous bed deformation did not occur (Chapter 7). Upon deglaciation, sediment transport by meltwater formed a disproportionately high contribution to overall sediment movement in the basin (11% of the basin sediment volume was redistributed during the final 2.5% of the glacial cycle).

This thesis has documented: (i) preservation of pre Younger Dryas moraines under Younger Dryas ice masses (Chapter 3); (ii) preservation of pre-Devensian sediments under multiple phases of warm-based ice flow (Chapters 4, 5); and (iii) preservation, or partial preservation of subglacial bedforms that were most likely generated during early phases of the last glacial cycle (Chapter 4). These observations, combined with the evidence for limited inflow and outflow of till in the Clyde basin (Chapter 6), suggest that continuous, spatially pervasive shear of subglacial sediments may be rejected as a dominant mode of glacier ice motion in the locations considered in this thesis. Rather, a mosaic of shallow deforming spots and basal sliding is probably a more appropriate model under warm-based conditions. However, it is noted that this thesis did not incorporate analysis of the former ice sheet bed in an area of known ice streaming.

Future research In an assessment of the role of bed deformation in generating widespread Pleistocene till sheets, Alley [1991] noted that 'there is a clear need for quantitative studies to determine how much material was transported how far and how fast by the ice sheets'. Since then, however, relatively few datasets have become available to do this [e.g. Hooke and Elverhøi, 1996]. Chapter 6 has attempted to address this in the context of a lowland glaciated basin. Further development of this work would be to use the data presented here as a test for recently developed numerical simulations of ice advance into unconsolidated sediments [e.g. Leysinger Vieli and Gudmundsson, 2010].

8.1.4 What problems and uncertainties are associated with using the modern land surface to represent former ice sheet beds, and how can these problems be reduced?

Key findings Large parts of the last BIIS (in particular major outlet glaciers) occupied lowland and coastal environments. In certain lowland environments examined in this study, the modern land surface is shown to differ in elevation and relief from the former ice sheet bed – a result of postglacial deposition. These differences affect measurements of glacier bedform morphometry and can lead to errors in quantitative palaeoglaciological reconstructions (Chapter 7). The potential for errors introduced by unrepresentative subglacial topography may need to be considered in future numerical simulations of the BIIS. Alternatively, these errors could be reduced by the use of borehole datasets to interpolate former ice beds, as demonstrated in Chapter 7.

Future research This work could be applied over a much larger scale. Using national datasets, it would be possible to derive an improved interpolation of former ice sheet bed topography (and potentially other basic geotechnical properties) for many parts of the BIIS (e.g. Figure 8.1). This could significantly improve the input basal conditions for future modelling experiments. Additionally, this type of information could be beneficial for engineering purposes by providing an estimated depth to well-consolidated sediments, or in landscape evolution studies by approximating volumes of postglacially mobilised sediment.

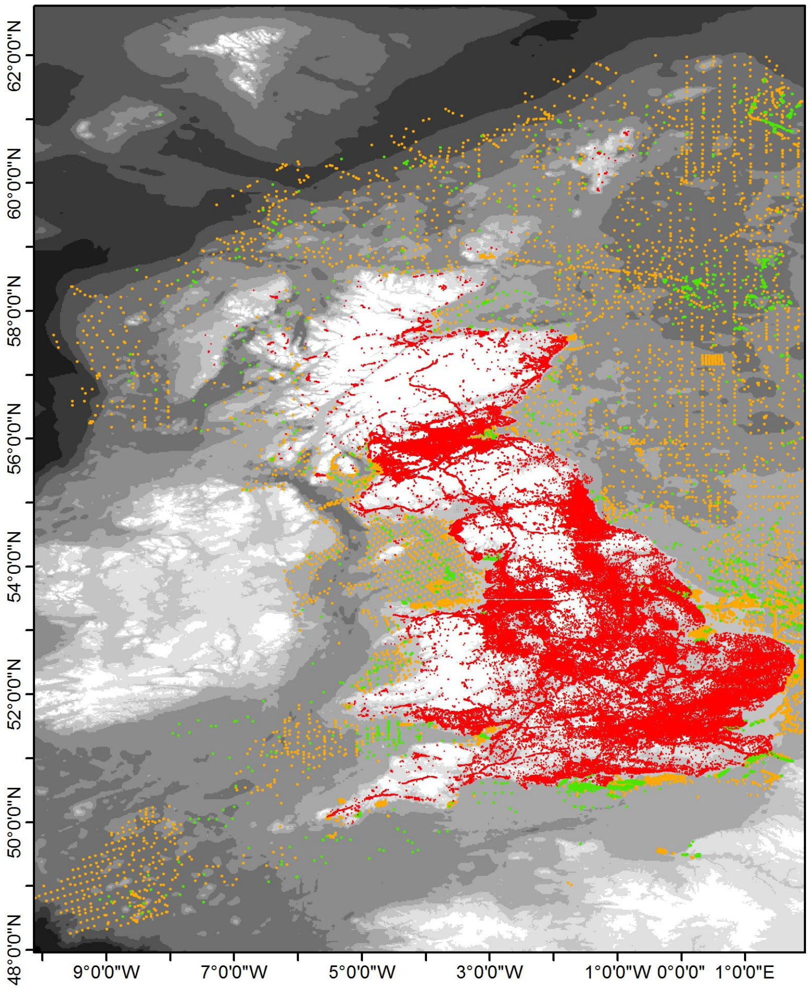

FIGURE 8.1: Subsurface borehole data that exists for Britain and adjacent offshore areas. Red - onshore borehole; green - offshore borehole; orange - offshore vibrocore. Note also that numerous seismic lines (not shown) also exist.

8.2 Thesis conclusions

The main objective of this thesis was to examine how parts of the last BIIS evolved and interacted with the underlying landscape during the last glacial cycle. The following conclusions may be drawn, regarding the areas studied.

- All parts of the former BIIS system that were examined in this thesis (a mountain ice cap, an ice sheet core area, and an ice sheet peripheral area) left a palimpsest landform and sediment record; and many examples exist where landforms or sediments deposited before, or during early stages of the last glacial cycle were not destroyed. By examining this composite geomorphological record, the build up, flow patterns, and decay of these ice masses have been reconstructed.

- Subglacial topography played a critical role in ice mass evolution. At the scale of an individual ice cap, the subglacial topography largely determined catchment size and relief, influencing spatial patterns of ice cap response to warming. Topographic highs under the ice sheet influenced the positions of stable ice divides, and some long-term frozen bed patches, while migration of ice divides and thermal boundaries was focused through subglacial topographic lows.
- The development of a marine-based ice sheet sector over the Malin Shelf significantly altered the flow regime of the BIIS in south-west Scotland, drawing ice flow at right angles to the main, geologically controlled, topographic corridors.
- The dominant geomorphological features in the coastal landscape of south-west Scotland do not relate to the period of maximum ice sheet extent. They were formed during an often-repeated, restricted, mountain ice sheet mode.
- Ice marginal and sub marginal processes (marginal glacitectonic deformation, meltwater transport) were key, and possibly the most important, agents of sediment movement in the Clyde basin during the last glacial cycle.
- In the areas examined, continuous, spatially pervasive shear of subglacial sediments was not a dominant mode of ice motion. Rather, wet-based basal motion was probably characterised by basal sliding and a mosaic of shallow deforming spots.
- In places affected by postglacial deposition, the former glacier bed can be interpolated from borehole data. There is an extremely large borehole dataset for Britain and the surrounding marine areas; this could be used to generate more representative ice-bed topographic grids for future BIIS numerical modelling experiments.

Collectively these findings contribute to our understanding of the last BIIS. The thesis demonstrates the important influence that the subglacial landscape had in the evolution of the BIIS, and highlights the need for continued geophysical surveys to map the detailed subglacial topography of modern ice sheets, in order to predict their long-term behaviour.

It is hoped that some of the findings from this research will go on to inform future reconstructions of the BIIS, in order to better constrain numerical models. Indeed, the research is timely, coinciding with the start of the NERC-funded BRITICE-CHRONO project, which is now attempting to make the last BIIS the best constrained former ice sheet anywhere, and a benchmark to improve and test predictive ice sheet models.

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