THE STYLE AND TIMING OF THE LAST DEGLACIATION OF WESTER ROSS, NORTHWEST SCOTLAND

A thesis submitted to the University of Manchester for the degree of

Doctor of Philosophy

in the Faculty of Engineering and Physical Sciences

2011

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THE STYLE AND TIMING OF THE LAST DEGLACIATION OF WESTER ROSS, NORTHWEST SCOTLAND

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Submitted for the degree of Doctor of Philosophy, 2011

Abstract:

The climate of the Wester Ross region of NW Scotland is particularly sensitive to fluctuations in the strength and latitude of the North Atlantic Gulf Stream. This was particularly apparent during the last deglaciation (14.7-12.9 ka), when overall climatic amelioration was interrupted by periods of cooling, the most significant being a 1.2 ka return to glacial conditions during the Younger Dryas (12.9-11.5 ka). Glacial readvances during these cooling episodes left behind numerous geomorphological features, which have been mapped and interpreted through a variety of methods, including fieldwork observations, aerial photography and digital elevation models, to form a detailed reconstruction of the style and timing of deglaciation. These methods were augmented by the study of 3D digital models, produced by combining 5cm resolution, Light Detection and Ranging (LiDAR) scans with colour photography, leading to the production of a detailed geomorphological map of a cirque formation in Torridon, Wester Ross, which was covered by an ice-sheet at the Last Glacial Maximum, and experienced localised ice flow during subsequent deglaciation and readvances.

Six statistically comparable cosmogenic $^{10}$Be bedrock exposure ages give a Younger Dryas age for sites in Torridon and Applecross (Wester Ross), and have also been used to constrain the vertical extent of these ice fields. Reconstructions of these ice bodies revealed that the Torridon ice field (mean ELA, 482m) covered ~100km$^2$, over twice the surface area covered by the Applecross ice field (~43km$^2$). This could have resulted from the survival of ice in Torridon prior to the onset of the Younger Dryas cooling, and is tentatively supported by pre-Younger Dryas cosmogenic $^{10}$Be exposure ages from this study and previous studies, which imply that ice existed close to the Wester Ross coastline and within central Torridon between 14-13ka. The Applecross ice field mean ELA (361m) was lowered by the presence of independent glaciers, which formed in low-lying troughs as snow was efficiently transferred to the NE by prevailing SW winds. Using empirical values from a global dataset, average annual Younger Dryas palaeoprecipitation values for the Torridon and Applecross ELAs are 2010 ± 266 and 2312 ± 534 mm a$^{-1}$ respectively, suggesting a wetter climate than today. Palaeoprecipitation calculated using equations based on a climate model of NW Scotland, yield lower values between 1005 ± 67 mm a$^{-1}$ and 1758 ± 118 mm a$^{-1}$ for the Torridon ELA and 1205 ± 233 mm a$^{-1}$ to 2109 ± 407 mm a$^{-1}$ for the Applecross ELA, perhaps a more reliable estimate which reflect enhanced continentality, promoted by the formation of sea ice on the NE Atlantic seaboard during the Younger Dryas.

Despite the rapid warming observed in palaeotemperature proxies, studies of glacial geomorphology and basal shear stress suggest that initial deglaciation was slow, oscillatory and warm-based, leading to the formation of prominent retreat moraines in the lower valleys. This prolonged transition can be related to the northward migration of sea ice and the gradual reintroduction of a Gulf Stream-dominated maritime climate. Ice remaining in the central area down-wasted in-situ as the regional ELA increased, creating hummocky landscape. Finally, cosmogenic $^{10}$Be exposure ages indicate that glaciers (probably characterised by a polythermal regime) retreated into the high north-facing corries at approximately 11.8ka, depositing a series of flutes.
**Declaration**

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Acknowledgements

I would like to thank my supervisors, Drs. Duncan Irving and Simon Brocklehurst, for their constant support as this fascinating project evolved. They have both given up many a weekend to troop up to NW Scotland with me in all weathers. Their insightful feedback has helped me to develop my scientific ideas. Big thanks to both of you for all of your time and effort.

Throughout my PhD I have also been lucky enough to have been guided by Dr. Giles Droop who kept me on-track as my PhD advisor, Dr. Phil Hughes who gave me some great feedback on my yearly progression and Dr. Derek Fabel, who patiently walked me through the world of cosmognics.

Fieldwork in the mist and rain of NW Scotland would not have been the same without some great companions. Many thanks to Frank Rarity who carried more than his body weight in LiDAR equipment and who patiently took me through the scanning process a step at a time. Massive thanks also to Annie Rowan, who played the “Schmidt Hammer game” so well...and to Rajasmita Goswami, whose unintentional one-liners will have me in stitches for years to come! I would also like to take the opportunity to thank the mechanics of Applecross for fixing my car on several occasions, and the midges, for making me work so much faster!

Throughout my time at the University of Manchester, I have had the pleasure of spending time with some great people, all of whom have made my experience here one that I will never forget, To: Drs. Claire Corkhill and Jon Martlew, for taking me under their wings when I first arrived here; Olly Duffy, for all the hijinks, brew-ha-ha-has and shenanigans; Dr. Dave Foster, for catching the greedy mouse and saving us from starvation; Myron Thomas, for trusting me with a rock drill and Dr. Julia Cartwright for our many chats. I would also like to thank Dr. Mike Lawson, Dr. Xavier Van Lanen, Jonathan Wood, Joss Smith, Alanna Juerges, Bridget Weston, Dr. Ivan Fabuel-Perez, Dr. Vicky Catterall, Amy Ellis, Laurent Petitpierre and Dr. Dina Vachtman, all of whom I have thoroughly enjoyed my time with. During my time in Manchester I have been fortunate to have had some great housemates. I would particularly like to mention Sarah Shelley for being a great friend, Laurence Caird for supplying chocolate in my hour of need, Nick Goodwin, for getting me away from Moss Side, Tom Lynch for walking an epic 55 miles with me and Lorraine Youds for making me laugh!

Through conferences and particularly through the organisation of the annual Quaternary Research Association postgraduate symposium, I made some great friends, namely Hannah Mathers and Rose Wilkinson. On a trip to Svalbard to look at Quaternary sedimentology, Lorna Linch, Johan Striberger and Andreas Nilsson were excellent company. Thanks to Olafur Ingolfsson for organising a trip of a lifetime, and to the polar bear for not eating us!

My parents have been a great support throughout this experience, particularly my mum, who took me in when I hit “The Wall.” She has been a fountain of wisdom and a source of encouragement throughout, as have my sisters, Katherine McCormack and Joanna Beckett.

Finally, I would like to thank my rock, my better half and my best friend, Tom Morgan. His constant patience, support and understanding have got me to where I am now. Thanks for making me laugh when I needed it the most! This one’s for you.
Chapter 1:

Introduction and background
1.1. Introduction

Mountain glaciers are some of the most sensitive and readily observable indicators of rapid climate change. The Intergovernmental Panel on Climate Change (IPCC) 2007 report on snow, ice and frozen ground (Alley et al., 2007) states that “Glaciers and ice caps provide among the most visible indications of the effects of climate change.” Research into the forces which drive glacial dynamics is therefore of key interest, not only to the academic community, but also to policy-makers and the general public. Rapid climate change is of particular importance in current affairs, and the study of past glacial climates can provide insight into glacier response to climate warming.

The 2007 IPCC report also states that “…some exceptional results indicate the complexity of both regional to local-scale climate and respective glacier regimes.” This is particularly pertinent to the situation in NW Scotland, where during the last glacial transition (~23 – 11.5 ka), the eastern N. Atlantic seaboard was particularly sensitive to fluctuations in climate, forced by variations in North Atlantic circulation, the principal control over glacial growth and retreat in this region (Lowe et al, 1994; Boessenkool et al, 2001; Weaver et al, 2003). The last glaciers in NW Scotland deposited a finely-tuned record of these fluctuations in the form of a unique suite of geomorphological features (e.g. Sissons, 1977; Bennett and Boulton, 1993a; 1993b; Benn, 1997; Wilson and Evans, 2000; Clark et al, 2004; Bradwell, 2005; 2006; Lukas, 2005; Benn and Lukas, 2006). This research aims to understand the processes which formed these features, provide a chronology for their formation and to ultimately place these findings in context with glacial dynamics and climate change.
Chapter 1. Introduction and Background

Figure 1.1. Field setting of Wester Ross, in the NW Scottish Highlands. The study areas of Torridon and Applecross are shown as insets. Contours are at 400 and 800 m. The base map is a NEXTMap Great Britain digital surface model, illuminated from the SW, at 10 m resolution.
Chapter 1. Introduction and Background

Figure 1.2. (Taken from inset in Figure 1.1). The place-names, mountains, glens, corries (cirques) and lochs of Torridon, taken from Ordnance Survey explorer © sheet 433. Contours are drawn every 100m. 1 km grid.

The study areas of Torridon and Applecross are located within Wester Ross, NW Scotland (Figure 1.1). Figures 1.2 and 1.3 provide an overview of the study areas. These maps are labelled with the names of the mountains, glens and lochs of Torridon and Applecross, which will be referred to regularly throughout this thesis. Where a single coordinate is needed for these areas (i.e. to calculate a difference in latitude), it is taken from a central location within Torridon or Applecross. Within Torridon, the coordinates are 9090E/6350N and correspond to the area named “Central region” on Figure 1.2. Within Applecross, the central coordinate 7875E/4470N corresponds to the breached area of the “Bealach nan Arr” (between Beinn Bhan and the “Knoll”).

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The Wester Ross region of NW Scotland provides a compelling setting for the study of past glacial climates and glacial dynamics because:

- The proximity of the region to the effects of the North Atlantic Gulf Stream, which fluctuates in strength between glacial and interglacial episodes according to the configuration of North Atlantic oceanic circulation (e.g. Ruddiman and McIntyre, 1977; 1981a; 1981b; Benn et al., 1992; McCabe and Clark, 1998).
• Previous studies of glacial geomorphology within Wester Ross still have a certain amount of controversy surrounding them (e.g. Sissons, 1977; Robinson; 1977; Robinson and Ballantyne, 1979; Ballantyne, 1986; Bennett and Boulton, 1993a; 1993b; Bennett, 1999; Jones, 1998; Ballantyne and Stone, 2009; Ballantyne et al., 2009). This research attempts to resolve some of these disparities through the employment of a variety of techniques, ranging from traditional (e.g. field mapping) through to modern (e.g. Terrestrial Light Detection and Ranging and cosmogenic \(^{10}\)Be isotope dating).

• The abundance of glacial geomorphological erosional and depositional features, the vast majority of which were produced during the most recent deglaciation in the British Isles (e.g. Ballantyne and Sutherland, 1987).

• The Wester Ross readvance moraine complex, which stretches from the south in Applecross to the regions north of Torridon, adds an extra dimension to the deglaciation chronology of the region. The presence of these moraines implies that ice was present in Wester Ross during an interstadial episode (14.7–12.9 ka). This has been corroborated by cosmogenic \(^{10}\)Be dating (Ballantyne et al., 2009; Ballantyne, 2010).

• During the last phase of deglaciation in the Loch Lomond Stadial (~12.9-11.5 ka), the main ice cap was centred over the NW Scottish Highlands, to the east of Wester Ross. Satellite ice fields that developed in Torridon and Applecross at this time (Robinson, 1977; Sissons, 1977) have left a detailed geomorphological record of the glacial response to climate change during deglaciation. This record can be used as a basis for the prediction of the glacial response under a variety of climate scenarios in the present day.
Chapter 1. Introduction and Background

1.2. Outline of thesis

This chapter provides an in-depth background to the study, as an introduction to the methods (chapters 2 and 4), results (chapters 3 and 5) and discussion (chapter 6). The aim of this thesis is to answer a number of questions regarding the style and timing of the last deglaciation in Wester Ross. These questions focus on the glacial geomorphology, chronology and reconstruction relating to the last ice bodies to reside in this location. They are stated at the end of this chapter (section 1.9), will be answered in the conclusions (chapter 7) and will be addressed throughout the intermediate chapters.

This chapter introduces the processes behind the magnitude and frequency of glacial events in the Northern Hemisphere. This includes an introduction to orbital (Milankovitch) theory, the roles played by tectonics and the oceanic circulation, a review of the evidence of millennial-scale change within the North Atlantic region, and the influence of the associated climate fluctuations on the glacial growth and dynamics, focussing on the British Isles and the study area in Wester Ross, NW Scotland. Previous studies of the glacial geomorphology of Wester Ross are covered from the onset of deglaciation, through readvance, to the final stages of deglaciation and the beginning of the current interglacial. To conclude this introductory chapter, the aims and objectives of this research are summarised in terms of four integral questions relating to mapping techniques, glacial geomorphology, timing and the glacial response to climate change during the last deglaciation in Wester Ross.

Chapter 2 encompasses the geomorphological and geochronological aspects of this study and covers the variety of methods employed to study the glacial landscape of Wester Ross. Geomorphological fieldwork techniques including research strategies, the identification of glacial geomorphological features and the collection and analysis of field evidence are outlined. The use of remote sensing techniques such as aerial photography and Digital Elevation Modelling (DEMs) are introduced and an overview of the processes involved in the compilation of mapping data into a Geographical Information System (GIS) is
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given. Ground-based Light Detection And Ranging (LiDAR) is introduced, and the processes involved in the collection of data, the processing of scans and the creation of a 3D digital model are described. The geochronological aspect of this chapter covers the considerations involved in collecting samples for cosmogenic $^{10}$Be analysis, including Schmidt hammer analysis and the justifications of the choice of sampling locations. The procedures involved in isolating the $^{10}$Be fraction from quartz are outlined in detail, as is the process of calculating an exposure age from the $^{10}$Be concentration. Chapter 3 provides the detailed results of the geomorphological and geochronological aspects of the study in preparation for discussion in chapter 6.

Chapter 4 outlines the methods involved in the reconstruction of past glacial climates, covering the relationships between glacial dynamics, climate change and the local equilibrium line altitude. Methods used for the reconstruction of the glacier limits, surface morphology and equilibrium line altitude of the Applecross and Torridon glaciers are provided and are integrated into equations used to calculate palaeo-temperature and palaeo-precipitation of individual glaciers and ice bodies. The steps taken to calculate basal shear stress for a variety of longitudinal glacial profiles are outlined in this chapter. A variety of glacier and palaeoclimate reconstruction methods are used and will be further assessed in the discussion (chapter 6). The detailed results of this reconstruction work will be presented in chapter 5.

Following the sequence of questions outlined in section 1.9, the discussion will firstly assesses the variety of methods used as tools for mapping the glacial geomorphology of Wester Ross (section 6.1-6.3). To allow for fair comparison, Coire Mhic Fhearchair in central Torridon (Figure 1.2) has been chosen as the location to assess the suitability of field mapping, aerial photographs, Digital Elevation Models (DEMs) and ground-based LiDAR surveys as mapping tools. The second part of the discussion (section 6.2) will relate to the glacial geomorphological features (erosional and depositional) of Wester Ross. The implications of these features for former ice dynamics, thermal regimes, ice extents and the interaction of the former ice bodies with their surroundings will
also be assessed. Focus will be placed on an area of Torridon, where a wide array of glacial geomorphological features can be observed over a ~19 km² area, with implications for LLS deglaciation in this ice-marginal region of Wester Ross. Section 6.3 will focus on the geochronological reconstruction of the deglaciation of the Applecross and Torridon areas of Wester Ross. The results of \(^{10}\)Be cosmogenic dating will be analysed in conjunction with the geomorphological observations from section 6.2, and the implications for the timing of ice retreat will be assessed in context with North Atlantic and Scottish climate proxies and models. Finally, section 6.4 addresses the implications of glacier reconstructions for the climate of Wester Ross during the Younger Dryas cooling episode. Glacier characteristics will be assessed on individual glacier, ice field and regional scales, and in terms of their morphology and basal conditions. This information can subsequently be linked back to section 6.2 to investigate the impact that these glaciers had on the post-glacial landscape of Wester Ross. The variety of ELA and palaeoclimate reconstruction methods introduced in chapter 4 will be scrutinised and the results of these calculations will be analysed in terms of their implications for local and regional ice dynamics. At the end of this chapter (section 6.5), a discussion of the wider implications of this research is presented, placing palaeoclimatic change and the associated glacial fluctuations during the last deglaciation in context with contemporary climate change. Finally, (section 6.6) will outline the wider implications of the results of this research and section 6.7 will highlight some of the outstanding unanswered questions and speculate on the potential for future research.

The conclusions chapter (chapter 7) addresses the integral questions posed at the end of this introductory chapter. The outcomes of this research will be outlined in terms of the implications of glacial geomorphological observations, geochronological estimates and glacial and climate reconstructions for the style and timing of the last deglaciation in Wester Ross.
1.3. Quaternary glaciation of the Northern Hemisphere

Tillites, dropstones and diamicites in the Palaeoproterozoic geological record all point to evidence of glaciation as long ago as 2.4 – 2.3 Ga (Sukumaran, 2003). Subsequent glaciations occurred as continents were repositioned (Hoffman and Schrag, 2002; Meert, 2003); mountain chains were uplifted and atmospheric CO$_2$ was drawn down and buried (Raymo, 1994); the orbital configurations of the planets varied (Zachos et al, 2001; Pälike et al., 2006) and ocean currents fluctuated (Davies et al., 2001). This chapter focuses on the causes and onset of a series of glaciations in the Northern Hemisphere, following the formation of a closed North Atlantic basin (2.6 Ma to 11.5 ka). Those who first realised the extent and causes of this series of glaciations will be introduced, as will the characteristics of the most recent glaciation on the NE Atlantic seaboard and the impact it has had in shaping our landscape.

The term ‘Pleistocene’ (1.806 Ma to 11.5 ka) was originally intended to cover the latest period of glacial cycles, although subsequent to this, earlier cooling events were discovered to have occurred in the late Pliocene. The ‘Quaternary’ was introduced to cover this extension of the Pleistocene, and extends back to 2.6 Ma (A Quaternary chronological breakdown can be seen in Appendix 1). In 2009, the International Commission on Stratigraphy agreed to lower the base of the Pleistocene to that of the Quaternary, and endorsed the usage of both Quaternary and Pleistocene to cover the period 2.6 Ma – present (Gibbard et al., 2010).

1.3.1. External forcing

At a time when scientific understanding of the polar ice caps was limited, observations of glacier movement in the Alps prompted Louis Agassiz (1840) to propose the theory that the whole of the Northern Hemisphere had been glaciated in the past. Agassiz also popularised the idea of a geologically recent ice age as the latest catastrophic event in the history of the Earth, in an age when James Hutton’s idea of uniformitarianism was widely accepted (Cannon, 1960).
Further investigation was required to explain the growth and decay of continental ice sheets and the associated complexities of earth-ocean-atmosphere feedback. The French mathematician, Joseph Adhemar (1842), recognized that the Earth’s orbital ellipse varies from year to year due to the gravitational pull associated with the sun, moon and other planets. The combined effect of these planetary bodies causes the axis of the Earth to wobble under the torque of gravity, controlling insolation patterns and therefore the potential for ice growth in favourable locations. Scottish physicist James Croll published “Climate and Time” (1875), which predicted that either hemisphere would experience an ice age whenever two conditions occur simultaneously: a relatively elongate orbit and an enhanced winter solstice. However, Milutin Milankovitch (1920) recognised that under these conditions, hemispheric glaciations would not be synchronous. Geological observations suggested that global synchronicity is a key feature of glaciations and that summer (not winter) insolation patterns play a key role in glacial growth and decay.

Milankovitch recognized that two-thirds of the Earth’s terrestrial cover is located in the Northern Hemisphere and that the insolation that controls terrestrial ice build up is the driving force behind a globally synchronous ice signal. He demonstrated that insolation is dominated by a 23 ka cycle and less-so by 41 and 100 ka cycles, and predicted the longevity of ice ages according to the incidence and intensity of insolation (Muller and MacDonald, 2000). Since his calculations were published (Milankovitch, 1941), the orbital parameters have been recalculated several times in order to account for the way in which changes in the Earth’s orbit affect temporal and spatial insolation patterns. A seminal paper by Hays et al., (1976), uses deep-sea sediment records to investigate the variance of the orbital cycles, which can be seen in figure 1.4. The 100 ka eccentricity cycle (Figure 1.4) is a measurement of the departure of the Earth’s orbit from that of a perfect circle. Slight variation (eccentricity) in the Earth’s orbit is caused by a torque effect of other planets (mainly due to the size of Jupiter and the proximity of Venus). This eccentric behaviour leads to contrasts in the seasonality of the hemispheres.
Chapter 1. Introduction and Background

Figure 1.4. The Milankovitch orbital cycles. Adapted from Zachos et al., (2001).

*Obliquity* is the oscillation of the tilt of the Earth's poles, in relation to the sun. The angle of tilt varies between 21.39° and 224.36° over a 41 ka cycle (Siegert, 2001). If obliquity was zero, the sun would pass overhead at the equator every day, only half of the sun would constantly be visible at the poles and seasonality would be much weaker. The ~23 ka *precession* parameter determines how close the Earth is to the sun in a Northern Hemisphere summer. The gradual change in direction of the North Pole as the moon and sun exert a torque on the equatorial bulge of the earth, means that the date when the Earth is closest to the sun (perihelion) varies on a seasonal basis over time. We are currently experiencing an interglacial due to the proximity of the Northern Hemisphere to the sun during the summertime, meaning that ice cover cannot be maintained throughout the year.

Since the discovery of the sedimentary record of climatic change, the establishment of the Deep Sea Drilling Project (DSDP) in 1968 has revolutionized the high-resolution study of palaeoclimates, and in particular the frequency and magnitude of the Milankovitch cycles. It became the Ocean Drilling Program (ODP) in 1983, and in October 2003 it evolved into the Integrated Ocean Drilling Program (IODP). The availability of high resolution and temporally extensive records as a result of these projects has led to an increase in detailed studies of the Pleistocene glacial record.
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Figure 1.5. The mid-Pleistocene transition: $\delta^{18}$O signal from DSDP core 607. Adapted from Ashkenazy and Tziperman, (2004).

The analysis of deep-sea sediment cores has answered a number of important questions relating to orbital theory and the complexities of the internal climate system, but has also highlighted a number of problems. Milankovitch’s insolation theory suggested that the 23 ka (precession) signal is the strongest, and the weakest is the 100 ka (eccentricity) signal (Muller and MacDonald, 2000). Therefore the dominance of the 100 ka cycle in the $\delta^{18}$O record was not an expected outcome from studies of deep-sea cores (Broecker and Denton, 1990; Hays et al., 1976). Studies by Broecker and Van Donk, (1970); Hays et al., (1976); Imbrie and Imbrie, (1980); Imbrie et al., (1984); Raymo, (1997) and Huybers, (2007) indicate that as Pleistocene ice expanded, a gradual insensitivity to the precession and obliquity cycles resulted, and the eccentricity (100 ka) cycle became dominant following the “Mid Pleistocene transition” (1.2-0.8 Ma; Figure 1.5). Since the mid-Pleistocene transition, the Northern Hemisphere has experienced major glacial cycles (stadials) every 100 ka, interrupted by major interstadials ~10 ka long. The last deglaciation took place over 10 ka ago, suggesting that we are due to enter another glacial period imminently.

Despite the current dominance of the eccentricity cycle, the three distinct orbital cycles are superimposed upon each other, and combine to yield a gross insolation anomaly, which is fairly constant over time on a global scale, but is unevenly dispersed over latitude, depending on orbital configuration. The orbital parameters oscillate at a steady rate over millions of years, and can explain the
regularity of climate change over time, but external forces alone cannot explain
the large amplitude of these variations. These changes must be subject to internal
feedback mechanisms, which exaggerate and complicate the signal observed in
climate proxies throughout the world. Throughout geological time, tectonic
reorganisation has defined global boundary conditions, which include the
geographic and topographic characteristics of the continents, the location of
oceanic gateways and circulation patterns, sea-floor bathymetry, greenhouse gas
concentrations and ultimately the extent of the ice sheets. The interaction of
these factors means that climatic change can be extreme, rapid, complex and
therefore unpredictable.

1.3.2. Internal feedback

The closure of the Isthmus of Panama during the late Cenozoic had a profound
effect on the circulation and climate of the North Atlantic (Haug and
Tiedemann, 1998; Bartoli et al., 2006). Closure lasted from 13 to 1.9 Ma as
indicated by the evidence from sediment cores on the Atlantic and Pacific sides
of the Isthmus. Significant changes in the $\delta^{18}O$ salinity signal of planktonic
foraminifera assemblages occur at 4.6 Ma implying a shallowing of the Central
American Seaway and the presence of shallow water species at 3.6 Ma indicates
that closure was nearly complete at this time. The timing of the final closure of
the Central American Seaway remains controversial, as does the relationship
between this event, the onset of Northern Hemisphere Glaciation (NHG) and
the physical parameters controlling it (Bartoli et al., 2006). Despite a cooling
trend throughout the Cenozoic, the climate failed to generate and maintain a
major ice age until 2.7 Ma when the optimum ice growth conditions were
achieved.
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Figure 1.6. The formation of the North Atlantic Deep Water (NADW) in the Arctic Ocean. Adapted from Tomczak and Godfrey (1994)

The three requirements essential for the formation and preservation of Northern Hemisphere glaciers include: reduced insolation during Northern Hemisphere summers; a cooler climate and an increased supply of moisture to the high latitudes. A reconfiguration of oceanic circulation patterns provided the conditions necessary for the establishment of massive ice sheets in the Northern Hemisphere. The closure of the Isthmus of Panama coincided with the formation of a closed gyre system in the North Atlantic, forming a Pacific-Atlantic divide, and increasing the efficiency of the North Atlantic oceanic heat conveyor.

The deepest and densest waters of the North Atlantic Deep Water (NADW) form in the Arctic Ocean and Nordic seas (Figure 1.6), where evaporation of the warm, fresh Atlantic inflow waters promotes the sinking of highly saline water (Imbrie et al., 1992). Atlantic inflow is drawn from both the sub-polar gyre and the sub-tropical gyre (Figure 1.7) and enters the Arctic Ocean through the Rockall Trough.
Figure 1.7. The north Atlantic conveyor. Thickness of line and arrowheads denote the relative depth of the current (i.e. thicker lines represent deep water currents). Colour represents temperature with dark blue at the cold end of the spectrum and red at the warm end. Light blue represents an intermediate temperature.

The sub-polar gyre directs the warm, saline waters of the Tropical Atlantic towards the NE Atlantic (e.g. Hátún et al., 2005). Evaporation of warm tropical water and incorporation of freshwater in sea ice promotes the formation of cold, dense, saline water which sinks to ~3000 m. The NADW then flows towards the SW Atlantic and the southern hemisphere, where it overrides the Antarctic Bottom Water in the southern oceans at the Antarctic convergence zone (~60ºS) and rises back to the surface.
Figure 1.8. North Atlantic temperature and current profile. Vertical arrows show where heat enters and leaves the surface waters and the horizontal white lines indicate net transport. The inset shows the location of the cross section. Adapted from Hegerl and Bindoff, 2005.

Together with sea-ice melt from the Antarctic and warm surface waters, the NADW forms the Antarctic Intermediate Water (AAIW), which flows northwards and is slowly modified to a less dense and warmer water body in the tropics. In the Gulf of Mexico, surface waters are warmed and carried towards the northern latitudes of Europe as the warm, moisture-laden Gulf Stream. The system can be likened to a thermal conveyor, warming the high latitudes (including the Wester Ross region of NW Scotland) and cooling the tropics. Hegerl and Bindoff, (2005; Figure 1.8) demonstrate how heat is transferred to the deep ocean from the surface layer, along surfaces of equal density, via wind-driven and thermohaline circulation. Deeper oceanic overturn and horizontal heat transport are indicated, and show the pattern of circulation throughout the Atlantic Ocean. The conveyor is shown to work efficiently as net heat gain is focused around the tropics, whereas heat loss is most prominent around the polar and temperate regions.

An interesting theory regarding the onset of NHG (Driscoll and Haug, 1998; Bartoli et al., 2005; Sarnthein et al., 2009) relies on a combination of ocean-atmosphere feedback loops and external forcing, which eventually lead to the shutdown of the conveyor.
While the Milankovitch variations in precession, obliquity and eccentricity play a part in the pacing of glacial cycles, it seems that long-term cooling between 2.9 – 2.4 Ma had a terrestrial cause (Raymo, 1994). The increased delivery of moisture to the high latitudes (due to increased evaporation from warmer surface waters in the North Atlantic) via prevailing westerlies (Figure 1.9) would have increased fresh water discharge in the Arctic via enhanced river discharge in Eurasia, resulting in the freshening of the Arctic Ocean (Driscoll and Haug, 1998).

Figure 1.9. A possible short-circuit in the North Atlantic conveyor, through freshening of the North Atlantic Ocean. An increase in atmospheric circulation and evaporation of warm water over the North Atlantic leads to the transportation of moisture to Eurasia via westerly air streams. A consequent increase in freshwater input into the Arctic Ocean leads to a decrease in salinity. A concurrent decrease in water density restricts the formation of the NADW. Figure adapted from Driscoll and Haug, (1998).
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Figure 1.10. The onset of glaciation in the Northern Hemisphere and the weakening of the North Atlantic conveyor. As the NADW weakens or is completely shut-down, the North Atlantic circulation gyre alters, and the Gulf Stream is deflected towards the Tropics, as sea ice expands. A climatic cooling in the Northern Hemisphere promotes the growth of terrestrial ice caps. This scenario is an approximation of the ice extent at the Last Glacial Maximum (LGM). Adapted from Steigert, 2001.

The volume of water discharged from these rivers (10% of the global river discharge) would have had a profound effect on the surface waters of the Norwegian-Greenland Seas, promoting sea ice formation through a decrease in salinity, increasing albedo, preventing NADW formation, slowing or preventing the formation of the Gulf Stream and thus forming a positive feedback loop to promote ice growth (Figure 1.10). The consequent dampening of the North Atlantic circulation gyre and the Gulf Stream would have lowered the critical temperature at which precipitation falls as snow. An increase in obliquity (reduced tilt) around 3.1 Ma would have led to a reduction in insolation in the
high latitudes during the summer months, meaning that ice could have been preserved throughout the year, extending its influence further south during the winter months.

The fluctuation and interaction of orbital parameters and internal feedback mechanisms over long (>10 ka) timescales affects the extent of ice cover in the Northern Hemisphere. Superimposed on this signal are a number of even shorter-term fluctuations, which were particularly important to climate change and the dynamics of ice bodies in NW Scotland during the last deglaciation. These climate fluctuations will be the focus of the proceeding section.
1.4. High-resolution proxies millennial and sub-millennial-scale change

Orbital cycles are thought to set the pace and modulate a climate system, which oscillates independently around a steady state (Broecker and Denton, 1990). The increase in non-linear behaviour can only be explained by the internal dynamics of the climate, and is supported by the appearance of millennial (and sub-millennial) scale variations in the atmospheric (ice core) and oceanic (sediment core) records from the Northern Hemisphere. Millennial scale change has been observed in climate proxies globally, throughout the Pleistocene and beyond.

1.4.1. The ice-core record

Ice cores drilled in Greenland provide high-resolution (annual) palaeoclimatic data, providing insights into the past ocean-atmosphere interactions in the North Atlantic region. Stable isotopes trapped in oxygen bubbles within cores, in particular the heavy $^{18}$O isotope and its deviation from the standard mean ocean water (SMOW) concentration ($\delta^{18}$O), provides valuable palaeo-temperature estimates (Clapperton, 1997). A chronology can be established through the counting of annual accumulation layers and the correlation of volcanic ash (tephra) layers (Mortensen et al., 2005; Davies et al., 2008). The European GReenland Ice core Project (GRIP) and American Greenland Ice Sheet Project (GISP) cores were drilled at the Greenland summit to ~3000 m depth, covering late Quaternary glaciation, with reliable records extending back to 105 ka. However, the lowermost 10% of these cores are unreliable due to basal ice folding and the distortion of the chronology of the Eemian interstadial (Appendix 1) (Andersen et al., 2004). A more reliable record was discovered through echo-sounding (Dahl-Jensen et al., 1997) and Ground Penetrating Radar (GPR) (Nixdorf and Göktas, 2001) studies of the Greenland ice cap. These studies led to the establishment of the European North GReenland Ice core Project (NGRIP). Drilling took place in a zone of minimal basal distortion (Nixdorf and Göktas, 2001) with the aim of extending the record back to the Eemian interstadial (~123 ka).
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Figure 1.11. Oxygen Isotope stages identified from $\delta^{18}O$ signal for benthic foraminifera species Pyrgo and planktonic species Globulina Bulloides. Percentage of IRD and foraminifera from Atlantic core Me 69-17 (Lat: 47 ºN., long 19 ºW, and depth 3.9 km). Ash layers 1 and 2 (black dashed line) provide a chronological framework. Heinrich events are indicated by the red dashed line. Adapted from Broecker et al., 1992; original records from Heinrich (1988).

Figure 1.12. The GISP2 climate record for the last glacial cycle. Dansgaard-Oeschger warming events are divided by lines below the curve, and are spaced 1,470 years apart. Heinrich event are labelled 1 to 12 above the curve. Adapted from Rahmstorf, 2003.

In contrast to the erratic data from the GRIP and GISP cores, $\delta^{18}O$ records from the NGRIP core suggest that the Eemian was roughly as stable as the Holocene, with temperatures ~5ºC warmer than today (Andersen et al., 2004). The NGRIP $\delta^{18}O$ (temperature) record is referred to throughout this study as a basis for the comparison of glacial change in Wester Ross with widespread climate change in the North Atlantic basin.
1.4.2. Heinrich events

The North Atlantic sedimentation record is an ideal proxy for millennial scale climate change, due to slow bottom-water currents and high sediment accumulation rates, which are favourable to preservation. “Barren Zones,” are 1cm-1m thick layers, rich in Ice Rafted Debris (IRD), abundant in the polar foraminifera, N. pachyderma, with low δ\(^{18}\)O signatures and have been identified stretching across the Atlantic for ~3000km. These barren zones represent cycles, superimposed on Milankovitch cycles, and are best defined during a glacial period as “Heinrich” and “Inter-Heinrich” events. Heinrich (1988) discovered 6 of these layers, which were deposited in the North Atlantic between 60 and 10 Ka at 47°N, 19°W and provided an explanation for the cooling events observed in the ice records. Broecker et al. (1992), matched 5 of the 6 Heinrich layers using the ratio of IRD:Foraminifera, which is noted to be higher during Heinrich events (Figure 1.11). Broecker’s core (ODP 609) was located at 50°N and 24°W (3°N and 5°W of Heinrich’s core) but very similar peak heights and widths were recorded, suggesting that ice-rafting was a basin-wide phenomenon.

Heinrich layers are the result of catastrophic melting events, and the subsequent transport of icebergs (and therefore freshwater) into the North Atlantic Ocean. As discussed previously, the introduction of vast amounts of freshwater into the North Atlantic can have dramatic consequences for the climate in adjacent areas, including NW Scotland. However, ice calving would only result in the observed magnitude of sedimentation if it were synchronous throughout the North Atlantic. The origin of the events can be identified through the lithological identification of the IRD source regions in Greenland (mainly the Laurentide ice sheet), Iceland and Scandinavia (Dowedeswell et al., 1999). Synchronous calving would imply a large-scale Heinrich event and a potentially large-scale climatic event.
1.4.3. Dansgaard-Oeschger events (D-O events)

The correlation of Heinrich events with sudden peaks in air temperature over Greenland (Dansgaard-Oeschger (D-O) events) was first noted by Bond and Lotti (1995). D-O events occur quasi-periodically and are characterised by a rapid warming, followed by a prolonged cooling (Bond et al., 1999) (Figure 1.12). Rahmstorf (2003) observed D-O events occur in a 1,470 year cycle, but this cyclicity is only characteristic of the past 50 ka, and is therefore unlikely to be related to the regular variations of the orbital cycles.

Maslin et al., (2001), suggest that the cycles associated with D-O events are determined by the stability of the ice sheet with which they are associated. Once the northern hemisphere ice sheets reach a threshold, freshwater enters the ocean system. A subsequent decrease in density-related sinking causes a weakening of NADW, North Atlantic circulation and an increase in heat transfer towards the South Pole, causing melting of the Antarctic ice sheet and reducing the strength of the AABW. The NADW then returns to its previous strength, and the cycle is reset. This phenomenon is supported by the asynchrony of Greenland and Antarctic D-O records suggesting that when the NADW is weakened, the Antarctic Bottom Water (AABW) can penetrate further north, thus triggering an inter-hemispheric climate oscillation, or a “bi-polar see-saw” (Seidov and Maslin, 2001).

1.4.4. The North Atlantic Oscillation (NAO)

In the North Atlantic, high-pressure (anticyclone) systems are centred over the subtropical Atlantic, while low pressures (cyclones) are centred over Iceland. The normalized sea surface pressure difference between these two systems during a North Atlantic winter determines the atmospheric conditions on a multi-annual timescale, thus defining the NAO index (Figure 1.13) and associated sub-millennial-scale oscillations. A positive index phase is defined by a large pressure gradient, which enhances the strength of the westerlies, storm intensity, precipitation and temperatures on the NE Atlantic seaboard.
A negative phase is characterised by a weakening of both pressure systems, and dry, stable conditions (Six et al., 2001; Bojariu and Gimeno, 2003). These changes can also affect the dynamics of North Atlantic circulation and the horizontal flow of the surface currents, including the Gulf Stream (Taylor and Stephens, 1998; Hurrell et al., 2001).

A record of the NAO can be observed in many proxies, including net ice accumulation in ice cores (Appenzeller et al., 1998) and $\delta^{18}O$ records of foraminifera in sediment cores (Brückner and Mackensen, 2006). The relationship between the NAO and glacier mass balance records has also been studied (Nesje et al., 2005) as an indication of precipitation patterns in Europe. Regional precipitation anomalies have been observed on glaciers in Alpine and Scandinavian locations. A positive correlation between Scandinavian glacier mass balance and the NAO (as opposed to a negative Alpine correlation), could
highlight the importance of a maritime setting to the geographical distribution of glaciers (Six et al., 2001).

The maritime setting of Wester Ross on the NE Atlantic seaboard could have affected local glacial response to such North Atlantic phenomenon as the NAO and D-O cycles. The effects of these millennial and sub-millennial oscillations on the glacial response in Wester Ross will be assessed in the discussion (chapter 6).

1.4.5. The Gulf Stream

The formation and strength of the Gulf Stream is highly dependent on the configuration of North Atlantic circulation. Since the formation of the conveyor, it has played a critical role in millennial scale climate change throughout the North Atlantic. A number of theories link variations in oceanic circulation to the quantity of glacial ice in the high latitudes, in particular the NE Atlantic seaboard, which is heavily influenced by the Gulf Stream and its associated warm air masses. The conveyor can oscillate between “on” and “off” modes throughout glacial cycles. When the conveyor is switched off, glacial conditions predominate as the Gulf Stream weakens (see Figures 1.9 and 1.10). When the conveyor is strengthened, the ice sheets of the high latitudes respond by melting. These Heinrich events are characterised by the formation of icebergs, which drift into the open ocean, depositing Ice Rafted Debris (IRD) and forming Heinrich layers in sediments. This influx of fresher water creates a perturbation in the system, dampening the conveyor once more and promoting ice growth. The “flickering” of this system may be able to explain the rapid climatic warming, which characterises D-O events.

Climate fluctuations in the west of Scotland are intimately linked with the location of the Oceanic Polar Front (Ruddiman, 1977) and the Gulf Stream. The position of the Oceanic Polar Front is established by the meeting of the warm, saline Gulf Stream and the cold, fresh currents from the Arctic Ocean. The position of this front is therefore dependent on how far north the Gulf Stream extends at a given point in time.
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Figure 1.14. Average annual precipitation and temperature from Oban, NW Scotland and Nain, E Newfoundland. Data taken from Met Office records (1931-1988).

Figure 1.15. The weather in the Wester Ross region can be highly changeable over time and within a relatively small area. The first picture was taken in late April in Torridon, whilst the second was taken in Gairloch, ~30km from Torridon in early May of the same year.
The Gulf Stream currently extends far north in the North Atlantic and the Oceanic Polar Front is located far to the north of the British Isles, even in the winter. The westerly flow of the Gulf Stream creates massive contrasts in the respective local climates of the east and western Atlantic seabords. Figure 1.14 shows the average annual conditions for Dunstaffnage (Oban) at 56.56°N/5.54°W on the west coast of Scotland and Nain, Newfoundland (56.56°N/61.68°W). Both of these locations are at the same latitude, on opposite coastlines of the North Atlantic Ocean. However, average rainfall and temperature (1931-1988) are considerably different. In Nain (Figure 1.14), the prevailing winds come from the west (offshore), bringing cold air from the continent, and leading to its extreme climate.

The dominant westerly movement of warm air masses over the North Atlantic provides Scotland with abundant precipitation, which tends to fall as snow in the mountains between November and April (Benn, 1997). The Gulf Stream currently promotes a mild, maritime climate in NW Scotland, where January sea surface temperatures can reach 4°C. This makes the local weather patterns somewhat unpredictable, varying from cloudless skies and highs of 20 °C plus one month, to squally snow showers the next (Figure 1.15). A combination of the dominance of the onshore SW winds in NW Scotland and the ability of the oceans to retain heat maintains this milder climate.
1.5. Glacial response to local climate and topography

Glacial dynamics are reliant on a number of factors, including local climate and surrounding topography. This section outlines how glaciers grow and decay (1.5.1), their response to changes in local climate (1.5.2) and topography (1.5.3) and the internal response of a glacier as exhibited by basal thermal regime (1.5.4). This section provides an introduction to the proceeding sections which focus on past glacial response in Wester Ross.

1.5.1. Ice-body dynamics

Knowledge of glacier mass balance is essential to the understanding of how ice bodies (on all scales) function and is therefore integral to the reconstruction of glaciers (Chapters 4 and 5). The inputs (accumulation) to the glacial system include direct precipitation (snow, hail, rainfall) onto the glacier surface, and indirect transference of wind-blown or avalanched snow. Ablation processes depend on the glacier type and can range from direct ice melt on terrestrial margins, to ice calving on marine margins (Glasser, In Gordon et al., 1997). Mass balance (Figure 1.16) is concerned with the overall balance between accumulation and ablation throughout an ice body (Bennett and Glasser, 1996).

In terrestrial glaciers, accumulation is usually the dominant process in the cold winter months, and is generally balanced by ablation in the warm summer months. In marine terminating glaciers, ice calving is a constant ablation process, but is enhanced in the summer when basal sliding is increased. Deviations from this net mass balance can lead to the growth or decay of an ice body. If accumulation exceeds ablation (positive mass balance) in successive balance years, an ice body will expand or readvance. When ablation exceeds accumulation (negative mass balance) over successive years, deglaciation can occur. Both accumulation and ablation are heavily dependent on local climate and altitude, and vary from year to year.
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Figure 1.16. Accumulation and ablation over a mass balance year. Mass balance varies according to location and climate, and can be used to infer past climates where glacial reconstructions are possible.

Figure 1.17. The theoretical position of an ELA within a corrie glacier (similar principles apply for a valley glacier). The surface ELA generally corresponds to the zone of maximum erosion on the valley floor, and divides the accumulation and ablation areas. Flow lines within the glacier indicate ice movement. Arrow length is indicative of velocity, which increases towards the ELA and decreases towards the glacier snout.
Accumulation tends to dominate in the upper reaches on an ice body where temperatures are lower and precipitation is high. Ablation generally occurs at the terminus where temperatures are higher and precipitation is lower. The Equilibrium Line Altitude (ELA) defines the point on a glacier at which accumulation and ablation are equal, over the period of a year (Benn and Lehmkuhl, 2000).

Figure 1.17 shows the profile of a corrie (cirque) glacier with the ELA dividing the accumulation and ablation zones. The ELA also generally defines the area of greatest thickness and velocity within a glacier, and therefore relates to the area of maximum erosion of the bed. ELA measurements can provide a sensitive record of local climate, and can enable the reconstruction of past climates. The altitude of ELA varies according to local climate, both of which are defined by the local precipitation and temperature conditions. This will be investigated further in chapters 5 and 6, where the ELAs of the Torridon and Applecross ice fields will be used to establish how glacial mass balance responded to changes in local climate.

1.5.2. Climate

By isolating the effects of climate on the growth and survival of ice, we can determine how a glacier responds to changes in local precipitation and temperature patterns. Glaciers can exist in warm maritime and Alpine climates such as New Zealand (Bradwell et al., 2008; Glasser, 1997), where snowfall is high enough to last through a warm summer. Where average precipitation is low, particularly in continental and polar regions (e.g. Antarctica), glaciers exist where low summer temperatures are insufficient to melt winter snowfall (Glasser, 1997). Glaciers in maritime settings are usually much more responsive to a decrease in temperature during the winter season, and tend to readvance, generally forming annual ice-marginal landforms. In arid, continental settings, ice bodies react at a much slower rate, and therefore do not display the high-resolution response observed in maritime glacial regions (Boulton, 1986, Wilson and Evans, 2000). The presence of sea ice in the Polar Regions enhances
continentality (i.e. dry and cold), reduces overall precipitation availability (Björck et al., 2002) and lowers temperatures, reducing ablation and extending the length of time during which a glacier exhibits a positive mass balance. During glacial episodes, previously maritime areas can be subject to the same phenomenon, changing mild, wet climates to cold, dry climates, not dissimilar to those of the Polar Regions today (Golledge et al., 2009).

Ice bodies, once established, can have impacts on local, regional and even global climate. Some large ice sheets are able to buffer themselves from climatic warming, and lag behind changes in climate in a state of disequilibrium. For example, the high albedo and elevation at the centre of the Antarctic ice sheet, leads to temperatures as low as -70°C (Benn and Evans, 1998). Even valley glaciers can create local temperature anomalies. A glacier of 10-20 km in length may decrease local temperatures by as much as 1.6 – 1.7°C, resulting in a positive mass balance due to the suppression of ablation processes (Golledge et al., 2009).

1.5.3. Topography

Although climate is the major driving force behind the growth and demise of glaciers, landscape form also determines the threshold conditions for their morphology. Table 1.1 is adapted from Golledge (2007a) and demonstrates the glacial landsystem elements for a range of ice masses on a range of scales. As noted in section 1.5.1, ice sheets and caps generally self-regulate, irrespective of climate. Ice bodies >1,000 km² also flow regardless of underlying topography.

Topographical control becomes increasingly important either during deglaciation in ice marginal locations, or where isolated mountainous areas (less than ~100 km²) exist. Ice fields, valley and cirque (corrie) glaciers which form in these regions are confined and channelled by the surrounding landscape. The glacial features deposited by such ice bodies are therefore concordant with the local topography, in contrast to ice sheets and caps, which generally deposit features irrespective of topography (Golledge, 2007a).
### Table 1.1. The landsystem elements of ice bodies at a variety of scales. As a general rule, as thickness decreases, topographical control becomes more important. Adapted from Golledge, (2007a).

<table>
<thead>
<tr>
<th>Landsystem elements</th>
<th>Decreasing ice thickness/increasing topographic control</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Ice sheet</td>
</tr>
<tr>
<td>Approximate scale</td>
<td>&gt; 10,000 km²</td>
</tr>
<tr>
<td>Geometry</td>
<td>Domed accumulation area with outlet lobes of variable gradient</td>
</tr>
<tr>
<td>Ice marginal landforms and sediments</td>
<td>Generally offshore at grounded or calving margins</td>
</tr>
<tr>
<td>Subglacial sediment deposition</td>
<td>Often complete removal of pre-existing sediments in upland areas. Ubiquitous deposition of subglacial till</td>
</tr>
<tr>
<td>Patterns of local ice flow</td>
<td>Determined by surface slope, irrespective of topographic slope. Flows in a radial pattern from the central dome/iceshelf</td>
</tr>
</tbody>
</table>

On a local scale, aspect and mountain shape are important to the formation and maintenance of glaciers. High-altitude mountain tops will favour accumulation of snow, but narrow, steep summits will not be able to accommodate the formation of a glacier (Manley, 1955), thus explaining the existence of nunataks. The size of the accumulation zone is of vital importance, not only for direct precipitation, but also for the indirect transference of wind-blown snow and avalanches (Sutherland, 1984). Studies of glaciers throughout highland Britain (e.g. Sissons, 1979a; 1979b; 1980) indicate that between 10-22% of a glacier's annual accumulation originates from these sources.
1.5.4. Basal thermal regime

Prevailing climate and local topography play a vital role in defining the thermal regime of a glacier. In turn, the nature of glacial erosion and deposition is heavily dependent on basal thermal regime (Kleman et al., 1999; Glasser and Hambrey, 2001). Cold-based or partially cold-based (polythermal) glaciers survive where average precipitation is low. In continental and polar regions (e.g. Antarctica), glaciers generally survive where low summer temperatures are insufficient to melt winter snowfall (Glasser, 1997). The glacial landsystems associated with such glaciers depend on the proportion of the glacier underlain by cold-based ice and are associated with zones of permafrost. Cold-based glaciers move via internal deformation. These glaciers tend to be thin and non-erosive, thus preserving the surface on which they reside and inhibiting the formation of glacial landforms (Kleman and Glasser, 2007; Glasser and Hambrey, 2001). These glaciers cannot flow by basal sliding due to sub-freezing temperatures and therefore flow at very low velocity via internal deformation of the upper part of the bed (Boulton, 1979). These low velocities are sufficient to achieve the balance between low annual accumulation and ablation and are usually characteristic of glaciers in continental settings.

Polythermal glaciers are mainly found in polar and permafrost regions and are generally characterised by basal melting in the accumulation zone and basal freezing at the margin (Glasser, 1997). The Arctic archipelago of Svalbard is home to a number of polythermal glaciers due to the fact that its climate is relatively mild for its northern latitude (Glasser and Hambrey, in Evans (2003)). This is because it lies at the most northerly extent of the Norwegian Current, a branch of the Gulf Stream. The glacial landsystems associated with a polythermal glacier depend on the proportion of the glacier underlain by cold-based ice and are associated with zones of permafrost.

Temperate, warm-based, high mass-turnover glaciers exist in maritime and Alpine climates such as New Zealand (Bradwell et al., 2008; Glasser, 1997), where snowfall is high enough to last through a warm summer, and the snowline is below the altitude of the accumulation area for the majority of the balance
year. The aspect of mountain ranges can also enhance the preservation capabilities of such areas throughout the warmer months. Warm-based glaciers flow via basal sliding, causing modification of the bed (Glasser, 1997). Warm-based glaciers have a high balance velocity, where ice velocity (or annual ice accumulation) is high enough to discharge annual ablation (Benn and Evans, 1998). Resultant landscapes are often characterised by glacially-streamlined bedrock (i.e. erosion) in the accumulation zone, and streamlined deposits (i.e. deposition) in the ablation zone (Glasser and Hambrey, 2001). Bias towards a warm-based, dynamic thermal regime is often observed in the deglacial record where the last regime prior to deglaciation was warm-based, as it only takes warm-based glaciers a fraction of the time to obliterate any evidence of a cold-based regime. This should be considered where a landscape exhibits predominantly warm-based characteristics.

Basal shear stress is of vital importance to the understanding of the processes occurring at the subglacial shear zone, as this is the area in which the glacier interacts with the landscape. The basal thermal regime of an ice sheet controls the mechanisms of glacier flow, including the extent of basal sliding; the potential for sediment deformation; the production of meltwater and the level of geomorphological activity in the basal region (Glasser, 1995; Kleman and Hättestrand, 1999; Glasser and Siegert, 2002). In this instance, glacier reconstruction (Chapters 4 and 5) can be assessed in terms of its accuracy by calculating basal shear stress. By comparing basal shear stress results to geomorphological evidence to zones of warm and or cold basal activity can be identified. This will be addressed section 6.5.2 of the discussion (Chapter 6).
1.6. Glacial response to climate change in the Late Devensian

1.6.1. Late Devensian chronology

The INTIMATE project (Hoek et al., 2008; Lowe et al., 2008) correlated palaeotemperature data from the NGRIP ice core (Andersen et al., 2006; Svensson et al., 2006) with marine and terrestrial proxies sampled throughout the North Atlantic region, to form a Lateglacial chronology (Figure 1.18). The chronozones associated with this record vary according to location, and differ within Europe, as well as across the globe. To avoid confusion, the British terms for chronozones will be used throughout. A table which matches British and NW European chronostratigraphic terms is available for reference in appendix 1. The NGRIP record extends back to the Ipswichian interstadial (~128 – 116 ka), prior to the onset of the Devensian (116 - ~11.5 ka). In this instance however, focus is placed on the most recent part of the record; the Late Devensian (~26 – 11.6 ka), and the transition from a glacial to an interglacial climate in the North Atlantic (Figure 1.18). Within the Late Devensian NGRIP record, three distinct Greenland Stadial (GS) and Interstadial (GI) chronozones have been identified, and are used as a benchmark, to put glacial fluctuations in the British Isles in context with basin-wide climate change.

1.6.2. The Dimlington Stadial (GS-2)

At approximately 39 ka, during oceanic reorganisation, the Gulf Stream was weakened and sea ice built up on the NE Atlantic seaboard (Rose, 1985) promoting a colder, more continental climate. Chiverrell and Thomas (2010), suggest that the build- up of the last British-Irish Ice Sheet (BIIS), took place between ~35 and 32 ka. At ~26 ka (figure 1.18 and appendix 1) a further episode of climatic cooling took place, leading to the onset of a dominant polar climate regime (Coope, 1977). The period ~26 ka – 14.7 ka is known as the Dimlington Stadial (Rose, 1985).
The maximum extent of the BIIS was probably achieved during the Last Glacial Maximum (LGM) which is thought to have lasted from 27–21 ka (Chiverrell and Thomas, 2010). The LGM has been classified according to the global maximum ice volume during the last glacial cycle, which also corresponds to a significant and prolonged trough in the marine $\delta^{18}O$ record (Martinson et al., 1987). This event is also associated with a low in global eustatic sea level (Yokoyama et al., 2004).

The BIIS covered all of Scotland and Ireland, much of Wales, and the majority of northern England (Figure 1.19), with the ice shed centred over the Western Scottish Highlands during the Dimlington stadial. The thickness of ice cover and the presence of nunataks at the LGM have been heavily debated. Ballantyne (1997) and Ballantyne et al., (1998) proposed the existence of ice-free summits, reconstructed from periglacial ‘trimlines.’ However, recent evidence suggests that this weathering contrast may be created by differential erosion within a glacier and that a trimline defines an englacial boundary between warm and cold-based ice. Work carried out to investigate this phenomenon in Wester Ross will be introduced in section 1.7.2.
Offshore surveys of areas of the British Isles have uncovered evidence of the BIIS and its associated ice streams, during the Last Glacial (Figure 1.19). The western margin of the British Isles has a gradually sloping continental shelf which breaks
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at 180-250 m depth. To the east of Scotland, the main basins are the central and northern North Sea. To the west of Scotland, depositional centres are divided along the outer continental margins by the Wyville-Thomson ridge (between Orkney and Shetland) which also divides the Faroe-Shetland channel and Rockall Trough. The north and south inner shelf is divided by the Rubha Reidh Ridge, running approximately E-W across the Minch Sound, between Lewis and mainland Scotland.

The continental slope is characterised by a series of Trough-Mouth Fans (TMFs) and prograding wedges, which are composed of glacial sediments, deposited during successive glacial maxima, including the LGM. There are four major TMFs/prograding wedges on the NW margin of the British Isles (Figure 1.19). The Barra/Donegal fan is ~400 m thick, the Sula Sgier fan is ~200 m thick and the Rona and Foula wedges are both <200 m thick. The heterogeneous composition of the NW margin fan systems is caused by changes in the proximity of ice, the glacial discharge, bottom current flow strength, and the mechanisms of down-slope sedimentation. Increased clay input and expanded sediment sections tend to be indicative of cold conditions (glacials) and evidence of erosion can be attributed to the stronger currents present during warmer interglacial periods (Sejrup et al., 2005).

At the LGM, the source areas of the BIIS were drained by several fast-flowing ice streams. The Minch ice stream has been identified in recent literature as a prominent feature which drained ice from NW Scotland during the Late Devensian. Bradwell (2005) and Bradwell et al., (2007) have identified 11 large-scale onshore sub-glacial bedforms, which give evidence of ice direction. It is also believed that these large-scale bedforms may be the geomorphological signature of ice stream onset zones (the point at which ice starts to “stream,” as defined by the surrounding topography). Sidescan sonar and multibeam echo sounding data from the Minch show numerous elongate features of a similar scale, which can be interpreted as ridges and grooves caused by sub-glacial processes. The “Minch palaeo-ice stream” has been tentatively reconstructed thorough the use of onshore and offshore evidence (Stoker and Bradwell, 2005;
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Bradwell et al., 2007). At the LGM, ice from mainland Scotland, north Skye and Lewis coalesced in the Minch Sound, from where it flowed across the continental shelf, depositing sediment in the Sula Gsier Fan (Figure 1.19). Models of ice cover over the present day North Sea suggest that the British ice sheet was confluent with the Fennoscandian ice sheet between 30 and 25 ka (Bradwell et al., 2008; Sejrup et al., 2009) and formed a large North Sea ice stream which deposited sediment at the Rona and Foula wedges (Figure 1.19). As sea level rose, a marine embayment opened in the northern sector of this ice stream, forcing the separation of the neighbouring ice sheets, and leading to the formation of moraines close to the present-day shoreline of NE Scotland between 24 – 18 ka (Bradwell et al., 2008).

The BIIS was also drained to the west towards the Barra-Donegal fan (Figure 1.19) and south by numerous ice streams, including the Irish ice stream (McCarroll et al., 2010) and the Tweed ice stream (Everest et al., 2005). During glacial episodes, ice streams transported ice and sediment away from the source regions, and deposited it in TMFs and wedges at the edge of the continental shelf. The climatic reversal observed in the NGRIP record at ~14.7 ka (Figure 1.18), is thought to have been prompted by a reorganisation of the global oceanic heat conveyor, which resulted in strengthened North Atlantic Deep Water (NADW) formation and an increase in Gulf Stream efficiency. This had a profound effect on the glacial response and ecosystems of the eastern North Atlantic seaboard during deglaciation, in particular the British Isles (Lowe et al., 1994; Boessenkool et al., 2001; Weaver et al., 2003).

1.6.3. The Lateglacial Interstadial (GI-1)

In the NGRIP curve (Figure 1.18), the Lateglacial (Windermere) Interstadial (~14.7–12.9 ka) (Ballantyne, 2010) is characterised by high amplitude temperature fluctuations, and can be divided into five distinct warm and cool episodes (GI-1a–e), each less than 600 years long (Figure 1.18). Chronozones GI – 1e, c and a were characterised by climatic amelioration, particularly during GI-1e, in the early interstadial, when temperatures rose almost to Holocene levels.
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This contrasted to the cold, dry climate of chronozones GI-1d (the Older Dryas) and 1b (the Intra-Allerod Cold Phase), which occurred at 14.0-13.9 and 13.2-12.9 ka respectively (Lowe et al., 2008). Chironimid (midge) records from Whitrig Bog in SE Scotland (NGR:622 349) indicate drops in mean summer temperature of 4.9°C in GI-1d and 2.2°C in GI-1b (Brooks and Birks, 2000), which are likely to reflect more widespread cooling and indicate that conditions were favourable to ice survival in the Lateglacial interstadial. Proxies sampled throughout the North Atlantic and NW Europe concur with observations of the NGRIP record, and indicate that temperature fluctuations and climate instabilities were widespread throughout the Lateglacial (e.g. Anklin et al., 1993; Witte et al., 1998; Mayle et al., 1999; Boessenkool et al., 2001; Brooks and Birks, 2000).

Figure 1.20 shows the positioning of some well-documented moraine complexes, focussed around the main Scottish ice-shed. Following the LGM, ice retreated towards the main ice shed in the British Isles. Retreat was interrupted by a series of readvances and stillstands, perhaps relating to millennial-scale climate oscillations. On the Scottish continental shelf, submarine moraines have been identified at the edge of the continental slope to the north and west of Lewis, to the west of Orkney and Shetland and also to the north east coast of Shetland, marking the northern-most extent of the Dimlington ice sheet. Off the east coast of Scotland, the Wee Bankie moraine and the Bosies Bank moraine in the outer Moray Firth (Hall and Bent, 1990), represent a stage in the retreat of the Dimlington ice sheet.

Retreat of the BIIS in eastern Scotland can be characterised by two distinct complexes of moraines. The Aberdeen-Lammermuir ‘moraines’ (Sissons, 1967), were formed between 15.4 ka (the minimum age for the onshore recession of ice onto coastal Aberdeenshire) and 12.6 ka (calibrated from 14C data from a bog in Braemar. Work by Clapperton and Sugden (In Gordon, (1977)), suggests that these moraines are in fact glaciofluvial kames, overlain by till.
Figure 1.20. Readvance moraines formed during overall deglaciation following the Dimlington Stadial, based on mapping by Sissons, (1967); Robinson, (1977); Clapperton and Sugden (In Gordon (1977)); Robinson and Ballantyne, (1979); Everest, (2003) and McCabe et al., (2007).

The Perth moraine (Figure 1.20) is thought to represent a stage in the retreat of the BIIS, but its glacial origins are also debated. McCabe et al. (2007), describe the complex as a suite of readvance moraines. Peacock et al. (2007), disagree with this interpretation due to a lack of evidence that sediments adjacent to the moraine were overridden by an ice body at this time, and suggest that the Perth moraines formed during a winter still-stand event.

In the northwest of Scotland, the Wester Ross Readvance (WRR; Figure 1.20) is a particularly well-documented event. It is thought to represent the first onshore stage in the retreat of the Dimlington ice sheet. The Lateglacial interstadial is thought by some authors to have been characterised by ice-free conditions in NW Scotland (Sissons, 1967; 1977; Lowe et al., 1994). However, recent cosmogenic exposure ages from moraine-ridge boulders on the WRR moraine
complex (Bradwell et al., 2008; Ballantyne et al., 2009; Ballantyne, 2010) imply that ice remained in Wester Ross between 14-13 ka, following the distinct warming trend in GI-1e (Figure 1.18). This will be discussed further in section 1.7.3, which focuses on historical studies of the glacial geomorphology in Wester Ross.

1.6.4. Younger Dryas (GS-1)

The onset of the Younger Dryas (YD) cooling event (12.9 ka) is recognised in palaeoclimatic records throughout NW Europe (e.g. Isarin et al., 1999; Renssen et al., 2001; Lowe et al., 2008), although the global signal is still heavily debated (e.g. Lowell and Kelly, 2008). The rapid onset, magnitude and non-correlation of the YD event with Milankovitch cycles suggest that it was caused by a catastrophic event (Broecker et al., 1989; Broecker, 2003). This theory involves a significant reduction or shutdown of the North Atlantic thermohaline circulation in response to a sudden input of fresh water from Lake Agassiz during deglaciation of the Laurentide Ice Sheet (LIS) (Broecker, 2006). Recent debate has surrounded the possibility of outburst flooding as a result of a meteorite impact in North America ~13 ka (Firestone et al., 2007). The discovery of nanodiamonds (supposedly of extraterrestrial origin) around the impact site in North America has since been called into question (Pinter et al., 2011).

In a recent paper by Broecker et al. (2010), contrary evidence implies that the Younger Dryas had a fundamental role to play in the last deglaciation, and that it was an integral part of the deglacial sequence of events which led to the last deglaciation. Studies of CO₂ levels in the ice core record for the past four terminations have revealed evidence of similar deviations from climatic amelioration during deglaciation.

Whatever the initial trigger, the Younger Dryas cooling was caused by a reorganisation of North Atlantic circulation, the shut-down (or weakening) of the NADW, and a diminishing Gulf Stream. The consequent drop in mean annual air temperature of 7.5-10°C in Scotland (Hubbard, 1999; Isarin and Renssen, 1999; Alley, 2000; Brooks and Birks, 2000) led to a rapid transition
back to full glacial conditions on the NE Atlantic seaboard, which lasted for ~1.2 ka. As sea ice migrated south again, it is likely that depressions and stormy weather were more common due to the juxtaposition of sea ice and open seas, and snowfall in NW Scotland was associated with south and south-westerly warm and occluded fronts. Snowfall was effectively scavenged by the ice cap and marginal ice fields to the west, leading to diminished snowfall in the rain shadow in the east (Sissons, 1979a; 1979b).

The renewal of glacial conditions in the British Isles during the YD is known locally as the Loch Lomond Stadial (LLS). During the LLS, an extensive ice cap was centred over the main ice-shed in NW Scotland (Sissons, 1979b; Evans, 2006; Golledge, 2010) and satellite ice fields were focussed around the main ice cap, in the Central Highlands/Eastern Grampians (Sissons, 1973); Sutherland (Benn and Lukas, 2006; Lukas and Bradwell, 2010); Assynt (Bradwell, 2006); Torridon (Sissons, 1977); Applecross (Robinson, 1977; Robinson and Ballantyne, 1979; Jones, 1998); Skye (Benn et al., 1992; Ballantyne, 1989); Mull (Ballantyne, 2002a); Arran (Ballantyne, 2007) and the Lake District (McDougall, 2001). Individual cirque and valley glaciers were also found in favourable locations on Lewis (Ballantyne and McCarroll, 1995; Ballantyne, 2006); in the Southern Uplands (Cornish, 1981), North Wales (Hughes, 2009) and in the Cairngorms (Everest, 2003).

Golledge et al., (2008) used a high-resolution model to simulate Younger Dryas glaciation in NW Scotland. An optimum fit model, based on empirical data collected during numerous studies, suggested a temperature depression of 10ºC relative to today, and a steeper north-eastward precipitation gradient. This simulation implies that the YD maximum was achieved within 400 years of the onset of the cooling, at 12.6 – 12.4 ka, with terminal positions being reached sooner in the north-west. Modelling studies by Hubbard (1999) and Golledge and Hubbard (2005) imply that LLS ice was sustained at its maximum extent by a cool, dry climate, maintained by an increasingly continental climate on the NE Atlantic seaboard. Short summer ablation seasons led to an enhanced seasonality (Isarin and Renssen, 1999; Denton et al., 2005; Lie and Paasche., 2006), which led to a much more arid climate than predicted by Scottish
glaciological reconstructions of precipitation and temperature (P/T) conditions (e.g. Benn and Ballantyne, 2005; Ballantyne, 2006; Lukas and Bradwell, 2010). This discrepancy will be discussed in detail in chapter 6.

It is suggested that ice survived until 11.5 ka in favourable areas of the British Isles (Atkinson, 1987; Alley et al., 1993; Stone et al., 1998, Ballantyne, 2010) despite the warming commencing at ~11.7 ka (Figure 1.16). The transition into the Holocene was characterised by a rapid rise in air temperatures across Europe (Coope et al., 1998) with the Greenland ice core $\delta^{18}O$ signal suggesting a climatic turn-around on a scale of decades, and a total warming of 10 ± 4°C (Kobashi et al., 2008). The causes of LLS deglaciation will be also be considered in chapter 6.

1.6.5. Late Devensian sea level in Scotland

Fluctuations in global sea level throughout the Lateglacial also had an impact on the pattern of glaciation. Massive lowering in sea level (by as much as 140 m) at the time of maximum glaciation meant that the BIIS and the Fennoscandian Ice Sheet (FIS) were confluent in the North Sea, and the BIIS extended west to the edge of the continental shelf (Holmes, In Gordon et al., 1997). Shennan et al. (2006) have estimated sea level records from 18 ka to present for Arisaig, NW Scotland using biostratigraphic indicators (e.g. diatom, pollen, foraminifera). This record infers non-monotonic sea level change, due to glacio-isostatic rebound and eustasy (or global meltwater flux), which dominated at different periods. Rapid sea level fall continued until ~14 ka, concurrent with the acceleration of global meltwater discharge. During the Younger Dryas, relative sea level fell at a steady rate as isostatic uplift occurred more rapidly than global sea level rise (McIntyre and Howe, 2010). Following a sea level low-stand ~11 ka, relative sea level rose to ~ 7 m above present in Arisaig (Shennan et al., 2006). This Holocene high-stand persisted for ~1 ka, before the onset of a drop towards modern sea level.

The local details of the NW Scottish coastline are the product of changes in sea level and coastal processes. Deposition and erosion along palaeo-shorelines can
provide evidence to support sea-level estimates, and can also provide evidence of the glacial configuration at a certain time. Depositional shorelines formed during ice sheet deglaciation are represented by terraces of sand and gravel which are fragmented in places. McCann (1966) recognised two main raised beaches in Loch Carron, Wester Ross at ~26 to 29 m and ~18 m. The higher beach is well developed along open coasts, but is absent from the inner sea lochs such as Upper Loch Torridon, and is thought to have formed during a sea-level high. The terrace at ~18 m can be observed at the head of Upper Loch Torridon, and was thought to have been deposited during the retreat of the main Scottish ice sheet, contemporaneous with a fall in sea level. Readvance of glaciers (during the LLS) subsequent to the deposition of these beaches, led to erosion in the inner lochs, hence the lack of terraces away from the open coast.

Erosion during the Lateglacial has led to the formation of a distinctive rock platform, the Main Postglacial Shoreline (McIntyre and Howe, 2010), which can be recognised along much of the western Scottish coastline. On the NW Scottish shoreline, a high coastal rock platform has been described by Sissons (1982), sitting at 18-51 m height, and reaching widths of 300 m wide on the Applecross Peninsula, Wester Ross. This part of the shoreline reaches 23.5 m in height, and is associated with the glacial till described by Robinson (1977), and the raised beaches described by McCann (1966). It is suggested that this platform was formed when the WRR was at its maximum extent, at a period of relatively stable sea level. During ice-sheet retreat, it is thought that erosion of the platform ceased as sea levels fell. Sissons (1982) favours the explanation of rapid periglacial weathering of the exposed rock platform during relatively stable glacial episodes.

The main (lower) rock platform is well-developed between the Firth of Clyde and Mull. It is generally 10-20 m wide, but can reach up to 100 m width in places, and sits at <12 m.a.s.l, where postglacial isostatic rebound is particularly apparent. Cosmogenic isotope exposure dating of this shoreline indicates that it formed during a postglacial event (Stone et al., 1996). It is thought that this event was the LLS, when periglacial weathering enhanced the susceptibility of the
exposed rock to freeze-thaw processes. The relative stability of sea-level during the LLS meant that marine erosion was concentrated at the height of the shoreline for several thousand years, and supports Sissons' (1982) theory of erosion during glacial episodes.
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1.7. Previous studies in Wester Ross

1.7.1. Geology

The Torridon and Applecross areas of Wester Ross are located to the west of the Moine Thrust Zone (Figure 1.21), which stretches for ~200 km from Loch Eriboll in the north down to the Sound of Iona. This western-foreland area is composed of the oldest rocks in the British Isles. The forelands avoided the effects of the Caledonian Orogeny during the closure of the Iapetus Ocean (~550-400 Ma) which completely reconstructed the area to the east of the Moine Thrust. The collision of Laurentia and Baltica during the Silurian and Devonian is a probable driving mechanism for the thrusting, and late-Palaeozoic tilting events during the led to a dip of 8-12°E throughout the Moine Thrust Zone. (Johnstone and Mykura, 1989; Woodcock and Strachan, 2002).

The oldest rocks in the British Isles are the Lewisian Gneisses. The initial formation of the Lewisian complex took place from 2,900-1,100 Ma (late Archean-Proterozoic) as the Earth’s crust cooled to form the Laurentian continent. Lewisian rocks of late Archean age (Scourian) were formed under high-grade metamorphic conditions and are composed of granular pyroxene gneiss. The low-grade metamorphic, early Proterozoic rocks (Laxfordian) are composed mainly of deformed gneiss. Lewisian gneiss basement rocks are chiefly exposed beyond the two study areas of Torridon and Applecross (Floyd et al., 2007). The lowland “knock and lochan” landscape outwith these areas, is characterised by an undulating landscape dominated by Lewisian gneiss bedrock, dotted with numerous small tarns (Geike, 1887; Robinson, 1977). A period of prolonged erosion following the Laxfordian led to the formation of an undulating landscape. In the late Precambrian (1000-750 Ma) this landscape was buried in thick layers of sediment, which belong to the Torridonian group. The ~7 km thickness of these sediments suggests that they accumulated in faultbounded subsiding basins, under fluvial and possibly shallow lacustrine conditions.
A particularly good example of a sediment-filled valley can be seen to the north of Loch Maree (Figure 1.22), where erosional processes created a much more pronounced sub-Torridonian landscape than elsewhere in the region (Johnstone and Mykura, 1989). 400 Ma after the deposition and erosion of the Torridonian sandstones, Cambrian sediments were deposited on flat surfaces (Figure 1.23). These quartz-rich sediments were subsequently metamorphosed to form layers of bright-white quartzites, which are found in outcrop to the west of the Moine Thrust. Quartzite rocks form resistant cap rocks on many of the higher Torridonian sandstone mountains (Johnstone and Mykura, 1989; Ballantyne et al., 1998b).

The North Atlantic began to open ~250 Ma, in the Mesozoic, initiating the separation of the European and North American plates. The Western Isles were separated from the mainland by the Minch Sound, which also opened at this time (Mitchell, In Gordon et al., 1997). Uplift from the Late Oligocene onwards (~25 Ma) led to the formation of rift basins off the west coast of Scotland.
Figure 1.22. Torridonian sandstone overlying Lewisian gneiss on the banks of Loch Maree.

Figure 1.23. Cambrian quartzite overlying Torridonian sandstone in Coire Mhic Fhearchair, Central Torridon. The feature shown is the "Triple Buttress" formation at the back wall of the corrie.
The concurrent subsidence of the North Sea basin and the uplift of the NW Highlands of Scotland led to a significant deepening of the west-coast sea lochs (fjords) by Quaternary glaciers (Hall, in Gordon et al., 1997). Within the foreland region of Wester Ross, regional-scale faulting and jointing associated with the Moine Thrust zone have been exploited by ice on all scales. Subsequent topographic evolution and extensive glaciations have created a terrain characterised by deep glacial valleys, knife-edge ridges, over-deepened corries and a glacially-dissected, flat-topped plateau in Applecross. The glacial geomorphological features of Wester Ross we see today are remnant of the Dimlington glaciation and subsequent glacial readvances (in particular the Loch Lomond Stadial readvance) have been studied in detail by a number of authors (outlined in the proceeding section) and have implications for the dynamics of the retreating BIIS in Wester Ross.

1.7.2. The Last Glacial Maximum

Much of the geomorphological evidence of the BIIS in Wester Ross has been erased during the deglaciation and readvance of LLS glaciers during the most recent cooling event in the YD. Beyond the extent of the LLS, there remains evidence of large-scale erosional landforms, thought to represent the subglacial dynamics and maximum vertical extent of the Dimlington ice sheet in Wester Ross.

**Mega-Scale Glacial Lineations (MSGLs)**

Bradwell, (2005) and Bradwell et al., (2007) recognised clusters of elongate ‘megagrooves’ in bedrock and superficial deposits throughout the NW Scottish Highlands and Islands. Giant glacial grooves or “Mega-Scale Glacial Lineations” (MSGLs) are longitudinally aligned corrugations, ~6-100km long. They form in sediment where ice flow accelerates, generally at ice-stream onset zones, where transverse strain transforms irregular bumps in the glacier base into longitudinally extensive keels of ice. These keels plough through soft sediment to create the characteristic groove and ridge pattern (Clark et al., 2003).
MSGLs have been identified in superficial deposits on the NW coastline of Applecross (Bradwell, 2005) using NEXTMap digital surface models (Figure 1.24). Counterpart ridges and grooves have also been observed offshore using sidescan sonar and multibeam echo sounding data from the Minch Sound (Stoker and Bradwell, 2005; Stoker et al., 2006). These features have been used as a basis for the reconstruction of the flow paths, catchment and basal shear stress of the Minch palaeo-ice stream (Bradwell et al., 2007) which streamed away from mainland Scotland to the NW, towards the edge of the continental shelf (Stoker and Bradwell, 2005).

**Glacial trimlines**

Glacial trimlines are weathering limits, defined to some extent by the vertical dimensions of a palaeo-ice body, separating glacially scoured bedrock below from frost-weathered bedrock above.

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**Figure 1.24. Mega–scale Glacial Lineations (MSGLs) on the Applecross Peninsula.** The lineations trend approximately N-S. Grid in 1 km increments.
Ballantyne et al., (1997) studied the high-altitude weathering limits throughout Wester Ross, and found trimlines at 700-960m, declining in height with distance from the ice-shed (the present water-shed). Trimlines have been inferred to represent the maximum vertical extent of an ice body (Figure 1.25a). In the instance of the British Irish Ice Sheet (BIIS), this would place its maximum altitude at ~900 m over the Western Highlands. However, investigations in Scandinavia and North America (e.g. Fabel, 2002; Briner et al., 2006),
demonstrate that high-altitude weathered surfaces are preserved by passive, cold-based ice. Figure 1.25b illustrates this scenario and the implications it could have for the altitude of an ice body. As basal shear stress, pressure and temperature decrease with height in a glacier, ice can adhere to the underlying substrate, thus preserving it. In this scenario, the trimline would be formed englacially, and would represent a minimum vertical extent. Recent evidence favours the englacial boundary theory as a more realistic situation at the LGM. Based on studies of high-altitude trimlines in NW Scotland, Ballantyne (2010) provides evidence to support this scenario:

1. Glacial erratics exist within blockfields above the weathering limit in certain locations, implying deposition by a glacial episode prior to the Dimlington stadial.

2. When the LGM surface of the BIIS is reconstructed, taking trimlines as a maximum limit, ice does not extend to the offshore limits unless unrealistically low basal shear stress is introduced to the model.

3. Modelling studies (e.g. Boulton and Hagdorn, 2006) cannot constrain the first scenario alongside climatic proxy data, but match optimal models very closely with the second (englacial trimline) scenario, meaning that the BIIS was at least 900 m thick.

The presence and origins of trimlines will be investigated in this case through geochronological (cosmogenic \(^{10}\)Be dating) and morphological (Schmidt hammer) studies. \(^{10}\)Be dates from bedrock above and below a lower altitude trimline than those identified by Ballantyne et al., (1997), should provide insights into the exposure history. Relative weathering of surface above and below the trimline will also be assessed using a Schmidt hammer. The detailed methodology of this investigation will be introduced in chapter 2 and the consequences of the results (shown in chapter 3) will be discussed in chapter 6.
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1.7.3. The oscillatory retreat of the Dimlington ice sheet

A series of 40 recessional moraines, up to 20 m high and 3 km long, have been identified up to 40 km offshore, using swath bathymetric imagery of the seafloor in the Minch Sound (Stoker et al., 2006; Bradwell et al., 2008). These moraines were formed during deglaciation following the Dimlington stadial, when ice extended beyond the present shoreline. The maximum extent of the WRR is demarcated by the Wester Ross moraine complex, which is located close to the coastline of Wester Ross and likely represents the first onshore stage in this sequence of offshore retreat moraines (Figure 1.26). The complex is comprised of discontinuous moraine ridges, boulder limits and eskers. These features form the maximum limit of a series of glacier lobes, which occupied Loch Torridon, Loch Gairloch and Loch Ewe (Robinson and Ballantyne, 1979) during the Late glacial interstadial (GI-1).

Robinson (1977) first described the extent of the Applecross moraine (Figure 1.24) and it was later mapped (Robinson and Ballantyne, 1979). The Applecross moraine demarcates the southern extent of the Loch Torridon lobe of ice, which existed during the GI-1. Striae on ground unaffected by the WRR show a consistent NW trend, in keeping with the observations of megagrooves, which represent the flow of the Minch ice stream (Bradwell, 2005; 2007). On ground overridden by the WRR, striae are dominantly directed towards the SW, asserting the definition of this moraine as a readvance moraine. In Torridon, the Wester Ross Readvance moraines on the northern flank of Baosbheinn and further west run in an E-W direction (Figure 1.26) suggesting that ice originated from the east and the Loch Maree trough (Ballantyne, 1986; Ballantyne and Stone; 2009). This however, does not discount the possibility that this lobe was also fed by valley glaciers from the south and the Torridon Mountains.
To the north, the Redpoint, Gairloch, and Aultbea moraines form the extent of a lobe of ice which resided in Loch Maree at the same time (Figure 1.26). Further north, the Achiltibuie moraine is thought to have formed contemporaneously with the more southerly moraines. Everest et al., (2006) used cosmogenic isotope dating to constrain the age of the readvance. They found that the mean age for the southern part of the Wester Ross Readvance system in Torridon was 16.3 ± 1.6 ka and suggested that it formed as a response to the most recent Heinrich event (H1). However, subsequent exposure dating by Ballantyne et al., (2009) reported a mean exposure age of 13.5 ± 1.2 ka from sandstone boulders on the southern-most moraines in the Torridon region. Further studies by Ballantyne et al., (2009) of a moraine thought to be a continuation of the Redpoint moraine at the north-west extremity of Baosbheinn (Figure 1.26) revealed a mean exposure age of 13.4 ± 1.2 ka. Bradwell et al. (2008) report a statistically similar mean exposure age (13.6 ± 1.4 ka) for moraines in the more northerly Assynt region. McCann (1966) mapped the extent of the WRR over the Applecross Plateau (Figure 1.26). However, these extents are not based on geomorphological evidence. It is not possible within the context of this research to decipher between the limits of the WRR and those of proceeding glaciation. These limits will therefore not be considered.

Where possible, the lack of consensus between previous studies will be addressed within this study. 

10Be cosmogenic dating will be used to gain insight into the timing of retreat prior to the onset of the LLS, through the collection of samples from relatively weathered bedrock above the extent of the LLS ice bodies which existed in Torridon and Applecross. Samples collected at relatively high altitude will be assessed in terms of their implications for the initial retreat of ice within these areas.

### 1.7.4. Loch Lomond Stadial

As the most recent glacial event in Wester Ross, the Loch Lomond Stadial (LLS) has been studied in great detail by numerous authors, who will be introduced in this section. The clarity and quantity of the glacial geomorphological features
within the boundaries of the LLS provide a reliable basis upon which to base the study of past climates, and the dynamic response of glaciers to climate change.

One of the most pressing issues regarding the glacial geomorphology of the most recent advance in the Younger Dryas is that of the origins and formation of “hummocky” moraine. This debate is relevant to much of the north western Scottish Highlands (including Wester Ross), where hummocky terrain characterises much of the ground within LLS ice limits.

**The ‘hummocky moraine’ debate:** ‘Hummocky moraine’ is a landform that looks disorganised when viewed from the ground, but when observed remotely, usually exhibits some form of organisation. Between the late 19th and mid 20th centuries, moraines were interpreted as glacial readvance or stillstand features (e.g. Geike, 1877; Charlesworth, 1956). Harker (1901) did however recognise that in some areas such as the Cuillin Hills on Skye, a type of disorganised, non-linear moraine existed. This idea had little impact on the prevailing view of moraine genesis until the 1960’s, when Sissons (1967) popularised the use of the term ‘hummocky moraine’ to describe the landscapes within the boundaries of the LLS in Scotland. This term was used to describe chaotic moraine and moraine which displayed some organisation, thought to have formed via stagnation processes. Sissons dismissed the idea that pattern in moraine sequences is the product of readvance or stillstand of the glacier toe, instead interpreting linear features as the products of crevasse squeezing. Stagnation was thought to be caused by the sudden climatic warming following the YD cooling. Sugden (1974) argued that localised areas of ‘hummocky moraine’ formed when ice bodies were isolated from their sources, and rather than representing a distinctive climatically induced stage in deglaciation, they reflect the distribution of locations conducive to stagnation.

Eyles (1983) studied moraine formation in a modern Icelandic setting, analogous to Wester Ross during the LLS. He concluded that push and dump moraines would be commonly found within the extent of LLS glaciers, and could form at the toe of an active glacier. The theory of active retreat was subsequently widely
used by several workers, including Benn (1990) and Bennett and Boulton (1993a; 1993b), who recognised that a pattern of LLS deglaciation could be mapped by defining successive palaeo-ice fronts, which can be interpreted in terms of climatic and glacial dynamics. Considerations of spatial and temporal variations in climate, sediment supply and topography, have led to a poly-genetic approach to the definition of ‘hummocky moraine,’ dividing it into three distinctive landforms, based on origins and landform associations: Ice marginal retreat moraine, chaotic stagnation terrain and subglacial flutes. All three landforms can be found within the LLS limits in Torridon and Applecross and will be discussed further in section 3.4.1.

**Torridon:** Sissons (1977) initially mapped the LLS on the northern mainland of Scotland, including the extent of LLS glaciers in Torridon (Figure 1.27). Within these limits, hummocky moraines, flutes, drift limits and boulder limits were also mapped. Sissons (1977) characterised much of the land below ~400 m as hummocky moraine formed during ice stagnation. He reconstructed the ice flow radiating out from central Torridon, supplemented by ice from the high north-facing corries. Valley glaciers were reconstructed extending to the north and east for ~5 km, through Strath Lungard, Glen Grudie and Coire Dubh Mor. A particularly extensive (~19 km²) valley glacier was thought to fill Coire Mhic Nobuil in the south of the study area, fed by a tributary glacier from Bealach a’ Chomhla. Hummocky moraine was used to define the extent of the LLS where end moraines were not developed in the north and east facing valleys. This contrasts to the clear end and lateral moraines, which demarcate the extent of the south and west-facing tongues of ice. Ballantyne (1986) adjusted Sissons (1977) limits to extend further into the valleys in the west and north of Torridon, recognising end moraines demarcating the extent of lobes of ice either side of Baosbheinn (Figure 1.27). The moraines defining the extent of the LLS truncate the Wester Ross Readvance moraines, implying that the LLS was a readvance, at least in this small area. Bennett and Boulton (1993a) studied the deglaciation of the LLS ice field in Torridon using aerial photographs at a scale of 1:10,000.
Ice limits were extended beyond the extent of hummocky moraine in the east of the study area, to converge with ice from the main ice cap (although the mechanical interaction associated with this confluence is negligible, and the radial arrangement of glaciers in Torridon implies that it acted as an independent ice body). The ‘hummocky’ moraine mapped by Sissons (1977) was reinterpreted as a sequence of retreat moraines, reaching from the lower valleys, to the high corries in the central area of Torridon. Each moraine arc was said to represent a stage in the retreat of the LLS ice, or a ‘palaeo ice-front,’ inferring an active pattern of LLS deglaciation in Torridon. The pattern of glacial retreat (i.e. was it active or passive?) will be investigated through the detailed study of the glacial geomorphology in part of Torridon. The deglaciation of the LLS ice field in Torridon can be reconstructed through the detailed field-based mapping of individual palaeo-ice fronts, providing a comprehensive account of the final retreat of the LLS glaciers, and their response to local climate change.
Bennett and Boulton (1993b) suggest that the application of dating techniques alongside decay histories clarify the deglaciation chronology. Cosmogenic isotope dating is utilized within the scope of this study to provide a framework for deglaciation in Torridon.

Sediment ridges elongated in the direction of palaeo-flow are also known as flutes and have been observed in Wester Ross on multiple scales. Hodgson (1986) studied flutes in the south Torridon region, and demonstrated that they achieved their morphology via subglacial deformation of the pre-existing till, i.e. the till deposited during the Dimlington glaciation, but found no unequivocal evidence of obstructions as the cause of initial deposition.

Wilson and Evans (2000) studied the glacial geomorphology of Coire a’ Cheud-cnoic, or ‘The Valley of 100 Hills’ in Glen Torridon, a valley which was part of Hodgson’s (1986) study. When viewed from the north (Figure 1.28), the landscape of this valley looks chaotic and hummocky, but when mapped using aerial photographs, Wilson and Evans found that many of the mounds were
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Elongate N-S, parallel to palaeo-ice flow. These flutings are ~1 m high, 10-20 m wide and up to 500 m long and were also found to be superimposed on arcuate ridges, forming a palimpsest landscape and corroborating Hodgson’s sedimentological observations that the flutes were reworked from a till deposited during a former (most likely a Dimlington stadial) glacial event.

A detailed study of the glacial geomorphology within Torridon will enable the identification of such streamlined features, which can be used to aid the reconstruction and identify palaeo-ice flow (i.e. flutes form parallel to ice flow) of the former ice body in Wester Ross. Care will be taken to look for any evidence of variation in ice flow direction, which could be indicative of palimpsest terrain.

**Applecross:** Robinson (1977) described the maximum extent of LLS ice on the Applecross Peninsula as a complex of four major valley glaciers in Coire Nan Arr, Coire na Ba/Bealach na Ba, Coire Attadale and Coire nan Cuileag flowing respectively to the east, south east, north west and west. These glaciers were joined at height by breaches, in particular Bealach nan Arr, a breached watershed at 600m, between Coire Nan Arr and Coire Attadale. Seven independent glaciers in the north-facing corries (Figure 1.29) dissected the plateau that characterises Beinn Bhan (896 m), the highest point on the Peninsula. Robinson (1977) suggests that the narrow, westward-dipping plateau around the Beinn Bhan ridge would not have promoted snow accumulation, rather periglacial activity during the YD. It is presumed that prevailing south-westerly winds caused snow drift into the north-facing corries (Sissons and Sutherland, 1976; Sutherland, 1984).

Jones’ (1998) investigations in Applecross (Figure 1.29) tested field-based geomorphological evidence against a high-resolution glaciological model (Hubbard, 1997). Trimlines were used as a basis for the vertical limit and ice marginal moraines were used to constrain the lateral extent of the LLS ice field. The interpretation of the field evidence agreed with the model, suggesting that ice cover was confined to individual valleys, rather than breaching the watershed.
To account for Robinson’s observations of scouring in the breaches (such as the Bealach nan Arr), Jones (1998) suggests that these surfaces were preserved by snow and ice cover during the maximum extent of the LLS, which would imply that they represent scouring during an earlier glaciation.

This research aims to investigate the contrasting results of Robinson (1977) and Jones (1998) through the collection of bedrock samples from the breached area in Applecross, for cosmogenic isotope analysis. It is hoped that the results of these investigations will provide an exposure age for the uppermost scoured area, placing it within the timescale of the LLS or an earlier glacial event.
A particularly interesting aspect of LLS lateral-terminal moraine morphology was investigated by Benn (1989), who considered the within-valley asymmetry of these features in Coire nan Arr, Applecross (Figures 1.29 and 1.30) and other valleys around Wester Ross and the Isle of Skye. Benn (1989) related moraine asymmetry to the distribution of steep rock walls (free faces), which is associated with the aspect of a particular valley. He concluded that a positive correlation exists between the volume of a moraine and the distribution of free faces, resulting from the variation in slope retreat rates between exposed, steep (40-45°) slopes which retreat at a rate of ~1.65-3.38 mm yr⁻¹, and debris mantled slopes, which retreat at one or two orders of magnitude lower. As steeper rock walls tend to develop on exposed north and east facing valley sides, it follows that larger lateral-terminal moraines tend to form on the same side. These principles will be investigated in the E-W trending Coire Mhic Nobuil in Torridon, where detailed glacial geomorphological investigations have taken place.
Since deglaciation, the landscape of Wester Ross has been shaped by paraglacial adjustment. The retreat of valley glaciers exposed landscapes in an unstable condition, which were prone to rapid and extensive modification (Ballantyne, 2002b; 2002c).
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Figure 1.33. An exposure (approximately 20 m wide and a maximum of 5 m high) in a roadside quarry above Inveralligin (Figure 1.2). The red line divides the upper and lower facies.

The exposure of oversteepened rock-faces, which experience stress-release following glacial unloading, can lead to slope failure and rock slides. Evidence of a large rock-wall failure can be observed in Toll a’ Mhadaidh Mor (Figure 1.2), under Beinn Alligin in Torridon (Figure 1.31). The corrie floor is covered by ~1 km$^2$ of boulders, extending from an obvious scarp on the south-west facing slope of Beinn Alligin. This feature was originally thought to be a rock glacier (Sissons, 1976a). However, Ballanytne and Stone (2004) have since classified it as a ‘Sturtzstrom’ or an ‘excess run-out avalanche,’ caused by paraglacial stress release due to the propagation of an internal joint network, which eventually resulted in the collapse of part of Beinn Alligin in the mid-Holocene (3950 ± 320 a).

Ballantyne and Stone (2009) also studied a boulder ridge below the north-west face of Baosbheinn (Figure 1.32). Previously interpreted as a protalus rampart by Sissons (1976a) and a rock glacier by Sandeman and Ballantyne (1996) this feature has been reinterpreted as runout debris, from a failure which can be also attributed to rock-slope weakening and paraglacial stress release. Cosmogenic $^{10}$Be dating implies that the failure occurred sometime between 14-13.3 ka, possibly due to freeze-thaw of ice within bedrock joints. As this debris covers part of the Redpoint moraine (see section 1.7.3), it is likely that it formed following the retreat of the WWR glacier.

Slope adjustments have led to the alteration of the immediate post-glacial landscape. This is particularly apparent in the study of a man-made roadside quarry on the northern shore of Loch Torridon (Figure 1.33). Bennett (1999)
carried out sedimentological analysis on this section, concluding that it comprised two distinct facies. The lower facies was interpreted as a debris fan composed of a subglacial diamiction, formed through a series of debris flows. The upper unit is a weakly stratified unit, interpreted as the product of a sudden mass movement. A chronology for the formation of the units has been compiled by Bennett (1999). It is believed that the lower unit was a lateral moraine deposited on the shore of Loch Torridon. This moraine collapsed down the hillside to form a debris fan, forming a series of sub units in the lower facies. The upper layer is thought to be the result of periglacial weathering or 'gelifluction' of the slope during the LLS. This example attests to the importance of landscape response and the extent of the reworking of glacial sediments on steep slopes, and highlights the value of the distinctive glacial-depositional record in Wester Ross.
1.9. Aims and objectives

The general aim of this research is to provide insight into the timing and style of the last deglaciation in Wester Ross, whilst employing a combination of well-known and in some cases innovative methods. Given the key ideas covered in this introduction, this aim can be broken down into four integral questions, which take into account and reconsider various aspects of the glacial geomorphology, past glacial dynamics (geochronology) and past glacial climates (glacial reconstruction). This suite of questions will be addressed in the discussion (chapter 6) and conclusion (chapter 7).

Question 1: What can the glacial geomorphological signatures of Wester Ross tell us about the last deglaciation in NW Scotland?

The glacial geomorphology of post glacial landscapes such as those in Wester Ross, can provide insights into the dynamics of a glacier in retreat, and can also be used to infer the response of the glacier to climate change and local topographical constraints. The sensitivity of this region to the effects of the North Atlantic Gulf Stream, the location of the oceanic polar front and related atmospheric oscillations such as the North Atlantic Oscillation (Taylor and Stephens, 1998) has led to the formation of a particularly detailed geomorphological record. Considerations include:

- What are the predominant glacial-geomorphological landforms within the study areas? Variations in geomorphological signature indicate a change in climate or topographical control during deglaciation (Benn, 1997; Benn and Lukas, 2006).

- Was ice warm based or cold based/polythermal during deglaciation? The thermal regime of glaciers provides an indication of local climate and the style of deglaciation (Kleman and Hättestrand, 1999; Glasser and Hambrey, 2001) and can be interpreted through the study of glacial geomorphology.
Question 2: Can terrestrial Light Detection and Ranging (LiDAR) be applied to the mapping of glacial-geomorphological features, and how does it compare to relatively traditional mapping methods?

I will address the variety of techniques employed to map glacial geomorphological features and introduce ground-based LiDAR as a novel approach to mapping. Ground based, or terrestrial LiDAR scanners are mainly applied to ‘hard-rock’ geological studies (e.g. Carabajal et al., 2007), landslide studies (e.g. Rosser et al., 2005) and initially applied in a glacial geomorphological context in McCormack et al., (2008). Since initial applications in the 1960s (e.g. Schuster, 1970) the precision of LiDAR data has improved from several metres to a few millimetres (Bellian et al., 2005), depending on the scanner used and the distances it is used over, with the potential to provide a positive contribution to the future of glacial geomorphological mapping. In this instance, terrestrial LiDAR is used to scan a north-facing corrie (cirque) in central Torridon. Scans are taken from a variety of positions within the corrie alongside high-resolution digital photographs. Scans and photographs are georeferenced and overlaid in RiSCAN software and a 3D model of the corrie is generated. A preliminary assessment of the benefits and drawbacks of the application of LiDAR to geomorphological mapping is made, in comparison with established techniques, such as field mapping, mapping from aerial photographs, and mapping from Digital Elevation Models (DEMs) such as NEXTMap Great Britain™. The glacial history of the corrie is assessed by compiling a detailed map, composed of the geomorphological features mapped using all four techniques.

Question 3: What insights can cosmogenic $^{10}$Be isotope dating provide into the deglacial history of Wester Ross?

The measurement of the cosmogenic $^{10}$Be component of quartz in a rock can be used to provide an estimate of the length of its exposure to cosmic rays, and therefore its exposure age (Balco et al, 2008). Cosmogenic dating can provide insights into the rate and style of change associated with glaciation and/or
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deglaciation. Numerous studies of continental and mountain scale glaciations (see Gosse and Phillips, 2001, for a comprehensive list) have refined the technique, enabling short-lived glacial events to be dated more precisely (e.g. Gosse et al., 1995), thus providing more succinct glacial chronologies. Samples taken from large boulders lying on terminal moraine complexes provide the age at which a glacial advance reached its maximum lateral extent (e.g. Ballantyne et al, 2009). Exposure ages from within these boundaries can offer insights into the vertical extent and style of maximum glaciation, and the timing of deglaciation.

In this contribution, cosmogenic exposure ages are calculated for post-glacial sandstone bedrock surfaces from the central Torridon and Applecross areas (Figure 1.1.). Cosmogenic exposure ages are used in context with geomorphological and sedimentological studies to provide a detailed deglacial chronology for the centre of the Torridon and Applecross areas of Wester Ross during the last deglaciation. This includes the investigation of a low altitude trimline in Torridon, to establish the maximum extent of ice during the Younger Dryas, and to investigate the possible survival of ice before the onset of climate cooling at ~12.9 ka. In Applecross, discrepancies between two ice surface reconstructions (Robinson, 1977; Jones, 1998) are addressed by taking samples form the areas in question.

By relating these results to those addressed in question 1, I will investigate how prolonged deglaciation was in Torridon compared to the rate of change observed in palaeoecological proxies (e.g. Brooks and Birks, 2000) and the ice core record (Andersen et al., 2006; Svensson et al., 2006). If there is a regional disparity, this could indicate a degree of non-linearity in the glacial response to climate change. I will also investigate how numerical models of the LLS ice in NW Scotland (e.g. Hubbard, 1999; Golledge et al., 2008) compare with the empirical data collected during field-based and remotely sensed mapping exercises. Anomalies may imply that inequalities existed between relatively maritime and continental regions of NW Scotland, or that existing models are insufficient to provide a clear overview of climate and glacial dynamics during the LLS.
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Question 4: What was the climate like in Wester Ross during the most recent glacial advance (Loch Lomond Stadial) in the Younger Dryas (~12.9 – 11.5 ka)?

Extensive oceanic reorganisation during the Younger Dryas led to a cooling episode, which lasted ~1.4 ka. An ice cap formed over the main watershed in NW Scotland, and satellite ice fields formed in marginal regions such as Torridon and Applecross, thought to be characterised by a complete re-growth of ice in some regions, and a rejuvenation of remaining Lateglacial Interstadial ice in others (e.g. Ballantyne, 2010; Golledge, 2010). The unique and prominent glacial landforms created during this readvance in Torridon and Applecross reflect the sensitivity of the region to fluctuations in local climate and in particular, fluctuations in the North Atlantic Gulf Stream (Benn et al., 1992). The ELA can be used to compare and contrast glacial response to local topographical variation, but the relationship between ELA and the local climate is particularly useful in analysing glacial response to climate change and in particular, variations in precipitation and temperature (e.g. Sissons and Sutherland, 1976).

Reconstructions of palaeo-precipitation and temperature are estimated using an equation derived from a global, empirical P/T dataset (Ohmura et al., 1992), which should provide a finely-tuned record of the response of NW Scottish coastal, marginal ice fields to Younger Dryas (YD) climate change. However, the use of a global dataset smoothes out any local anomalies. The validity of this model in the context of the highly sensitive climate of NW Scotland is therefore also considered. Basal shear stress calculations provide an indication of thermal regime and glacier dynamics during the last glaciation, giving an insight into YD climate. The direct interaction of the glacier base with the landscape can be related to question 1, as the glacial geomorphological features of Wester Ross are the direct result of basal conditions at any given point.
Chapter 2:

Methods I: Field and Laboratory

Methods
Chapter 2. Methods I: Field and Laboratory methods

This chapter covers the methods used to compile an overview of the geomorphological characteristics and the timing of deglaciation in the Wester Ross region, focusing particularly on the Torridon area. The methods used (field mapping, mapping using remote sensing methods and ground-based LiDAR) to compile a detailed geomorphological map (displayed in chapter 3 and Appendix 2) are outlined in sections 2.1-2.3. These results and the methods used to achieve them will be assessed in the discussion (chapter 6) with the aim being to answer questions 1 and 2 (section 1.9). Geomorphological mapping will also provide the basis for glacial reconstruction, which is the focus of chapters 4 and 5. Section 2.4 focuses on the methods used to collect and process samples for cosmogenic $^{10}$Be dating, to provide a deglacial chronology for the Torridon and Applecross areas. The considerations taken into account while collecting these samples will also be outlined. These considerations are particularly important in the accurate presentation of results, which will be presented in chapter 3. These results should provide some answers to question 3 (section 1.9).

2.1. Field mapping methods

2.1.1. Research strategy

As outlined in section 1.6, geomorphological and sedimentological fieldwork studies have been previously undertaken in Wester Ross. In this instance it was deemed necessary to carry out further fieldwork, to be able to put observations from previous studies in context, to develop personal understanding of the glacial history of the region and to reach unbiased interpretation. It was also important to produce maps in line with contemporary views of glacial landforms.

Reconnaissance fieldwork in Torridon and Applecross was of vital importance to decision-making processes, with regards to the collection of samples for cosmogenic $^{10}$Be dating. Suitable sites were noted and assessed in terms of their potential to offer insight into the Lateglacial history of Wester Ross, and their reliability for accurate dating.
### Table 2.1. Glacial geomorphological landform identification criteria. Adapted from Hubbard and Glasser, (2005).

<table>
<thead>
<tr>
<th>Landform</th>
<th>Identification criteria</th>
<th>Boundary</th>
<th>Significance</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Ice scoured bedrock</strong></td>
<td>Widespread exposures of bare bedrock with smooth, striated or plucked upper surfaces</td>
<td>Outermost extent of bare bedrock</td>
<td>Evidence of glaciol ice at its pressure-melting point</td>
</tr>
<tr>
<td><strong>Drift limit</strong></td>
<td>The edge of the cover of a glacigenic deposit that is also marked by a moraine ridge. Drift limit may be identified by a change in vegetation type or by a change in the density of glacially deposited boulders</td>
<td>Outermost extent of glacigenic deposit</td>
<td>Drift limit marks the extent of a glacier advance</td>
</tr>
<tr>
<td><strong>Moraine ridge</strong></td>
<td>A single ridge or a collection of ridges composed of glacigenic material, typically 1-10 m (but exceptionally 100 m) in height. Sharp or rounded crests and either linear, curved, sinuous or sawtooth in plan.</td>
<td>Lower concave break of slope. The restline orientation can also be recorded if this is well defined</td>
<td>Moraine ridge marks the lateral or terminal extent of a glacier advance (except in the case of “hummocky moraine”)</td>
</tr>
<tr>
<td>“Hummocky moraine”</td>
<td>A seemingly chaotic assemblage of irregular hummocks and hollows. “Hummocky moraine” often shows order when viewed on aerial photographs, or when mapped in detail in the field.</td>
<td>Outermost extent of “hummocky moraine.” The crestline orientations of individual hummocks can also be recorded if the map scale permits</td>
<td>Formed by the slow melting of stagnant, debris covered ice, by deposition at receding glacier margins, or the release of material from proglacial or englacial thrusts</td>
</tr>
<tr>
<td><strong>Flutes/flutings</strong></td>
<td>Groups of straight, elongated ridges of glacigenic material. Typically 10-100 m in length and 1 m wide.</td>
<td>Upper convex and lower concave breaks of slope of individual features</td>
<td>Subglacial modification of sediment. Flutes are formed parallel to direction of flow</td>
</tr>
<tr>
<td><strong>Meltwater channels</strong></td>
<td>Channels cut into rock or sediment, often with abrupt inception and termination and lack of modern catchment.</td>
<td>Thalweg of channel, location of channel inception and termination. Arrow to indicate direction of former drainage can also be recorded if known</td>
<td>Evidence of former meltwater discharge routes</td>
</tr>
<tr>
<td><strong>Trimline</strong></td>
<td>Line separating areas of solifluction from extensive gullting or areas of mountain-top detritus/weathered material from</td>
<td>Lower limit of solifluction or weathering and upper limit of extensive gullying</td>
<td>Former vertical dimensions of a glacier or englacial</td>
</tr>
</tbody>
</table>
Chapter 2. Methods I: Field and Laboratory methods

This will be discussed further in section 2.4.4. As fieldwork progressed, it was clear that the geomorphological and sedimentological characteristics of Wester Ross could be studied in vast detail, outside the scope of this doctoral research. Therefore field-based geomorphological mapping became focussed on Torridon, in particular within Coire Mhic Nobuil (Figure 1.2), where glacial-geomorphological features were especially abundant and prominent. Reconnaissance surveys concluded that this valley would provide the clearest and most sensitive record of the final stages of the last deglaciation.

2.1.2. Collection of field evidence

Landform recognition was undertaken using my own experience of glacial environments, following trips to a variety of modern and ancient glacial environments in the Swiss Alps, the Basin and Range region of Idaho, USA, and the Svalbard archipelago. Identification criteria outlined in Hubbard and Glasser (2005) were also used as a comprehensive guide to the recognition of a series of glacial geomorphological features in the field. Table 2.1 is an adaptation of this guide.

Detailed geomorphological mapping was undertaken on foot in Coire na Caime and Coire Mhic Fhearchair (Figure 2.1). A combination of remotely sensed and field-based analysis was undertaken in this central region of Torridon when ground-truthing of remotely observed features was necessary. Landforms in Torridon were mapped using a GARMIN GPSmap 60CSx. With 5 m accuracy, this allowed some of the more subtle features (striae, moulded bedrock and small linear features) to be measured and located accurately. GPS readings were taken at points of interest, and where linear features were present; readings were taken at both ends of the feature, and at any obvious change in orientation. Where linearity was questionable or of particular interest, detailed descriptions and drawings were made in a field notebook. Extensive wet-heath cover, and a ~30 cm deep layer of clast-filled peat throughout much of Wester Ross, made it logistically difficult to assess the subsurface stratigraphy.
Figure 2.1. Mapping coverage, sedimentary exposure and cosmogenic isotope dating in the Coire Mhic Nobuil area of Torridon. The area mapped entirely on foot is coloured light grey and the darker grey covers the area mapped using a combination of aerial photograph interpretation and ground-truthing on foot. Cosmogenic sample sites are denoted by the crosses and exposure ages are provided.

For this reason, landform interpretation is based mainly on the geomorphological interpretations described in section 2.1.2. Some exposures of glacial sediments were found along pathways and streams, but often proved to be too small to provide a complete picture. Localities were plotted on base maps at a scale of 1:10,000 to enable the determination of moraine distribution, and imported into a GIS (ArcGIS). Waypoints were interpolated to define the boundaries of linear features and the orientations of striae were added following correction from true north to grid north.

Within Torridon only one exposure was deemed suitable for detailed study. This exposure is located at the junction of Coire Mhic Nobuil and Bealach a’ Chomhla, on the western bank of the stream emerging from Bealach a’ Chomhla (Figure 2.1).
The exposure was cleaned and enlarged, large-scale detailed drawings were made of the whole section, and particular areas of interest were sketched at a smaller scale. Individual units were defined based on grain size and shape, the nature of the matrix, sedimentary structures and compaction. Lithofacies codes from Evans and Benn (2004) were used for this purpose and can be seen in appendix 3.

2.1.3. Analysis of field evidence

The final geomorphological map, which can be seen in the results section (Chapter 3 and Appendix 2), was used as a basis for reconstruction. Only ridges with a linear form, i.e. flutes and concentric moraine ridges were used for the reconstruction of successive ice-fronts and palaeo-ice direction. These were distinguished from each other according to their orientation within the glacial valleys and corries. Fluted ridges were identified trending down-valley, at right angles to the concentric ridge chains that form the palaeo ice-fronts. Concentric ridges were connected to their conjugates on the opposite side of the valley, to form a stepwise pattern of deglaciation. Where bifurcations are obvious (i.e. differential retreat of the glacier front), two or more ridges have been drawn adjacent to each other. Flutes, striae, and roches moutoneés (mainly found in the high corries) were used as directional indicators.

Moraine asymmetry was assessed using the methods outlined in Benn (1989). Lateral-terminal moraines were isolated from the other geomorphological features in Coire Mhic Nobuil, Torridon and projected onto 1:10,000 basemaps in ArcMap. Ice margin shapefiles for Torridon (reconstructed in chapter 4) were also projected in ArcMap. Free faces (slopes of 40-45° or more) above the ice margin were identified and added to the map. Slope area was calculated using the average gradient (Equation 2.1).

\[
A_r = \frac{A_m}{\cos a}
\]

*Equation 2.1.*
Where $Ar$ is the real area of the slope segment, $Am$ is the mapped area of the slope gradient and $a$ is the average gradient calculated from map contours. By quantifying the total area covered by free faces on opposite sides of the valley, potential debris sources could be identified and compared with the area covered by lateral-terminal moraines in Coire Mhic Nobuil.
2.2. Remote Sensing

Remote sensing methods were combined with field-mapping and sedimentological investigations. This multi-disciplinary approach allowed an unbiased and accurate assessment of the glacial-geomorphological landscape. By combining remotely sensed and field data, individual features were assessed in terms of their recorded morphology in the field, and the way in which their morphology is reflected in aerial photography and DEMs. This enabled the accurate reconstruction of glacial geomorphology in regions of central Torridon where remote and rough terrain was logistically difficult to map on foot.

2.2.1. Aerial Photography

The interpretation of aerial photographs from the University of Cambridge Unit for Landscape Modelling, was undertaken in conjunction with field-mapping excursions. Digital copies of black and white aerial photographs at a scale of 1:25,000 (e.g. Figure 2.2) were imported into ArcGIS. In ArcMap, georeferencing was achieved by matching distinctive geomorphological features, (Ground Control Points; GCPs) to Ordnance Survey (OS) basemaps. A minimum of 10 GCPs were picked, to maximise the accuracy of georeferencing. These points were taken from a variety of locations, and from lowest and highest-altitude points on the images, to take into account distortion associated with different altitudes. The most reliable GCPs are generally man-made features such as railway lines, buildings and roads. In this instance, some of the GCPs were taken from footpaths, but an obvious lack of other man-made features meant that GCPs were taken from mainly from natural features, which are less reliable as they are subject to change. Mountain peaks and bedrock features were believed to be a fairly reliable location for GCPs, and water features such as lochs were only used where essential, due to variations in their shorelines.
Google Earth imagery from the 2009 image library, reconstructed from Infoterra coloured aerial photography (2004), was georeferenced to the aerial photographs and base-maps as an (30% transparent) overlay. This imagery added colour to the landscape and enhanced the clarity of individual glacial-geomorphological features.
Chapter 2. Methods I: Field and Laboratory methods

Figure 2.3. Aerial photograph from figure 2.2 georeferenced to an Ordnance Survey base map using Ground Control Points (GCPs).

Figure 2.4. Aerial photograph (Figure 42.2) overlay on Google Earth (2009) imagery. Areas of high resolution are shown in the top part of the imagery. The low-resolution area is shown at the bottom of the imagery in brown.

High-resolution Google earth imagery was not however available for the whole of the Torridon area as can be seen in Figure 2.4.
2.2.2. Digital Elevation Models (DEMs)

A NextMap Great Britain™ Digital Surface Model (DSM) at 10m resolution can be seen in figure 2.5, draped with an OS base-map, with an aerial photograph superimposed. When studying small areas in detail, resolution was increased to up to 5m to allow for more detailed study of geomorphological features. DEMs are particularly useful when investigating the glacial geomorphology of vast areas. NEXTMap Great Britain™ is based on radar technology, thought to provide the most realistic approximation of field mapping at a 1:10,000 scale, compared to other remote sensing datasets (Smith et al., 2006). It has been used in the BRITICE project (Clark et al., 2004) as a basis (control) for comparing mapping from various projects throughout the British Isles.

NEXTMap Great Britain™ tiles were downloaded from www.neodc.rl.ac.uk, as DSM tiles and imported into Arc GIS. The relevant tiles were stitched together and clipped to size in Arc. Post-processing took place in ArcScene and ArcMap.
2.3. Ground-based Light Detection and Ranging

Light Detection and Ranging (LiDAR) is a highly accurate tool for the acquisition of 3D spatial data. It is an optical remote sensing technology that uses the properties of scattered light to measure the range of a distant target. LiDAR is used in airborne or ground-based (terrestrial) operations and has been widely employed in a number of scientific disciplines, including geological studies (e.g. Carabajal et al., 2007 and Bates et al., 2008); atmospheric science (e.g. Cook et al., 2007); planetary science (e.g. Baize et al., 2005) and oceanography (e.g. Lee et al., 2005). Non-scientific applications include surveying, law enforcement and construction. Since its initial applications in the 1960s (e.g. Schuster, 1970) the precision of LiDAR data has improved from several metres to a few millimetres (Bellian et al., 2005), depending on the scanner used and the distances it is used over. Ground-based LiDAR has been used here to create a 3D ‘virtual reality’ model of Coire Mhic Fhearchair in central Torridon.

2.3.1. The RIEGL LMS-Z420i scanner

The ground-based LiDAR scanner used in this instance was the RIEGL LMS-Z420i (Figure 2.6). The scanner is capable of scanning 12,000 points per second, with a range of 800m and 80º vertical and 360º horizontal fields of view. A light laser emitted by the scanner is backscattered from the target object and is recaptured by the scanner. This two-way travel time is halved and multiplied by the speed of light to derive an accurate z distance. The x (latitude) and y (longitude) positions are calculated by the mirrors within the scanner, and depend on the relative positions of the pulse when it leaves and returns (Bellian et al., 2005). LiDAR scanners such as this are capable of scanning many thousands of x,y,z points per second, and are therefore capable of scanning large areas in short spaces of time. Scanning density can be adapted according to the required resolution, the amount of time available for scanning and the capacity of the hard-drive for storage of data.
A Nikon D100 Digital SLR camera is mounted above the scanner (Figure 2.6), and provides high-resolution colour images to provide colour and texture for the scans. Photographs are taken automatically to coincide with the scan data through a controlling laptop.

Georeferencing of the LiDAR data is undertaken by a Trimble Differential Global Positioning System (DGPS), which is also mounted upon the scanner head. Global coverage and sub-metre scale accuracy allows the accurate mapping of scan positions. The LiDAR system is controlled by a Panasonic Toughbook laptop. This laptop has been modified for fieldwork, incorporating a touch screen, which allows successful viewing even with direct sunlight. The laptop runs software to initiate and control the scan, including the digital camera and DGPS. It also acts as a portable data storage device to record the scan data. The whole system is powered by modified car batteries.
2.3.2. The LiDAR expedition

In a pioneering trial, LiDAR was used to assess the glacial geomorphology in a corrie in the centre of Torridon. LiDAR data were collected during a three-day expedition to Coire Mhic Fhearchair in July 2007. Issues with precipitation and mist during this time hampered attempts to collect data on several occasions, and prevented the scanning of the very top of the corrie walls. However, during weather
windows, nine panoramic (~200 mm-spatial resolution) and five detailed (~50 mm) scans were chosen according to the presence of glacial geomorphological features of interest (i.e. distinctive features which could provide information on glacial history, whilst also testing the capabilities of LiDAR for this purpose). In this instance, 50 mm was chosen in order to produce a dataset of a manageable size within the time frame of the expedition. Where a particular area of the corrie was of interest, several scans were taken from different perspectives, to account for undulations in the ground surface and shadowing.

2.3.3. Post-processing

The individual scans were geo-referenced using DGPS data and overlapping scans were merged by matching like points to create a unified coordinate system in PolyWorks. Point clouds from the 9 panoramic scans were imported into PolyWorks, to act as a matrix. Corresponding points were picked using the n-point pair alignment function, and were automatically aligned using the best-fit function, that uses a ‘least squares’ algorithm, to achieve a statistical best-fit between two scans. The 9 panoramas were then locked together to provide a solid framework for the rest of the project. DGPS data was then imported into PolyWorks. A reference point was assigned to the 9 scan stations, and the z value of each was offset to accommodate the height difference of the scanner and the DGPS receiver. The auto-match function was used to match the scanner and DGPS coordinates, and georeferencing was complete. The 5 higher resolution scans were subsequently georeferenced using the coordinate framework provided by the panoramic scans. A 3D surface was created from the point cloud using the triangulation capabilities of the software package RiSCAN Pro. Once triangulation has taken place, a mesh is formed, which is composed of hundreds of thousands of individual points. Each individual pixel on a digital photograph was linked to its x,y,z coordinate within the triangular mesh. The addition of digital photographs to this mesh effectively produces a high-resolution photo-textured model. In many instances, shadows were an issue, particularly on the eastern side of the corrie. This was due to time constraints and a lack of battery power during the 3 day excursion. This was
particularly obvious where triangulation had joined points which were widely spread, creating an unrealistic surface. Therefore a different approach was taken. In RiSCAN Pro, the area covered by the scans was filtered into individual cubes with specified edge lengths (50 mm) and a representative data point for each cube was used to create an ‘octree’. Digital colour photography was subsequently added to form a true colour image (Figure 2.7). Analysis of glacial geomorphological features (i.e. measurement of height, length width) was carried out on-screen where features of interest could be observed from various angles and using intensity and false colour functions in RiSCAN. At this stage, the suitability of this particular digital model could also be assessed in terms of its suitability for glacial geomorphological mapping.

2.3.4. Ground-based LiDAR case study: Coire Mhic Fhearchair, Torridon

Ground-based LiDAR has been compared to existing mapping techniques to assess its validity as a geomorphological mapping tool. Existing techniques include field mapping, aerial photography, and DEMs. Coire Mhic Fhearchair (see figure 1.2 for location) was chosen as a suitable site on this occasion due to the variety and clarity of the geomorphological features within this corrie. For this reason, each of the mapping techniques could be assessed in terms of the ease of observation of geomorphological features, the ability to pick out features at a variety of resolutions and the ability to study an individual feature using the given technique.
2.4. Cosmogenic isotope dating

In the context of this research, cosmogenic isotope dating is utilized to provide a chronological framework for the last deglaciation of Wester Ross. Through the quantification of $^{10}$Be concentrations in quartz, and with knowledge of the production rate and half-life of $^{10}$Be, we can measure how long a surface has been exposed to cosmic rays, and as a result, we can calculate its exposure age.

2.4.1. Cosmogenic $^{10}$Be isotope production

When high-energy cosmic radiation enters the upper atmosphere, a plethora of secondary particles (mainly neutrons) are produced. Cosmogenic isotopes are produced in situ, via spallation and muon capture reactions when minerals in rocks are hit by these particles. Spallation is the primary source for $^{10}$Be production within the atmosphere and quartz grains. Within quartz grains, a nuclear reaction takes place, during which the nucleus of an oxygen atom within a quartz grain is bombarded and broken down by high-energy particles, liberating protons, producing daughter nucleons, and reducing atomic weight.

Atmospheric $^{10}$Be is carried to the Earth’s surface via precipitation, and is considered as a contaminant within samples used for cosmogenic analysis of bedrock (Everest, 2003), although recently it has been applied to the study of sediment transport and deforestation (e.g. Reusser and Bierman, 2010). Muon capture accounts for $\sim$3% of $^{10}$Be production (Stone, 2000), and occurs in similar locations to spallation reactions. Muons are captured by nuclei, which then decay to form stable isotopes such as $^{10}$Be. Isotopes produced within exposed quartz grains; $^{36}$Cl, $^{10}$Be and $^{26}$Al are all suitable for cosmogenic analysis due to their long half-lives (Child et al., 2000). However, $^{10}$Be has the longest half life (1.5 Myr), which makes it ideal for the study of processes occurring on glacial timescales. Beyond 5 Ma, the measurement of $^{10}$Be is no longer effective (Ivy-Ochs, 1996).
2.4.2. Sampling considerations

When collecting bedrock samples for cosmogenic exposure ages, a number of factors were taken into consideration, to maximize the reliability of the results. These considerations are also illustrated in Figure 2.8.

- **Decrease in isotope concentration with depth:** Cosmogenic isotope production rate decreases with depth in bedrock (Gosse and Phillips, 2001, Fabel et al., 2002). Sites characterised by ice-erosional features such as scouring and plucking were chosen to enhance the reliability of the exposure ages. Where striae were present, it could be assumed that erosion since glaciation was minimal. Surfaces which appeared more weathered were sampled with the understanding that non-erosive (cold-based) ice cover may lead to an anomalously young exposure age. To account for the decrease in $^{10}$Be concentration with depth only the top 2 cm of rock were removed. Measurements of protruding quartz granules were also
taken to assess the extent of post-glacial weathering at individual sites. The steps taken to do this and how this was accounted for in the final calculations is outlined in section 2.4.6.

- **Potential for intermittent cover:** Studies by Lal (1991) imply that $^{10}$Be production rate decreases exponentially with depth in rock. Masarik and Reedy (1995) however, assume that the effects of shielding are negligible above the top $\sim$4.6 cm (in granite), and suggest that corrections for production rate are only need below this level. This correction also accounts for intermittent cover of the surface. Careful assessments were therefore made at each site with regard to the proximity of the sample site to steep slopes and associated rock-fall, and the likelihood of sediment, vegetation or snow cover following deglaciation. Exposed and upstanding sites were selected to minimise the effects of intermittent cover by snow, ice or vegetation during the Holocene, and to negate the need to account for shielding of the rock surface. Estimates of palaeo-snow coverage can be very subjective and vegetational histories can be incomplete, all of which could have obscured the bedrock surface from cosmic rays for periods of time, thus providing an age which does not reflect constant exposure since deglaciation. Prolonged or intermittent cover of the surface in question would have result in the underestimation of the exposure time since deglaciation.

- **Topographic shielding:** Where a sample site is surrounded by completely open sky, a value of zero topographic shielding would be applied. In the instance of Torridon and Applecross, obstructions exist in the form of undulations in the surface surrounding the sample site, high peaks and corrie walls. Corrections were therefore made according to how much of the horizon was obscured by the surrounding terrain. In this case, the angle of elevation of the horizon was measured every for 30º of the azimuth. Topographic shielding values for each of the sites can be seen in appendix 4.
2.4.3. Schmidt hammer tests

Schmidt hammer values provide an indication of the relative weathering of a surface, and are therefore extremely useful in the analysis of post-glacial landscapes, particularly in conjunction with cosmogenic isotope dating. A Schmidt Hammer records the percentage rebound or coefficient of restitution of a spring loaded mass impacting with constant force on a bedrock surface (Ballantyne, 1986). Highly weathered, frost-shattered surfaces have a lower elasticity due to the presence of deep joints and fractures. Glacially-scoured surfaces tend to have a higher elasticity, due to their homogenous nature and shorter exposure history, thus yielding higher Schmidt Hammer values than relatively weathered surfaces. Ballantyne et al., (1998) suggest that this test is not suitable on Lewisian Gneiss and Moine schists, as the lithology of these rocks varies over short distances. More reliable are measurements made on the more consistent Torridonian sandstones and Cambrian quartzites.

Fifty Schmidt Hammer measurements were taken at each sampling site (Appendix 5). Statistical analysis, including mean and standard deviation calculation was undertaken on each set of data. Where two adjacent sampling sites were characterised by contrasts in weathering, t-testing was undertaken out to assess the significance of the difference. Equations 2.2, 2.3 and 2.4 outline the steps taken to achieve these results. The results of these calculations can be seen in chapter 3 and will be discussed further in chapter 6.

\[
S_p = \sqrt{s_a^2(N_a - 1) + s_b^2(N_b - 1)} \quad \text{Equation 2.2}
\]

\[
t_{\Delta x} = \frac{x_a - x_b}{S_p \sqrt{\frac{1}{N_a} + \frac{1}{N_b}}} \quad \text{Equation 2.3}
\]
Where $t_{x}$ is the test statistic, $N_a$ and $N_b$ are the sample sizes of the sets believed to have a lower and higher Schmidt hammer readings respectively, $x_a$ and $x_b$ are their means, $S_a$ and $S_b$ are their standard deviations. The aim of this test is to establish whether the Schmidt hammer values (and therefore the degree of weathering) below the trimline (sample set a) are significantly higher than those above the trimline (sample set a). If the means of the two samples are $x_a$ and $x_b$, the null hypothesis ($H_0$) may be stated thus: "$x_a$ is not significantly larger than $x_b$." The alternative hypothesis ($H_a$) therefore reads "$x_a$ is significantly larger than $x_b$." To establish which hypothesis to accept, the t statistic or "critical-t" or margin of error is calculated by establishing the degrees of freedom (the number in the sample -1; in this case 50-1=49). Using the level of significance (0.05 for a two-tailed test) and the degrees of freedom and the number of tails, the critical-t is established using the TCRIT function in Microsoft Excel.

If tcrit (the result of equation 2.3) is greater than tcalc (equation 2.2), we accept the null hypothesis. If tcalc is greater than tcrit, we accept the alternative hypothesis. In this instance, acceptance of the alternative hypothesis would imply that the mean of the second sample set is significantly higher than the first.

**2.4.4. Sampling locations and justifications**

Eight samples were collected from Torridonian sandstone bedrock for cosmogenic $^{10}$Be exposure dating. Four of these samples (WRC 1a, 1b, 2 and 3) were collected from sites in central Torridon and the remaining samples (WRC 4, 5, 6, and 7) were collected from sites around the central breached area of the Applecross plateau. As well as the considerations listed above, samples were collected in locations that would yield vital information about the deglaciation history of the Torridon and Applecross areas. Samples were removed from their various locations using a heavy mallet and sharp chisel. At each site, care was taken to remove only the top 2 cm of bedrock. Fifty Schmidt Hammer readings were taken at each sample site and from the surrounding surfaces.
Figure 2.9. Locations of the cosmogenic sample sites within the LLR ice limits in Torridon. (a) A roche moutonnée on the lip of Coire Mhic Fhearchair (Figure 2.1) from which WRC1a and b were collected. The dashed white line shows the estimated area which has been quarried from the lee side of the roche moutonnée, indicating that ice movement was from right to left (south to north). (b) The low-altitude trimline at 500 m on Am Beacan (Figure 2.1). Glacially scoured bedrock below is separated from weathered bedrock above. (c) The WRC2 sample site on glacially-scoured bedrock below the trimline. Photograph taken facing SE. (d) The WRC3 sample site on weathered bedrock above the trimline. Coire Mhic Fhearchair can be seen in the background, to the south of Am Beacan.

Topographic shielding measurements were taken with a compass every 30° in a 360° panorama of the surroundings. Photographs were taken from all angles of each sampling site and detailed descriptions of the surroundings were recorded.

**WRC1a and WRC1b:** WRC1a and 1b were collected from a Torridonian sandstone roche moutonnée on the lip of Coire Mhic Fhearchair (Figure 2.9a). This feature has a characteristic asymmetry indicating that ice flow was from S-N (right to left on Figure 2.9a). WRC1a was taken from the glacially-smoothed stoss face, which was characterized by a relatively weathered surface, with quartz granules protruding out by a maximum of 20 mm. WRC1b was taken from a relatively less weathered quarried surface on the lee face.
**Justification:** To test whether the lee face of the roche moutonnée was created via plucking mechanisms during the YD. If this were the case, a significant difference in age between WRC1a and 1b (1b being younger), would provide an indication of when the lee face was quarried. If there is no significant difference between WRC1a and 1b, and both sites are YD in age, this would imply that at least 2 metres of erosion took place during the YD at both WR1a and 1b, thus resetting the cosmogenic clock for both sites. Any lesser amount of erosion would result in the retention of inherited cosmogenic isotopes at the time of deglaciation, producing an anomalously old (i.e. pre YD) exposure age. A much older exposure age would therefore imply a negligible amount of YD erosion. In this instance we can gain insight into the erosional and, by inference, basal thermal regimes of the last glacier to reside in the area.

**WRC2 and WRC3:** WRC2 and WRC3 were taken from below and above (respectively) an obvious trimline (Figure 2.9b) in Torridonian sandstone bedrock at ~500m altitude on Am Beacan, Torridon (Figure 1.2), approximately 2 km north and across-valley from the lip of Coire Mhic Fhearchair. WRC2 was collected from a raised ice-polished bedrock platform (Figure 2.9c), where glacial striae and quarried surfaces indicate dominant SE-NW ice movement. WRC3 (Figure 2.9d) is characterised by a relatively weathered surface, but evidence of glacial quarrying, and the presence of erratics of Lewisian Gneiss imply that it was covered by ice during a less recent, yet more extensive event. This sample was taken from a site 111 m higher than WRC2 and 26 m above the trimline.

**Justification:** These samples were collected to provide a basic chronology for the trimline. Significantly different dates above and below it would explain the weathering contrast and would indicate glacial events of varying extent. A YD age for WRC2 and pre-YD exposure age for the site at WRC3 would imply that the trimline defines the limit of the LLS. A Lateglacial interstadial date for WRC3 would have implications for the survival of ice in this region, providing an estimation of when ice was retreating away from Torridon.
Figure 6.7. Locations of the cosmogenic sample sites on the Applecross Peninsula. (a) The WRC4 sample site on the glacially-scoured Bealach nan Arr breach between Coire nan Arr and Coire Attadale (Fig. 4). (b) The WRC5 sample site, ~50 m from the WRC4 site on the Bealach nan Arr Breach. The sample was taken from a step in the bedrock to test for signs of inheritance. (c) The WRC6 sample site on the highly weathered knoll, 50 m above and slightly to the west of WRC4. (d) Glacially scoured surface with pressure release cracks (Boulton, 1979) from which WRC7 was taken.

WRC 4, 5 and 7 are taken from glacially-scoured breaches and WRC 6 is taken from a heavily weathered knoll.
**WRC4 and WRC5:** Unlike Am Beacan in Torridon, there is no obvious trimline in Applecross, therefore samples were collected from sites where the weathering contrast was most obvious, i.e. the most scoured (WRC4, WRC5) and the most weathered surfaces (WRC6). WRC4 was collected from the lowest point on the Bealach nan Arr breach, in an area of extensive glacial scouring (Figure 2.10 and 2.11a). WRC5 was taken from a bedrock step (Figure 2.10 and 2.11b), approximately 200 m from WRC4, 1 m below the top of the step and 40 cm above a vegetated surface.

**Justification:** These samples were collected to test for signs of inheritance and to assess the extent of erosion by the last glacial event. Statistically similar exposure ages at WRC4 and WRC5 would imply that over 2 m of erosion took place across the breach. As discussed previously, any lesser amount of erosion would result in an anomalously old age. Statistically different exposure ages would indicate that differential erosion took place across the breach, or given that both samples were taken at a very similar altitude, it would be more likely that an error exists in the dating technique. These samples were also collected to test Robinson (1977) and Jones' (1998) empirical ice-extents. A YD age for the breach would imply that Robinson's more extensive LLS glaciation in Applecross is a more realistic situation.

**WRC6:** WRC6 was taken from a highly weathered and frost shattered area at the highest point on the knoll ~50 m above and to the west of the breach. The top of the knoll is covered in boulders, some are up to 3 m$^2$, and are perched on top of small boulders and cobbles (Figure 2.10 and 2.11c).

**Justification:** Compared to the breached area to the east, this area has been subject to a high level of weathering, and is therefore likely to have been subject to a longer exposure history. WRC6 was collected to test for this, and also to provide an approximate age of exposure for this area (and similar areas) of high ground in Applecross during deglaciation.
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**WRC7:** WRC7 was collected from the area to the south of the knoll and at the top of the over-steepened wall of Coire Nan Arr. This ledge is characterised by erosional landforms which will be investigated further in chapter 3.

**Justification:** WRC7 was collected from this area (Figure 2.10 and 2.11d), to address the issue of the extent of the LLS ice field in Applecross, interpreted differently by Robinson (1977) and Jones (1998) (Figure 1.27), and also to gain insight into the local subglacial thermal regime of LLS ice bodies. A YD exposure age for this ledge would again indicate that Robinson's model of a more extensive LLS glaciation is more realistic than Jones’ suggestion that this area was covered by non-erosive, cold-based ice.

2.4.5. **Sample Processing**

To enable accurate measurement of $^{10}$Be by Accelerator Mass Spectrometry (AMS), a number of conditions must first be met. The target must be pure, in-situ $^{10}$Be, and therefore all contaminants must be removed in the sequence outlined below. Throughout the whole process, laboratory protocols were followed to prevent contamination and to avoid accidents when using strong acids.

The first step of sample preparation (quartz separation and purification) took place at the Department of Geographical and Earth Sciences, the University of Glasgow using procedures modified from Kohl and Nishiizumi (1992).

**Sample preparation:** 2 kg of each sample was crushed and sieved to isolate the 250-500 µm fraction. Fractions >500 µm would be too large to break down effectively during separation, and <250 µm would be likely to suffer a greater percentage-loss of quartz during separation. The 250-500 µm fraction was rinsed in an aluminium dish, to remove any finer grains, and was subsequently dried in a drying oven.
Carbonate and metal removal (HCl/HNO₃ leaching): Enough 10% HCl/HNO₃ (hydrochloric/nitric acid) was poured onto the sample to cover it. It was then covered with a watch glass and left overnight on a low hotplate setting. Following disposal of the acid, the sample was washed several times and was then dried in the sample oven overnight at ~90ºC. This stage removes any metals present following the crushing and grinding process, and any carbonate contaminants.

Flotation: The frothing technique was undertaken to remove feldspar and mica. The sample was pre-treated in a solution of 1% hydrofluoric (HF) acid, which altered the surface chemistry of the feldspar, mica and quartz grains, making the feldspar and mica hydrophobic and the quartz hydrophilic. After an hour, the HF was decanted and the mineral mixture was combined with a few drops of pure eucalyptus oil and a carbonated solution of 1 ml/L of acetic acid and 1 g/L of Lauryl amine in 1 L pure water. After a few minutes, the frothy top layer of the mixture was removed, and along with it, the feldspar and mica. The remaining quartz and heavy mineral mixture was left to dry in the oven.

Magnetic separation: A Frantz magnetic separator was used to remove magnetic minerals such as biotite minerals from each sample. Each sample was added to the separator, which was set at 0.1 mA. Magnetic and non-magnetic fractions were collected, the separator was cleaned out, and the non-magnetic fraction was returned, but this time, to a separator set at a magnetic field strength of 0.5 mA. This process was replicated with the magnetic field set at maximum, and was repeated until the magnetic fraction was removed.

Phosphoric leaching: The purpose of this stage was to remove aluminosilicates from the sample. To do this, 40 g of the sample was added to a 1000 ml round flask along with 400 ml of 85% ortho-phosphoric acid and was heated in a heating mantle to 240ºC. This temperature was maintained until the acid converted to pyrophosphoric acid at 220 ºC. After 30-60 minutes of acid digestion, the mixture was left to cool and the acid was carefully decanted into a beaker of cold tap water. To remove the
gelatinized silica solution in the sample, 200 ml of water and 200 ml of 50% sodium hydroxide solution were added and the sample and solution were boiled for ~ 10 minutes. The solution was safely disposed of and the remaining sample was dried in a beaker. At this stage, the pyrophosphoric acid had dissolved most of the aluminosilicates in the sample.

**Removal of feldspar and meteoric $^{10}$Be (HF leaching):** 60 g of the remaining samples were each added to 500 ml polypropylene bottles and covered with a 2% HF (hydrofluoric acid) solution to within 2 cm of the top. The bottles were tightly sealed and inverted several times to mix the contents. The bottles were then placed in an ultrasonic bath for 72 hours, and were shaken several times during this period. After 72 hours, the bottles were left to cool, the HF was safely discarded and the samples were rinsed with de-ionised water. The process of etching with HF was repeated two more times. As well as dissolving silicates such as feldspars, the HF leach removed the outer shell of quartz grains; the part of the grain most likely to be contaminated by meteoric $^{10}$Be.

**Determination of quartz purity:** It is essential to obtain the lowest possible Al concentration in the sample to maximise accuracy. Acceptable levels of Al are between 10–100 ppm and anything higher than this can indicate the presence of impurities such as feldspar and muscovite. To test for evidence aluminium content, 0.3–0.6 g of each sample was placed in a Teflon round-bottomed vial with 10 ml of concentrated HF (40–50%) and 1–2 drops of 1:1 H$_2$SO$_4$ (sulphuric acid). Quartz was allowed to dissolve overnight in capped vials. The HF/ H$_2$SO$_4$ solution was allowed to fume off under a fume hood whilst sat on a 120°C hotplate. These procedures were repeated if any quartz remained in the vials.

Once the quartz was fully dissolved, the vials were cooled and the dry residue was dissolved in 8 ml of 2% HNO$_3$. The vials were capped and left to stand overnight again, for the remaining fluorides to dissolve. The vials were then inverted several times to homogenise the solution, and were weighed to record any change in mass.
If the samples appeared pure, i.e. uniform in appearance and not caking the floor of
the bottle, the samples were rinsed in de-ionised water, dried out, cooled, and
transferred to labelled bags. Samples were decanted into centrifuge tubes and were
sent for analysis by ICP-OES (Inductively Coupled Plasma optical emission
spectrometer) analysis to determine sample purity. If samples returned with a >100
ppm concentration, the process would be repeated once more.

The second stage of the process was carried out in the Glasgow University
Cosmogenic Isotope Laboratory housed at the Scottish Universities Environmental
Research Centre (SUERC) in East Kilbride, using methods from Child et al., (2000).

**Carrier addition and sample digestion:** The aim of this second stage is to extract Be
from the quartz and to measure the $^9$Be/$^{10}$Be ratio using AMS. However as Be is a
very rare element in quartz, the sample is spiked with a known quantity of $^9$Be. The
ideal carrier has not been exposed to cosmic radiation and therefore has very low
concentrations of $^{10}$Be. In this instance the carrier was derived from a sample of
Archean beryl from a deep gold-mine in Australia.

**Anion exchange:** Anion exchange columns were used to remove any remaining Fe
and Ti from the sample. Anion exchange resin was added to ion exchange columns
and 10 ml 1.2M HCl, followed by 10 ml 6 N HCl solutions was washed through the
resin. 2 ml of 6 N HCl was also added to the sample and stirred through. The Fe
(III) formed a range of Cl$^-$ complexes in the presence of strong HCl, which binded to
the anion exchange resin, could be seen collecting and forming a brown stain at the
top of the column. Be and Al (and some Ti) did not form the same bonds, and
washed straight through the resin when a further 6 ml of 6 N HCl was added.

**Cation exchange:** Cation exchange removed any remaining Ti and ultimately
separated the remaining Be and Al, ready for AMS analysis. Cation resin was
slurried into ion exchange columns, which were subsequently conditioned with 10
ml 0.2 M H$_2$SO$_4$. Sample solutions (the sample in 0.2 M H$_2$SO$_4$) were dripped
slowly into the columns. 1ml of 0.5 M H$_2$SO$_4$ was added to the columns to remove
the remaining Ti. The Be was subsequently eluted into a separate bottle by adding 10 ml 1.2 N HCl, and finally the Al was removed by adding 6 ml of 6 M HCl.

**Centrifuging, drying and oxidation:** The Be and Al solutions were transferred to labelled centrifuge tubes and using a 1:1 NaOH solution, were brought to a pH of 8 to precipitate Be and Al hydroxides. Following a few hours of precipitation, the tubes were centrifuged at 3200 rpm for ~10 minutes, then left to dry and cool. The Be(OH)$_2$ and Al(ON)$_3$ precipitate pellets were added to a designated (and carefully weighed) quartz crucible. The crucibles were transferred to a furnace where they were left for 2 hours at 1000ºC. Once cooled, the crucibles were weighed and the initial weight was subtracted to calculate the weight of the oxides. The crucibles were then packaged, and sent to the AMS (Accelerator Mass Spectrometer) to be pressed into targets and analysed. Beryllium isotope ratios in the samples and a procedural blank were measured at the SUERC AMS Laboratory.

**2.4.6. Calculation of exposure ages**

The accurate calculation of $^{10}$Be production rates requires the consideration of a number of factors, including latitude, altitude and atmospheric pressure and variations in the Earth’s magnetic field, all of which vary according to the situation of the sample. These affect the exposure age calculation and therefore need to be accounted for before they are incorporated in the final equations. A variety of production rate scaling schemes exist (Table 2.2), each giving a different exposure age for the same site (Balco *et al*., 2008), as latitude and altitude, and therefore cosmogenic $^{10}$Be production rates vary between samples.
Table 2.2. List of scaling schemes for spallogenic production. Adapted from Balco et al. (2008)

<table>
<thead>
<tr>
<th>ID</th>
<th>Reference</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>De</td>
<td>Desilets et al. (2006)</td>
<td>The scaling factor is a function of cutoff rigidity and atmospheric pressure. Production rates vary over time, according to changes in the Earth’s magnetic field.</td>
</tr>
<tr>
<td>Du</td>
<td>Dunai (2001)</td>
<td>The scaling factor is a function of cutoff rigidity and atmospheric pressure. Production rates vary over time, according to changes in the Earth’s magnetic field.</td>
</tr>
<tr>
<td>Li</td>
<td>Lifton et al. (2005)</td>
<td>The scaling factor is a function of cutoff rigidity, atmospheric pressure, and a solar modulation parameter. Production rates vary according to changes in solar output as well as changes in the Earth’s magnetic field.</td>
</tr>
</tbody>
</table>

The atmosphere acts as a partial shield to incoming radiation, which increases with distance from the source. Rates are therefore adjusted using a calibrated value determined for the situation of a sample. Lal (1991) employs a scaling function, which is based on a mathematical model, relating atmospheric pressure to altitude. $^{10}\text{Be}$ production rates are further complicated by the inter-annual variation of atmospheric pressure with location. It is not yet known how these anomalies affect the overall production rate when averaged-out over timescales of $10^3$ to $10^5$ years. Stone (2000) has studied air pressure variation in Scotland from records of the last 50 years, and suggests that latitudes 54 – 58° N experience lower pressure, and therefore $^{10}\text{Be}$ production rates at sea level are 2% higher than average. The location of this study in Scotland makes the Lm scheme a particularly appropriate scaling scheme for this study. This production rate is scaled according to spallation, with a 3% muon capture contribution. Through this scaling, the production rate has been
estimated to be $4.39 \pm 0.37$ atoms g$^{-1}$/yr$^{-1}$. The Lm production rate scheme also includes the variation of production rates with time due to changes in the Earth’s magnetic field strength according to Nishiizumi et al. (1989), and is currently thought to be the most reliable. This scheme has also been employed by similar studies in NW Scotland (e.g. Bradwell et al., 2008; Ballantyne et al., 2009), and the results can therefore be used as a direct comparison, negating the need for recalculation.

The equation used to calculate an exposure age (Equation 2.4) includes the considerations that have been referred to throughout this chapter. Latitude, altitude, pressure, sample thickness, topographic shielding, potential erosion rate, laboratory measurements of $^{10}$Be concentration and the internal error is calculated for each sample (and site) before this final calculation can be made. These parameters are essential to ensure an accurate result. Scaling scheme factors are also accounted for.

$$N = S_{thick} S_G P_{ref,sp,Xx} \int_0^T S_{Xx}(t) \exp(-\lambda t) \exp\left(\frac{-\varepsilon t}{\Lambda}\right) dt$$

$$+ P_{\mu} \int_0^T \exp(-\lambda t) \exp\left(\frac{-\varepsilon t - z/2}{\Lambda_{\mu}}\right) dt$$

_Equation 2.4_

Where:

- $N$ is the nuclide concentration in atoms g$^{-1}$;
- $S_{thick}$ is a non-dimensional thickness correction;
- $S_G$ is a non-dimensional geometric shielding correction;
- $P_{ref,sp,Xx}$ is the reference production rate due to spallation for the scaling scheme Xx (i.e. St, De, Du, Li or Lm) in atoms g$^{-1}$ yr$^{-1}$;
- $S_{Xx}(t)$ is the non-dimensional scaling factor for scaling scheme Xx, which may or may not be variable over time, according to the scaling scheme;
- $\lambda$ is the decay constant for $^{10}$Be (yr$^{-1}$);
- $\varepsilon$ is an independently determined exposure rate (g cm$^{-2}$ yr$^{-1}$);
- $\Lambda_{sp}$ is the effective attenuation length for spallogenic production (g cm$^{-2}$);
- $P_{\mu}$ is the surface production rate in the sample due to muons (atoms g$^{-1}$ yr$^{-1}$);
- $z$ is the sample thickness (g cm$^{-2}$) and $\Lambda_{\mu}$ is an effective attenuation length for production by muons (g cm$^{-2}$).
Chapter 2. Methods I: Field and Laboratory methods

Table 2.3. Parameters for entry into the CRONUS (MATLAB) calculator (Version 2.2.1): ‘Pressure flag’ specifies how to treat the elevation/pressure value. ‘std’ is entered here if elevations are in metres and the standard atmosphere is applicable the site (locations outside Antarctica). ‘Shielding factor’ refers to the amount of topographic shielding around the sample site. For samples with no topographic shielding, 1 is entered. For shielded sites, a number between 0 and 1 is entered. The ‘Erosion rate’ is inferred from independent evidence. In this instance only N10 (10Be concentrations) and N10 errors are entered. As the samples are not being tested for N26 (26Al concentration), a ‘0’ has been entered into the final two columns.

<table>
<thead>
<tr>
<th>Lab ID</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Elevation (m)</th>
<th>Density (g cm⁻³)</th>
<th>Shielding factor</th>
<th>Erosion Rate (cm yr⁻¹)</th>
<th>N10</th>
<th>N10 error</th>
<th>N26</th>
<th>N26 error</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRC-1a</td>
<td>57.593488</td>
<td>-5.447982</td>
<td>578</td>
<td>std</td>
<td>2.70</td>
<td>0.9602</td>
<td>0</td>
<td>100808</td>
<td>3511</td>
<td>0</td>
</tr>
<tr>
<td>WRC-1b</td>
<td>57.593488</td>
<td>-5.447982</td>
<td>577</td>
<td>std</td>
<td>2.70</td>
<td>0.9602</td>
<td>0</td>
<td>95145</td>
<td>3427</td>
<td>0</td>
</tr>
<tr>
<td>WRC-2</td>
<td>57.609094</td>
<td>-5.467938</td>
<td>415</td>
<td>std</td>
<td>2.70</td>
<td>0.9964</td>
<td>0</td>
<td>87049</td>
<td>3163</td>
<td>0</td>
</tr>
<tr>
<td>WRC-3</td>
<td>57.611502</td>
<td>-5.465049</td>
<td>526</td>
<td>std</td>
<td>2.70</td>
<td>0.988</td>
<td>0</td>
<td>109483</td>
<td>3959</td>
<td>0</td>
</tr>
<tr>
<td>WRC-4</td>
<td>57.437781</td>
<td>-5.69019</td>
<td>595</td>
<td>std</td>
<td>2.70</td>
<td>0.9988</td>
<td>0</td>
<td>103309</td>
<td>4571</td>
<td>0</td>
</tr>
<tr>
<td>WRC-5</td>
<td>57.437767</td>
<td>-5.667096</td>
<td>602</td>
<td>std</td>
<td>2.70</td>
<td>0.9997</td>
<td>0</td>
<td>102893</td>
<td>3462</td>
<td>0</td>
</tr>
<tr>
<td>WRC-6</td>
<td>57.437767</td>
<td>-5.631666</td>
<td>648</td>
<td>std</td>
<td>2.70</td>
<td>0.9988</td>
<td>0</td>
<td>160302</td>
<td>10705</td>
<td>0</td>
</tr>
<tr>
<td>WRC-7</td>
<td>57.431608</td>
<td>-5.693168</td>
<td>609</td>
<td>std</td>
<td>2.70</td>
<td>0.9988</td>
<td>0</td>
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<table>
<thead>
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<th>E=0.17mmka-1</th>
</tr>
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<tr>
<td>WRC-1a</td>
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<td>WRC-1b</td>
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</tr>
<tr>
<td>WRC-6</td>
</tr>
<tr>
<td>WRC-7</td>
</tr>
</tbody>
</table>

Exposure ages were calculated using the CRONUS-Earth (Version 2.0) exposure age calculator at http://hess.ess.washington.edu/ (Balco et al., 2008), which is based on MATLAB software. The aim of CRONUS-Earth is to standardise methods for calculating exposure ages, therefore providing a common basis upon which to compare published exposure ages. The use of universal schemes allows published production rates to be directly compared, and future ‘tweaks’ to the production rates as advances in science are made, will allow the easy re-calculation of existing data (Gosse and Phillips, 2001). The CRONUS exposure age calculator solves equation 2.5, for the exposure age $T$:

Table 2.3 shows the information entered into the CRONUS calculator to calculate exposure ages for the 8 sampling sites. 13 parameters are required to calculate an exposure age in version 2.2.1 of the CRONUS calculator. Exposure ages have been
calculated for 2 erosion rates (since exposure). The first erosion rate is an absolute minimum ($\varepsilon = 0$) and is an extremely unlikely scenario. The presence of rounded quartz clasts protruding a maximum of ~20 mm from the surface of bedrock in the study area indicates a minimum post-glacial erosion rate of ~1.7 mm ka$^{-1}$ (taking 11.5 ka as the final age of LLS deglaciation (Ballantyne, 2010; Stone et al., 1998).
Chapter 3

Results I: Geomorphology and Geochronology
3.1. The post-glacial landscape of Wester Ross

Observations of the post-glacial landscape and glacial geomorphological features of Wester Ross were made during numerous field trips to the region. Focus was placed on the Applecross and Torridon areas, which supported small ice fields during the Younger Dryas. General observations of glacial geomorphology in Torridon and Applecross are presented in sections 3.1.1 and 3.1.2 respectively. The results of a more detailed glacial geomorphological study, focussing on the Coire Mhic Nobuil area of Torridon, are presented in section 3.2.

Wester Ross is dominated by a mountainous landscape, with knife-edge ridges characterising many of the higher peaks (particularly in Torridon). Gently sloping and low-lying terrain is found mainly close to the coast and in valley bottoms. Deep glacial troughs cut through the coastline in an E-W direction forming sea lochs such as Loch Carron and Loch Torridon, and freshwater lochs such as Loch Maree. Corries are numerous throughout Wester Ross, particularly on the north-facing slopes of some of the major mountains. A profusion of exposed, scoured bedrock, roches moutonnées, and striae attest to the impact of successive episodes of glacial erosion on the landscape. Small lochs and lochans are numerous in the gently sloping areas of the valleys and coastal plains, and in places are confined by glacial depositional features. The Tertiary drainage pattern of rivers and streams in Wester Ross has also been defined in places by glacial erosional and depositional features. Many of the glacial valley bottoms are covered by ‘hummocky’ moraine landscape, the origins of which were discussed in chapter 1.

3.1.1. Torridon

The Torridon area (Figure 1.2) can be divided into north and south portions by Glen Torridon. ‘Torridon’ is used in this instance to refer to the land directly to the north of Glen Torridon, where much of this study has been focussed (Figure 3.1).
Figure 3.1. The study area in Torridon, taken from the south of Loch Torridon, facing northwards.

Torridon is extremely popular with hill walkers due to its spectacular scenery and the proximity of several Munros (mountains over 3000 feet/914 m) in one area. Beinn Eighe, Liathach and Beinn Alligin each have two summits which achieve Munro status. The summits of these mountains have been incised on their north facing slopes by a series of corries. The largest of these is Coire Mhic Fhearchair (Figure 3.2), the floor of which covers an area of ~1.5 km². The corrie floor is characterised by an expanse of bedrock, broken by a large lochs and numerous lochans. The oversteepened corrie walls are covered with scree slopes, debris chutes and talus cones. Flutes are abundant in all of the corries of Torridon, running perpendicular to the back walls, and parallel to the direction of palaeo-ice flow.

The valleys of Torridon are separated by knife-edge ridges and radiate from a central (unnamed) region (Figures 1.2 and 3.3), which is covered by an expanse (~15 km²) of ‘hummocky’ terrain. This terrain is broken by a series of lochs and lochans, which fill the hollows between hummocks. Valleys extend in all directions from this central expanse and the most gently sloping valleys (Strath Lungard, and the valleys either side of Baosbheinn) reach to the north and west towards the low-lying terrain of the coastal regions. The gentle inclination of these valleys has led to the formation of two large lochs: Loch na h-Oidhche and Loch a’ Bealaich (Figure 1.2). Extending to the south-west and east (respectively), Coire Mhic Nobuil and Coire Dubh Mor have a steeper incline towards sea level, in Loch Torridon and Glen Torridon. Within Coire Mhic Nobuil in particular, lateral-terminal and retreat moraines are especially numerous and prominent.
Chapter 3: Results I: Geomorphology and Geochronology

Figure 3.2. Coire Mhic Fhearchair, Torridon. Loch Coire Mhic Fhearchair covers most of the corrie floor. The distinctive "triple buttress" feature can be seen at the back wall.

Figure 3.3. Central Torridon as viewed from Coire Mhic Fhearchair, looking north. The vast hummocky terrain is dotted with several large and numerous small lochans.
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3.1.2. The Applecross Peninsula

The Applecross Peninsula (Figure 1.3) is located to the south-west of Torridon on the west coast of Wester Ross. Beinn Bhan is the highest mountain on the peninsula (896 m) and is characterised by a narrow dissected plateau, cut to the north and east by a series of small corries, and to the NW and SE by deeply incised glacial valleys, such as Coire Attadale, Coire nan Cuileag, Coire nan Arr (Figure 3.4) and Coire na Ba. End moraines demarcate the extent of the valley glaciers, with the exception of that marking the terminus of the Coire nan Arr glacier, where an end-moraine is exposed periodically with the ebbing of the tide at the top of Loch Kishorn. Lateral moraines become less distinct up-valley where the slopes steepen and meet the valley walls.
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Figure 3.5. Ice-scoured area above Coire nan Arr, Applecross. Joints in the Torridonian sandstone have been exploited during glacial erosion.

Figure 3.6. Large pressure-release crack above Coire nan Arr, Applecross. These cracks could have formed during deglaciation as pressure from glacial loading in Coire nan Arr decreased, leaving the walls of the corrie unsupported. This area was chosen as the location for one of the cosmogenic $^{10}$Be sampling sites (WRC7). See section 2.4.4.
Within the LLS limits, retreat moraines are less abundant and less distinct than those in the Torridon Mountains. Flutes are abundant throughout the valleys, and individual (< 2 m high) hummocks characterise the upper valleys. A particularly extensive area of chaotic moraine characterises the upper reaches of Coire Attadale, and is also common close to the back wall of many of the north-facing corries. Ice-scoured breaches characterise the watershed at the head of the steep corrie walls, particularly at the Bealach nan Arr breach (between Coire nan Arr and Coire Attadale) and above the NW wall of Coire nan Arr, where the terrain slopes gradually into Coire nan Cuileag. The Bealach nan Arr sits at ~600 m, and is characterised by glacially smoothed bedrock. This breached area contrasts to the heavily weathered steep slopes leading up to Beinn Bhan in the east, and a rocky knoll to the west. The area above the north-west wall of Coire nan Arr is also characterised by glacially scoured bedrock (Figure 3.5), and evidence of a very high-pressure subglacial environment where large pressure release cracks are abundant (Boulton, 1979). Figure 3.6 shows one of the many cracks in an area adjacent to the top of the western wall of Coire nan Arr, which probably formed when the Coire nan Arr glacier retreated, leaving the over-steepened wall unsupported. It is possible that this area was breached by localised ice flow during the LLS. This, and the distinctive terrain in the area warranted further investigation, hence the choice of this site for cosmogenic isotope dating (section 2.4.4).

A general reconnaissance of the Applecross area revealed that previous mapping by Robinson (1977) was accurate. It was therefore deemed unnecessary to re-map the area, and focus was placed on Torridon, where earlier glacial geomorphological reconstructions were considered to be lacking the detail necessary for reconstruction of the LLS ice body and inconclusive with regards to deglacial chronology. The following section outlines the results of this mapping.
3.2. Torridon mapping results

Mapping was focussed on the Coire Mhic Nobuil and Central areas of Torridon (Figure 1.2) and was undertaken mainly on foot, supplemented by aerial photography where required (Figure 2.1). Figure 3.6 is a glacial geomorphological map of the area of Torridon shown in Figure 2.1. Mapping results are outlined in terms of the depositional and erosional features identified in figure 3.6, a larger version of which can be found in Appendix 2. In Figure 3.7, individual retreat stages have been identified and palaeo-ice flow has been mapped using the glacial geomorphological features defined in figure 3.6.

3.3. Erosional features

Features of glacial erosion were recorded in abundance above 400m altitude (Figure 3.6), particularly in the high corries. These included small-scale features (cm) on bedrock, and intermediate to large scale features (>m) caused by the sub-glacial moulding and plucking of the bedrock substrate.

3.3.1. Small-scale features

Small-scale erosional features can be seen throughout Torridon, although they are perhaps not as prolific as they once were, due to sub-aerial Holocene weathering of the Torridonian sandstone substrate. Preserved micro-scale features were only observed within the boundaries of the most recent glaciation.

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*Figure 3.6* (Page 146) Geomorphological map of part of the Torridon study area. The three types of moraine are assigned a unique colour. Type 1 recessional moraine is coloured red, type 2 hummocky moraine is coloured black and type 3 fluted moraine is coloured purple. Ice directional features are also shown. A larger version of this map can also be seen in Appendix 2.

*Figure 3.7.* (Page 147) Geomorphological reconstruction based on Figure 3.6. Black arcs represent the individual retreat stages reconstructed from recessional moraines in Coire Mhic Nobuil. Black arrows represent general ice direction and dashed black arrows represent ice directions inferred from areas of geomorphology where directionality is more ambiguous.
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Figure 3.8. Small-scale glacial erosional features. From left to right: Striae on bedrock in Coire na Caime; Striae and chattermarks on glacially-smoothed bedrock behind Beinn Alligin; Crescentic gouges on bedrock in Toll a’ Mhadaidh Mor (see figures 3.6 and 1.2 for locations). Arrows indicate palaeo-ice flow direction.

There is negligible evidence of similar features beyond these extents due to prolonged exposure. Features on a micro (cm) scale include striae, crescentic gouges and crescentic scars. Striae are scratches incised into bedrock, instigated by the presence of particles embedded in basal glacial ice. They are a direct result of glacial abrasion, and provide an indication of palaeo ice-flow. However, where ambiguities arise in directionality (i.e. ice could have been flowing in either direction), it is not always easy to decipher this. Abrasion is favoured by situations where basal pressure exceeds 10 bars (1 Mpa); Boulton (1979), and where sliding velocity is low. Striae were generally observed on bedrock, which had been protected from sub-aerial weathering by vegetation or gravel/sediment. The striae on figure 3.8a were observed in Coire na Caime (Liathach, Figure 1.2) and have obviously been preserved by a thin cover of sediment. In this instance, striae are indicative of ice flow out of the corrie (i.e. S-N). Striae were mainly observed in the high corries and regions where exposed bedrock is abundant.

The area to the north of Beinn Alligin (Figure 1.2) is characterised by extensive areas of undulating bedrock, covered in striae and crescentic gouges (Figure 3.8 b and c).
Figure 3.9. (Above) Coire na Caime, Liathach, Torridon (Figure 1.2). The corrie floor is covered with bedrock, with some evidence of striae (Figure 3.8). The vast majority of the sediment in this corrie originates from the debris fans and scree slopes, which originate from the over-steepened walls.

Figure 3.10. (Left) Glacially-smoothed bedrock in Coire Mhic Fhearchair, Beinn Eighe, Torridon (Figure 1.2). This glacially-smoothed surface is covered in striae.

Figure 3.11. (Below) Lewisian Gneiss Roche Moutonnée trending E-W (right to left). This feature is found close to Wester Alligin, Torridon (Figure 1.2) and is remnant of a time when Wester Ross was covered by ice flowing from the main ice cap to the east. Beinn Alligin and the shoreline of Loch Torridon can be seen in the background.
Crescentic gouges and scars are created usually by a single clast entrained in basal ice, and result from repeated impact or pressure of the clast on the bedrock surface. Once the elastic limit of the bedrock surface is reached, fracturing occurs in front of the clast, and a wedge of bedrock is broken off, leaving behind a gouge. Crescentic gouges are generally convex down-glacier, such as those observed below Beinn Alligin (Figure 3.8c), and are generally found in series. The close association of these features with striae indicate that they are formed under warm-based ice, with sliding of ice across bedrock surfaces (Glasser and Bennett, 2004)

### 3.3.2. Intermediate to large-scale features

Roches moutonnées, streamlined bedrock and glacial grooves are the predominant intermediate to large (>1 m) scale erosional features in Torridon and Wester Ross. These features can also be observed in the wider Wester Ross region outside the mountainous area of Torridon, and on a variety of rock types. Glacial streamlining has led to the formation of smooth, gently undulating terrain and to elongation in the direction of flow. Streamlining is especially common in the north-facing corries of Torridon, where bedrock is exposed on the corrie floor.

Figure 3.9 shows the extent of glacially-smoothed bedrock in Coire na Caime (Liathach). Loch and lochans fill troughs in the bedrock, which have been excavated by high-pressure, fast moving ice. Streamlining is generally associated with abrasion, and occurs under the pressure range associated with striae formation. Figure 3.10 shows streamlined sandstone bedrock with superimposed striae in Coire Mhic Fhearchair (Beinn Eighe). Where bedrock is streamlined parallel to striae, this is unambiguous evidence for warm-based glacial erosion.

Glacial grooves range from a few metres to hundreds of metres long, several metres wide and ~1m deep (Glasser and Bennett, 2004). They are thought to form due to fast-flowing ice, and their development is determined by the efficiency of glacial abrasion and the production of high-pressure meltwater. They can also form following the exploitation of structural weaknesses in bedrock. Such features can be
observed in Coire Mhic Fhearchair (Beinn Eighe, Torridon) where large glacial grooves have exploited the natural dip of the sandstone bedrock. This will be discussed further in section 6.3, where the results of the Coire Mhic Fhearchair case study will be discussed.

Roches moutonées are asymmetric bedrock bumps, with abraded stoss (up-glacier) faces and quarried lee (down-ice) faces. These features have formed through plucking and regelation mechanisms and are most likely to form where low-pressure (<1 bar) cavities exist at the glacier bed, where ice cover is thin, where high sliding velocities exist and where subglacial water pressure fluctuates due to meltwater supply from the glacier surface (Sugden et al., 1992; Glasser and Bennett, 2004). Figure 3.11 shows a roche moutonnée formed from Lewisian Gniess bedrock near Wester Alligin, outside the extent of the LLS. Within the LLS limits, glacially scoured bedrock and roches moutonées are particularly abundant in the north-facing corries.
3.4. Depositional features

Distinctive contrasts in depositional glacial geomorphology can be observed on Figure 3.6. The main valley of Coire Mhic Nobuil which was mapped on foot is characterised mainly by a densely packed sequence of lateral-terminal retreat moraines which are identified in red on Figure 3.6, and can also be seen in Figure 3.12a. Toll a’ Mhadich Mor (Beinn Alligin), Coire Mhic Fhearchair (Beinn Eighe) and Coire na Caime (Liathach) were also mapped on foot (See Figure 1.2 for locations).

Toll a’ Mhadich Mor is characterised by both retreat moraines and flutes, and is also home to the large rock slide (Ballantyne and Stone, 2004) described in section 1.7.4. The higher north-facing corries, Coire Mhic Fhearchair and Coire na Caime, are characterised mainly by flutings. The vast area of Central Torridon (Figure 3.12b), which was mapped mainly using aerial photographs with ground-truthing in places (Figures 2.1 and 3.6), is characterised mostly by chaotic terrain, with some minor, localised streamlining. An Coire Mor and Coire Beag (Beinn Dearg; see Figure 1.2) were also mapped using aerial photographs due to accessibility issues. Like the north-facing corries mapped on foot, these corries are mainly characterised by superficial fluted terrain. Results are presented below in terms of these three contrasting landforms.

3.4.1. Type 1 feature: Ice-marginal retreat moraine in the lower valleys

Much of Coire Mhic Nobuil is characterised by closely spaced (1-20 m) and occasionally overlapping retreat moraines (Figure 3.12a), mostly asymmetric in cross-profile, with steep (≈35°) ice-proximal slopes and relatively gentle (≈15°) ice-distal slopes. When viewed in plan form, the ridges can be interpreted as individual retreat stages (Figures 3.6 and 3.7).
Small channels generally run parallel to adjacent glacial topography, but most of the larger streams are controlled by large-scale valley form, incising through ridge sequences, giving the landscape a fragmented appearance. Alt Coire Mhic Nobuil (the major river in this valley) is particularly established in the lower valley. Streamlining of moraines by this channel and its tributaries accounts for the ‘V’ shaped ice-fronts in figure 3.7.

Between Loch Torridon and Coire Mhic Fhearchair, there are 84 interpreted retreat stages. The retreat moraines shown in figure 3.6 vary in size with distance from Loch Torridon, the largest being the outermost ridge, the terminus of which (if it exists) is underwater. A prominent lateral moraine on the NW face of Liathach delimits the maximum extent of the LLS in the lower reaches of Coire Mhic Nobuil, and has a larger conjugate, ~10 m high, close to the shore of Loch Torridon.
Figure 3.13. Free faces and end moraines in Coire Mhic Nobuil, Torridon.

Figure 3.14. Sedimentological exposure in Coire Mhic Nobuil. **Dml:** Diamicton, matrix supported with evidence of sub-parallel laminations and sand lenses. Mostly coarse sandy matrix, some areas are clast supported. Clasts are sub-angular-sub-rounded. Gradual coarsening-up sequence to gravels above. **Sd(d):** Fine sandy matrix with deformed bedding. There are concentrated areas of well-sorted granules (GRcu), gravel (Gm) and lenses of fine sand. Large clasts are present within the sandy matrix. The upper boundary is a sharp erosional contact. **Dcm:** Clast-supported, massive diamicton. Contains rip-up lenses of fine sands and a raft of medium-coarse sand which coarsens upwards (Suc). **Dml:** Massive diamicton, matrix supported with evidence of lamination. Areas of heavily weathered, angular class orientated parallel to bed. **Dms:** Massive diamicton, matrix supported with little evidence of layering.
The contrasts in size of these conjugate lateral/terminal moraines have prompted further investigation, following the steps outlined by Benn (1989) and in section 2.1.4. Figure 3.13 shows the extent of the free faces above the reconstructed glacier margin (for more information on reconstruction see chapter 4) in the Coire Mhic Nobuïl Glacier catchment. The areas covered by free faces on both sides of the valley were calculated using equation 2.1. On the south and east-facing slopes of the valley, the map area (Am) was calculated as 1,200 m$^2$ and the average gradient (a) as 0.63, giving an actual slope area (Ar) of 1,485m$^2$. On the north and west-facing slope, the mapped area (Am) covered by free faces was considerably lower at 500m$^2$, although the average gradient (a) was calculated to be much steeper at 1, indicating a total area (Ar) of 925m$^2$. These calculations were made to assess the relationship between moraine asymmetry and the availability of supraglacial material from free faces (Benn, 1989). This is to follow-up field observations and subsequent mapping (Figure 3.6), which reveals a degree of asymmetry between the sizes of the lateral-terminal moraines on opposing sides of the valley. It is obvious in the field and from Figure 3.6 that the moraines on the south and east-facing side of the valley are much larger than those on the north and west facing slopes. The long lateral-terminal moraine on the north-west facing slope below Liathach becomes indistinct close to Loch Torridon, whereas the moraines on the slopes of Loch Torridon on the other side of Coire Mhic Nobuïl are extensive and less distinct in form, reaching heights of up to 12 m and maximum widths of ~50 m.

Subsurface exposures were lacking in Torridon. However, one particularly clear exposure was found at OS Grid Reference 88164/59009 (Figure 2.1) in Coire Mhic Nobuïl. It is part of a recessional moraine on the banks of the stream emerging from Bealach a’ Chomhla (Figure 1.2), just before it meets Alt Coire Mhic Nobuïl, the main channel running the length of Coire Mhic Nobuïl. The exposure has been divided into 5 distinct lithostratigraphical units, which are described in figure 3.14. The lens of fine sand, well-sorted granules and gravel (unit 2), contrasts to the surrounding diamicton units (1, 3, 4 and 5). Clast roundness varies throughout the units, the majority of the clasts being well-rounded to rounded, but a minority are
also angular. This exposure should provide some insight into the formation of the sequence of recessional moraines in Coire Mhic Nobuil during deglaciation.

3.4.2. Type 2 feature: Chaotic terrain in the central region

Recessional moraines and associated ice marginal landforms are sparse above ~400 m in Torridon, with the exception of the retreat stages in Bealach a’ Chomhla (Figure 1.2) and the single retreat stage in An Coire Mor, Beinn Dearg. Above this, hummocky and fluted moraines are abundant. The chaotic moraine in the central region of Torridon is composed of individual mounds (Figure 3.6), some with short linear ridges, some enclosing hollows. The whole area is dotted with lochans and small ponds, which fill the spaces between hummocks. Apart from a few small areas that show some evidence of streamlining, much of the area is disorganised. Some hummocks can also be observed below ~300 m, between lateral terminal moraines, although these are minimal.

3.4.3. Type 3 features: Flutes in the corries

The majority of the fluted moraine can be seen in the larger, high north-facing corries (Coire Mhic Fhearchair (Figure 3.7), Coire na Caime, An Coire Mor), providing an ice directional indicator, perpendicular to flow. Flutes are particularly distinctive in Coire Mhic Fhearchair, where angular clasts of Cambrian Quartzite are superimposed on the underlying bedrock in a series of ‘stripes’ (McCormack et al., 2008). Erosional landforms such as roches moutonees, striae and crescentic gouges are abundant in the corries, particularly in Coire na Caime and Coire Mhic Fhearchair. Glacially-streamlined bedrock is also common throughout the corries and in areas below prominent ridges, such as Am Beacan (Figure 3.6).
3.5. Holocene Reworking

Paraglacial (Holocene) re-working is evident throughout Torridon in the form of talus cones and debris mantled slopes, which are mainly found on the over-steepened slopes and densely jointed walls of the north-facing corries (Figures 3.6, 3.15a and b). The most obvious example of Holocene reworking is the large ~1 km$^2$ rockslide in Toll a’ Mhadaidh Mor, under Beinn Alligin (Figures 3.6 and Figure 3.15c). As discussed in section 1.7.4, it has been classified as a ‘Sturtzstrom’ or an ‘excess run-out avalanche’ by Ballantyne and Stone (2004) and was caused by paraglacial stress release due to the propagation of an internal joint network, which eventually resulted in the collapse of part of Beinn Alligin in the mid-Holocene (3950 ± 320a).

Figure 3.15. (a) A debris chute which has exploited a fault on Beinn Eighe; (b) Scree slopes in Coire na Caime, Liathach; (c) The Beinn Alligin rock avalanche or “Sturtzstrom” below Beinn Alligin. The avalanche is approximately 1km$^2$. 
3.6. Case Study: Coire Mhic Fhearchair, Torridon

As explained in section 2.3.2, Coire Mhic Fhearchair (Beinn Eighe; Figure 1.2) was chosen as a case study site, to assess the suitability and validity of a variety of traditional (field mapping, Figure 3.16; aerial photography, Figure 3.17), contemporary (Digital Elevation Modelling, Figure 3.18) and innovative (ground-based LiDAR, Figures 3.19 and 3.20) of geomorphological mapping methods. In the proceeding sections the results of this survey are presented for all four mapping methods. The results of these four methods have also been compiled into a final map (Figure 3.21). This map provides a detailed overview of the glacial geomorphology of Coire Mhic Fhearchair.

3.6.1. Comparison of mapping methods

Field mapping (Figure 3.16) has revealed some of the finer and most distinctive features including striae, small moraines and the elongated quartzite fluting/debris stripe. Figure 3.17 shows the results of mapping using aerial photography. The results reveal evidence of debris fans, and several further elongated quartzite flutes. The DEM (at a resolution of 5m) shows evidence of undulations in the terrain, namely steps in the bedrock (Figure 3.18). Finally, Figure 3.19 highlights what could be seen using ground-based LiDAR at a resolution of 200-50 mm. Bedrock steps, debris fans and elongate quartzite flutes could all be observed by rotating the images shown in Figure 3.20.

3.6.5. Final glacial geomorphological map of Coire Mhic Fhearchair, Torridon

Figure 3.21 shows the map compiled from the information gathered during each mapping exercise. The most striking features in the corrie are those formed from clasts of Cambrian quartzite. These include the numerous debris fans, which line the over-steepened corrie walls and the distinctive stripes (or flutes). These features extends out of the corrie for ~2 km (S-N), before diverting to the east, and are composed of large boulders and cobbles (>20 cm³ to ~3 m³).
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Figure 3.16. Geomorphological fieldwork map of Coire Mhic Fhearchair. Contours are spaced at 100m. 1 km grid.

Figure 3.17. Geomorphology of Coire Mhic Fhearchair. Mapped using a digital aerial photograph (courtesy of the University of Cambridge Unit for Landscape Modelling). Image covers the same area as figure 3.16.
Figure 3.18. Bedrock geomorphology of Coire Mhic Fhearchair. Mapped using NEXTMap DSMs in ArcMap. (a) Curvature (b) Slope: green represents gentle gradients, red represents steeper gradients (c) Hillshade DSM resolution is 5m, with illumination from 340° and an angle of 30°.

Figure 3.19. Geomorphology mapped using ground-based LiDAR data at 50mm resolution in RiSCAN PRO 1.4.3. (a) Intensity model (note the highest intensity reflections denote the scan stations) (b) False colour model (c) true colour octree with features (transferred from 3D model via geo-referencing).
Figure 3.20. Views of a 50 mm resolution octree created from LiDAR data and digital photography in RiSCAN PRO (Version 1.4.3). The top image is taken looking down the scree slopes on the eastern wall of the corrie, towards the triple buttress. Loch Coire Mhic Fhearchair is the black area in the centre. Other black areas were out of view when the various scans were taken. The bottom image is taken looking towards the back wall of the corrie. Again the triple buttress can be seen at the back of the Corrie and Loch Coire Mhic Fhearchair can be seen in the foreground.
The most prominent stripe is at its widest and densest close to the back wall of Coire Mhic Fhearchair, and thins out towards the lip of the corrie. The small moraines in the central corrie trend dominantly in a S-N direction, with some deflecting slightly to the north-west. One moraine trends in a SW-NE direction. The majority of the striae indicate flow from the back (south-east) wall of the corrie into the central corrie. Some minor deviations from this trend can be observed. The striation close
to the corrie lip was much less distinctive than those at the back of the corrie (personal observation) and trended in a S-N direction. Several bedrock steps were observed at the lip of the corrie and towards the back wall. The two sinuous undulations which run approximately perpendicular to the long fluted features in the north-east part of the corrie differ from the other bedrock steps. Ground-based observation and the digital model produced from the LiDAR data revealed that these undulations are actually in the form of troughs rather than steps. A thick layer of vegetation prevented further investigation, but the possible origins of these troughs will be discussed further in chapter 6.
3.7. Results of cosmogenic $^{10}\text{Be}$ dating

Cosmogenic exposure ages for the eight samples, along with systematic and analytical uncertainties, are shown in Table 3.1. Post-glacial sub-aerial erosion of rock surfaces results in underestimation of exposure ages, due to the reduction with depth in cosmogenic $^{10}\text{Be}$ production rate (Gosse and Phillips, 2001), thus violating a “zero-erosion” assumption. The presence of rounded quartz clasts protruding ~20 mm from the surface of the roche moutonnée (WRC1a) indicates a minimum post-glacial erosion rate of ~1.7 mm ka$^{-1}$ (taking 11.5 ka as the final age of LLS deglaciation (Ballantyne, 2010; Stone et al., 1998). Analytical uncertainties and minimum erosion exposure ages (1.7 mm ka$^{-1}$) are used for comparing our exposure ages, and are plotted on Figure 3.22 in context with the NGRIP chronology. Internal (analytical) uncertainty only takes the measurement errors in the nuclide concentrations into account. External (systematic) uncertainty also accounts for uncertainty in the reference nuclide production rate, due to altitude and both latitudinal and temporal variations in Earth’s geomagnetic field strength (Balco et al., 2008).

Six of the exposure ages (WRC 1a, 1b, 2, 4, 5 and 7) are within ± 1σ (analytical uncertainty), and in the minimum erosion scenario ($\varepsilon = 1.7$ mm ka$^{-1}$), the calculated mean is a YD age of 11.8 ± 1.1 (0.4) ka BP. Samples WRC3 (from above the low-altitude trimline in Torridon) and WRC6 (the knoll in Applecross) are outliers and are significantly older than the mean YD age by ± 1σ (analytical uncertainty) for WRC3 and ± 2σ for WRC6. The results of the calculations in chapter 2 are shown in table 3.1. In this instance, exposure ages have been calculated according to the ‘Lm’ scaling scheme, according to Lal (1991), Stone (2000) and Nishiizumi et al., (1989). The effects of bringing post-glacial erosion into the equation can be seen in the results tables below. The range of differences between the two scenarios for all sample sites using the Lm scaling scheme is 174 - 409 years (the older samples at the higher end of the range). This variation emphasises the importance of considering post-glacial erosion.
### Table 3.1. The results of the CRONUS (version 2.2.1) calculator for the Lm scaling scheme.

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<th>Lab ID</th>
<th>Lat (ºN)</th>
<th>Long (ºE)</th>
<th>Elevation (m)</th>
<th>Shielding and thickness factor*</th>
<th>$^{10}\text{Be}^\prime$ (x10³ atom/g)</th>
<th>$^{10}\text{Be}$ age (ka) $\epsilon=1.7$ mm/ka</th>
<th>$^{10}\text{Be}$ age (ka) $\epsilon=0$ mm/ka</th>
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<td>WRC-1a</td>
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<td>0.9602</td>
<td>100.89 ± 3.51</td>
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<td>11.9 ± 1.1 (0.4)</td>
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<td>0.9602</td>
<td>95.15 ± 3.43</td>
<td>11.4 ± 1.1 (0.4)</td>
<td>11.2 ± 1.0 (0.4)</td>
</tr>
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<td>87.05 ± 3.16</td>
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</table>

- **Figure 3.22.** North Greenland Ice core Project (NGRIP) $\delta^{18}$O curve and associated chronozones (Andersen et al, 2006; Svensson et al, 2006; Hoek et al, 2008; Lowe et al, 2008). Minimum erosion cosmogenic $^{10}\text{Be}$ results are shown in context with climatic chronozones, at their relative heights above sea level and with analytical and systematic errors. Systematic and analytical uncertainties are represented by the error bars.
### Table 3.2. Cosmogenic exposure ages showing all scaling schemes (erosion=1.7 mm ka⁻¹)

The significance and implications of the given exposure ages will be further discussed in chapter 6. The results from the other relevant scaling schemes described in table 2.2 are also displayed in Tables 3.2 and 3.3, showing the results for a minimum erosion situation of 1.7 mm ka⁻¹ and no erosion respectively.
Table 3.3. Cosmogenic exposure ages showing all scaling schemes (erosion=0)

The De scaling scheme (Desilets et al., 2006) has also been applied to the calculation of exposure ages in NW Scotland. The range of differences between the De and Lm scaling schemes (using a 1.7 mm ka⁻¹ scenario) is 1,196 for the oldest (WRC6) and 753 for the youngest (WRC2) exposure ages. This difference highlights the importance of considering nuclide production rates according to variations in latitude, altitude, magnetic field and atmospheric pressure.
3.8. Schmidt hammer results and associated statistical calculations

### Table 3.4: Mean and standard deviations for the Schmidt hammer sample sets taken from each of the cosmogenic dating sites. 50 rebound reading were taken at each site to assess relative weathering.

<table>
<thead>
<tr>
<th>Sample site</th>
<th>Mean rebound value</th>
<th>Standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRC1a</td>
<td>35.84</td>
<td>6.02</td>
</tr>
<tr>
<td>WRC1b</td>
<td>39.84</td>
<td>6.35</td>
</tr>
<tr>
<td>WRC2</td>
<td>41.94</td>
<td>3.30</td>
</tr>
<tr>
<td>WRC3</td>
<td>35.58</td>
<td>4.92</td>
</tr>
<tr>
<td>WRC4</td>
<td>38.76</td>
<td>4.44</td>
</tr>
<tr>
<td>WRC5</td>
<td>41.10</td>
<td>4.63</td>
</tr>
<tr>
<td>WRC6</td>
<td>35.44</td>
<td>4.59</td>
</tr>
<tr>
<td>WRC7</td>
<td>33.76</td>
<td>4.91</td>
</tr>
</tbody>
</table>

### Table 3.5: T-test results for Schmidt hammer sample sets from WRC2 and WRC3 in Torridon.

<table>
<thead>
<tr>
<th>T-test at 95% confidence</th>
<th>T-test at 99% confidence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tcrt</td>
<td>2.01</td>
</tr>
<tr>
<td>Sp</td>
<td>4.74</td>
</tr>
<tr>
<td>Tcalc</td>
<td>6.20</td>
</tr>
</tbody>
</table>

### Table 3.6: T-test results for Schmidt hammer sample sets from WRC4 and WRC6 in Applecross.

<table>
<thead>
<tr>
<th>T-test at 95% confidence</th>
<th>T-test at 99% confidence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tcrt</td>
<td>1.98</td>
</tr>
<tr>
<td>Sp</td>
<td>5.50</td>
</tr>
<tr>
<td>Tcalc</td>
<td>3.11</td>
</tr>
</tbody>
</table>

Weathering contrasts between WRC2 and WRC3 (Torridon), and WRC4 and WRC6 (Applecross) are indicated by Schmidt hammer rebound values. The statistics related to these results are summarised in Tables 3.4 to 3.6. Table 3.4 shows the mean Schmidt Hammer rebound value from a sample set of 50 results and the deviation around this value.
Schmidt Hammer reading datasets for the heavily weathered sites at WRC3 and WRC6, and from adjacent glacially smoothed sites nearby (WRC2 and WRC4 respectively) were statistically scrutinised using t-testing to investigate the differences between the datasets. Tests indicated a significant difference between the two data sets at the 99% confidence level. For WRC2 and WRC3 (Table 3.5), \( T_{calc} \) is greater than \( T_{crit} \) at the 99% confidence limit, therefore the null hypothesis is rejected and the alternative hypothesis is accepted. The same situation is evident for WRC4 and WRC6 (Table 3.6), showing that the weathering contrasts (as indicated by the 50 Schmidt hammer samples) between these sites are significant.
Chapter 4:

Methods II: Glacier Reconstruction
Chapter 4. Methods II: Glacier Reconstruction

4.1. Reconstruction methods

The glacial landscapes introduced in chapters 2 and 3 are shaped by the dynamics of an ice body during the most recent glacial episode. The growth and decay (mass balance) of a glacial mass is heavily dependent on global location, local climate and topographical setting, which all combine to ultimately define internal thermal regime and glacial dynamics (section 1.5.1). These methodologies behind these principles form the basis of the reconstruction of the Torridon and Applecross ice fields and their implications for palaeoclimate at the LLS maximum. The reconstruction of the extent and surface area of the former Loch Lomond Stadial ice fields in Torridon and Applecross is based on geomorphological data, collected in the field and from remotely sensed imagery. Glacier reconstruction enables the calculation of the ELA, facilitating the estimation of palaeo-precipitation values for the ice field.

4.1.1. Reconstruction of ice extents in Torridon and Applecross

The maximum vertical and lateral extent of LLS ice in the study area was established from geomorphological evidence collected on foot, from remotely sensed imagery and through the assessment of previously mapped extents. The geomorphological map of the Coire Mhic Nobuil and central areas of Torridon (Figure 3.6), where the most detailed mapping was undertaken can also be seen enlarged in appendix 2. This section describes the assumptions made in the establishment of ice extents by means of geomorphological mapping.

Boulder and drift-limits were identified in conjunction with meltwater channels and lateral moraines in places. These are particularly obvious in the lower valleys, for example, the large lateral-terminal moraine on the NW face of Liathach in Coire Mhic Nobuil (see appendix 2) was easy to follow for ~2 km, and is bordered by a small channel on its eastern flank. Areas of heavy boulder cover were easily identified from remotely sensed imagery, as were prominent lateral moraines. Drift
limits were defined by a transition from sediment cover to bedrock, although they had to be extrapolated where geomorphological evidence was absent, particularly in the upper valleys and cirques, where paraglacial movement has erased any evidence of former ice limits. Where interpolation of a drift limit was necessary, the upper extent of glaciers was defined as being ~20-30 m below the top of corrie scarps according to Wilson and Clark, (1998); Carr, (2001) and Hughes, (2004).

Weathering limits (trimlines) divide glacially scoured bedrock below form relatively weathered bedrock above (Section 3.5). Two obvious trimlines were observed in the field at ~500m on Am Beacon and below Beinn Alligin (Figure 3.6). In Applecross, weathering limits were less distinct, and were generally interpolated where no discernable drift limit could be seen. It must be considered at this point, that a potential source of error in glacier reconstruction could arise from the use of trimlines as a tool for reconstruction (Section 1.6.2). It is possible that non-erosive, cold-based ice was present above trimlines when the LLS was at its maximum extent and therefore, ice margins depicted in this study represent the minimum ice thickness.

As many of the large valley glaciers in Torridon and Applecross have separate snouts, and flow in a variety of directions, they must logically be considered as separate entities (e.g. Hughes, 2004). The division of these individual glaciers will enable the investigation of the effects of local topographical and climate variations according to the ELA of each glacier. Therefore ice divides were estimated using modern watersheds, ice surface contours, underlying topography, modern drainage pattern, and divergent patterns observed in the local geomorphology (Appendix 2).

4.1.2. Reconstruction of ice surfaces

Former ice field surfaces were reconstructed by drawing ice surface contours at 50 m intervals following procedures introduced in Sissons (1974). Contours were plotted based on the morphology of the reconstructed glacier snout, becoming increasingly less convex upslope, in accordance with modern glacier morphology. The
directionality of geomorphological features including fluted moraines, striae and glacially smoothed surfaces was assessed to ensure that contours were perpendicular to prevalent flow trends. The planimetric area between each contour was then measured to allow calculation of ELAs.
4.2. Equilibrium line altitude (ELA) calculation

As stated in section 4.1.1, the ELA of a glacier is the theoretical altitude on a glacier at which accumulation and ablation are balanced over an astronomical year (Benn and Lehmkuhl, 2000). The relationship between an ELA and the local climate can provide information regarding the glacier's response to climate change and in particular, variations in precipitation and temperature (e.g. Sissons and Sutherland, 1976). It is therefore vital that the ELA is calculated using a sound methodology, incorporating variables such as hypsometry, glacier type and catchment topography. Numerous methods of ELA calculation have been employed in the past, and will be reviewed here.

4.2.1. Approximations based on glacial geomorphology

Methods concerning the approximation of ELA from glacial geomorphological features are thought to be the least accurate. The cirque floor method assumes that a cirque floor occupies a level midway between the ELA during a period of maximum glaciation and the modern ELA (e.g. Foster, 2009). However, cirques usually form over multiple glacial episodes, and in this instance, cannot be relied upon to provide an accurate ELA for the most recent glacial event (Benn and Ballantyne, 2005). The maximum lateral moraine height is thought to provide an indication of former ELA. Lateral moraines develop where ice flows in a radial motion, towards the glacier margins, in the ablation zone. The theory of lateral moraine height assumes that such moraines are preserved in their complete form, that moraine formation took place directly adjacent to the ELA, and that lateral moraine did not form during the recession of said glacier. This technique is also inappropriate for areas where lateral moraines are not present, particularly in high-relief or high-altitude catchments (Benn and Lehmkuhl, 2000).
4.2.2. Area-Weighted Mean Altitude method (e.g. Sissons, 1974)

The Area-Weighted Mean Altitude (AWMA) method takes into account the hypsometry (distribution of surface area with respect to altitude) of a glacier, assumes that accumulation and ablation gradients are identical, and has a linear relationship with altitude (Equation 4.1). The AWMA method provides a good approximation of an ELA, but overestimations are common, based on the fact that ablation gradients are generally steeper than those in the accumulation area (Benn and Evans, 1998).

\[
ELA = \frac{\sum_{i=0}^{n} A_i h_i}{\sum_{i=0}^{n} A_i}
\]

*Equation 4.1*

Where \(A_i\) is the surface area of the glacier between each contour interval \(i\), \(h_i\) is the altitude at the mid-point of each contour interval, and \(n\) is the number of contour intervals.

4.2.3. Accumulation Area Ratio method

The Accumulation Area Ratio (AAR) method assumes that the accumulation area occupies a fixed proportion of the total glacier area, and that the accumulation and ablation areas contribute equally to the net mass balance (Equation 4.2).

\[
AAR = \frac{A_c}{A_c + A_b}
\]

*Equation 4.2*

Where \(A_c\) is accumulation area and \(A_b\) is ablation area. At steady state, modern, mid-high latitude glacier AARs are generally in the region of 0.5 – 0.8, with the majority of AARs between 0.55 and 0.65. The closer the AAR is to 0.6 ± 0.05, the closer the net mass balance is to zero (Porter, 1975). Glaciers with debris-covered...
snouts will generally have lower AARs (~0.4) due to the insulating effect of the debris, the lowering of the ablation rate and the consequent increase in the relative size of the ablation area. This method is most applicable to regular-shaped glaciers without complex morphology, as this method does not take into account the hypsometry of a glacier.

4.2.4. Area Altitude Balance Ratio method

The Area Altitude Balance Ratio (AABR) method attempts to address the issues highlighted by the AWMA and AAR methods. This technique takes into account glacier mass balance and hypsometry (Furbish and Andrews, 1984), by accounting for the variations in accumulation gradient \( b_{nc} \) and ablation gradient \( b_{nb} \). The AABR method assumes that net annual ablation \( d_b \) multiplied by ablation area \( A_b \) is equal to the net annual accumulation \( d_c \) multiplied by the accumulation area \( A_c \) (Equation 4.3).

\[
d_b A_b = d_c A_c
\]

\textit{Equation 4.3}

For equilibrium conditions, the ELA is determined on a trial and error basis using equation 4.4:

\[
b_{nb} / b_{nc} = z_c A_c / z_b A_b
\]

\textit{Equation 4.4}

where \( z_c \) is the area-weighted mean altitude of the accumulation area and \( z_b \) is the area-weighted mean of the ablation area. Osmaston (2005) provided a spreadsheet, which uses equations 4.4 and 4.5 to find elevation of the ELA according to the balance ratio (BR). The BR is defined as \( b_{nb} / b_{nc} \) (equation 4.4), and has been predetermined as ~1.8 for high altitude, continental glaciers with large ablation areas (Furbish and Andrews, 1984) and 2 for mid-latitude, maritime glaciers with small ablation areas (Benn and Gemmell, 1997).
In this instance, ELAs have been calculated using AAR and AABR methods. LLS maximum AARs of 0.5 and 0.6, and AABRs of 1.67, 1.8, and 2.0 were calculated to allow for comparison with previous reconstructions of ELA in Scotland and to provide a range of ELA possibilities to reflect climate at the LLS maximum. AABR calculations were undertaken in Microsoft Excel, following instructions outlined in Osmaston (2005).

The AABR method is considered to be the most reliable (Osmaston, 2005) and in this instance, an AABR ratio of 2.0 is deemed the most reliable for Wester Ross, as it reflects a mid-latitude, maritime climate and glaciers with small ablation areas (Benn and Gemmell, 1997). The possibility that a colder, more continental climate existed during the YD might suggest that an AABR of 1.8 would be appropriate. However, a value of 2.0 is used here to avoid circularity of argument. The results of these reconstructions (chapter 5) did not warrant further investigation, as the difference between the two reconstructions methods is small.
4.3. Palaeoclimate reconstruction

At the ELA, annual accumulation and ablation are equal and therefore positively correlated. As ablation is linked to summer temperature and accumulation to precipitation, there is also a positive correlation between summer temperature and accumulation at the ELA (Benn and Ballantyne, 2005). It is therefore possible to estimate palaeoprecipitation, if palaeotemperature values are available. In this research, we calculate palaeotemperature and precipitation using two different methods. Palaeoclimate calculations based on LLS ELA estimates in NW Scotland (e.g. Ballantyne, 1989; 2002a; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010) have generally been based the first model, summarised by equations 4.5 and 4.6. Equation 4.6 (palaeoprecipitation) is derived from a global, empirical dataset (Ohmura et al., 1992). In contrast, equations 4.7 and 4.8 have been derived from an optimum climate model of NW Scotland at the LLS maximum (Golledge et al., 2008) and account for P/T anomalies integral the Scottish Lateglacial climate (Golledge et al., 2009). Both models will be considered further in the discussion in chapter 6.

4.3.1. Model 1 (Ohmura et al., 1992)

Proxies are used to provide the mean summer temperature, and are generally taken from coleopteran (e.g. Coope et al., 1998) and chironomid (e.g. Brooks and Birks, 2000) assemblages from a site as close as possible to the reconstructed glacier. \( T_3 \) (the estimated average summer temperature derived from modern mean June/July August temperatures) can be reconstructed using mean July temperatures \( T_j \), using equation 4.5.

\[
T_3 = 0.97T_j \quad \text{Equation 4.5.}
\]

Mean July temperature estimates, published in Brooks and Birks (2000), were estimated from Whitrig Bog (362203E/189238N), ~278 km SSE of the study area in Torridon. Temperature can be adjusted for altitude by assuming an adiabatic lapse
rate of 0.006-0.007 °C m⁻¹ (Häckel, 1999). To calculate winter accumulation (i.e. precipitation values) in mm a⁻¹ at the ELA (Pₐ), Ohmura et al., (1992) use a regression curve of T₃ against Pₐ, shown in equation 4.6, which represents the total accumulation needed to balance ablation at the ELA.

\[ P_{a} = a + bT_{3} + cT_{3}^{2} \]  

*Equation 4.6.*

Where a = 645, b = 296 and c = 9. The standard error for this estimate is ± 200 mm a⁻¹. Accounting for this and for the error surrounding the temperature lapse rates, total precipitation is 2294 ± 235 mm a⁻¹ at the mean ELA.

4.3.2. Model 2 (Golledge et al., 2009)

Palaeotemperature estimates (equation 4.8) include the effects of glacially induced cooling, according to the relationship of Khodakov (1975). This relationship alters local temperature according to the length (L) in km of the adjacent ice body. The latitude of the study area and the altitude of the ELA are also incorporated in equation 4.7, where Δx, Δy, and Δz are the easting, northing and altitude differences (km) from the study site used to calculate T_max is equivalent to T_j (equation 4.5)

\[ T = 0.97(T_{\text{max}} + 0.00205\Delta x - 0.00373\Delta y - 6.15\Delta z) - 1.10^{0.28\log L - 0.07} \]  

*Equation 4.7.*

Palaeoprecipitation at the Torridon and Applecross ELAs is calculated using equation 4.8 below. A variety of precipitation and mass balance scenarios are accounted for by a scaling coefficient (S) in equation 4.8, where S is 1 for neutral precipitation seasonality, 1.4 for precipitation distribution centred on the warmest month, and 0.8 for precipitation distribution centred on the coldest month, i.e. annual precipitation dominated by wintertime snowfall. The summer and wintertime end members account for uncertainties regarding unknown temporal variation in precipitation patterns during the YD.

\[ P = S(14.2T^{2} + 248.2T + 213.5) \]  

*Equation 4.8.*
4.4. Modern climate comparison

To enable comparison of palaeo-temperature and palaeo-precipitation with modern values, the complete temperature and precipitation Met Office records (1971-2000) for Kinlochewe, Wester Ross (57.60°N 5.30°W), approximately 10 km ENE of Torridon village has been chosen. This is due to its close proximity to the Torridon mountain range. The site is 25 m above sea level and this has been accounted for in lapse rate calculations for temperature and precipitation. Equation 4.9 accounts for the effect of altitude on precipitation, to allow a direct comparison with YD values (Lukas and Bradwell, 2010).

\[ P_{Z1} = P_{Z2} / \left(1 + P^*\right)^{0.01(Z2-Z1)} \]  

_Equation 4.9._

Where \( P_{Z1} \) and \( P_{Z2} \) are the precipitation values at sea level at Kinlochewe \( (Z2) \) and at the ELA in Torridon or Applecross \( (Z1) \). \( P^* \) is the proportional increase in precipitation per 100m increase in altitude. In the example of Ben Nevis (Ballantyne, 1989) a value of 0.0578 was ascribed to \( P^* \), and is also used here.
4.5. **Calculation of basal shear stress**

Basal shear stress ($\tau_b$) is calculated (in bars) using equation 4.10, where $\rho$ is the density of ice taken as 910 kg/m$^3$, $g$ is the acceleration due to gravity (9.81 m/s$^2$), $H$ is the isostatically adjusted ice thickness based on an ice/Earth mantle density conversion factor of 1.36 and $\alpha$ is the angle of glacier slope.

$$\tau_b = (\rho g H \sin \alpha) / 100,000$$  \hspace{1cm} \text{Equation 4.10}

$H$ and $\alpha$ were calculated using the ice bodies reconstructed following the methods outlined in sections 4.2.1 and 4.2.2. To calculate these parameters, a flow line transect representative of the thickest cross section of each glacier was added to the reconstructions on glaciers deemed of particular interest. Where a glacier was fed by two source areas, two transects were drawn to investigate basal shear stress in these respective source areas. Readings of the glacier surface and ground surface height were taken at 500m intervals along-transect, from the glacier margin to the far reaches of the source areas. On smaller glaciers (specifically those located to the east of Beinn Bhan in Applecross), readings were taken at 250 m intervals to account for their relatively small size. Ice thickness ($H$) was calculated by subtracting the ground height from the glacier surface height (metres above sea level) and was isostatically adjusted by multiplying the result by 1.36. $\alpha$ was calculated using equation 4.11:

$$\frac{h_2 - h_1}{d_2 - d_1}$$  \hspace{1cm} \text{Equation 4.11}

$h_2$ represents the ice surface height at a specific site where $h_1$ is the height of the ice surface at the previous (downslope) site. $d_2$ and $d_1$ represent the distance from the glacier margin for the specific site and the previous (downslope) site respectively.

The results of these readings can be seen in section 5.5 as profiles, showing the glacier hypsometry (glacier surface) the ground surface and the basal shear stress for each site. These are displayed next to plan views of each glacier studied. Six profiles
were produced for valleys in the Torridon area including: Coire Mhic Nobuil and Coire na Caime; Glen Grudie and Coire Mhic Fhearchair; Glen Grudie and Coire Ruadh-staca; Strath Lungard; Flowerdale Forest and Shieldaig Forest. In Applecross, valley profiles included: Coire Attadale; Coire nan Cuileag; Coire na Ba; Coire nan Arr; Coire nan Arr and Coire a’ Chaorachain; Coire Sgamhadail; the Beinn Bhan Corries; Coire Toll a’ Mheine and Coire Gorm Beag.
Chapter 5:

Results II: Glacier Reconstruction
Chapter 5. Results II: Glacier Reconstruction

5.1. Torridon ice extent reconstruction

Figure 5.1. Torridon LLS ice surface reconstruction.

Reconstructions of the Torridon ice field have previously been undertaken by Sissons (1977) and Bennett and Boulton (1993a). The present reconstruction agrees closely with the more extensive reconstruction of the ice field by Bennett and Boulton. However, as their study was focussed on the retreat of the ice field and not the maximum extent, they did not account for features at the centre of the ice field. In this instance, trimlines on nunataks have been added to the reconstruction to provide realistic constraints in the centre of the ice field and the high north-facing
corries on Liathach, Beinn Eighe and Beinn Dearg have also been accounted for in this reconstruction. Two distinct trimlines were observed in the field at ~500m on Am Beacon and below Beinn Alligin (Figure 3.6; Appendix 2) and weathering contrasts at a similar altitude were used as a tool for reconstruction elsewhere. Minor amendments were made to the ice limits in the north, mapped by Ballantyne (1986) and Bennett and Boulton (1993a). In Strath Lungard, ice extents have been altered to include the polished bedrock present an A’ Choich, and the land between 400-500m on the north-facing slopes of Beinn an Eoin. The extent of ice in the Flowerdale Forest valley has also been altered to account for variations in topography close to the terminus (For place names see Figure 1.2).

The Torridon ice field has known links with the ice field to the south (Robinson, 1977) and possible links with ice from the main ice-shed in the east. Geomorphological evidence indicates that ice flowed in a radial pattern from an extensive accumulation area in central Torridon (Figure 3.6; Appendix 2) and therefore flowed independently of ice in the east. However, as it is difficult to define the maximum lateral extent of the ice flowing towards the east and south-east, these glaciers have been excluded from reconstruction and are referred to as areas of uncertainty on Figure 5.1. The exclusion of this ice enabled the accurate reconstruction of ice divides for the five remaining glaciers: Coire Mhic Nobuil Glacier, Shieldaig Forest Glacier, Flowerdale Forest Glacier, Strath Lungard Glacier and Glen Grudie Glacier (Figure 5.1). The ice field was divided to examine the relative effects of topography and local climate conditions on the ELA of the individual glaciers, and to assess the effects of the ice field on their interdependence.
5.2. Applecross ice extent reconstruction

Robinson’s (1977) geomorphological interpretation of this ice field (Figure 7.7) has been corroborated by cosmogenic $^{10}$Be exposure ages from the high breaches in central Applecross (McCormack et al., 2011) and ground-truthing of geomorphological features.
Robinson’s interpretations were found, in most cases, to match closely with field-based observations. Only minor adjustments were made to the lateral extent of the ice field where topography required them. The area of ice flowing over the breached area between Coire nan Cuileag and Coire nan Arr was extended slightly to account for the cosmogenic sampling site at WRC7 and to cover the glacially-scoured bedrock. Cosmogenic $^{10}$Be exposure ages suggest that this area was covered by ice for at least part of the duration of the LLS (Figure 3.22). The Coire nan Cuileag glacier has also been narrowed towards the terminus to due to the topographical constraints of this valley. The Applecross ice field is composed of four main valley glaciers; Coire nan Arr Glacier, Coire Attadale Glacier, Coire nan Cuileag Glacier and Coire na Ba Glacier (Figure 5.2). Beyond the boundaries of the ice field, an independent valley glacier existed to the west in Coire Sgamhaidail, three corrie glaciers resided to the east, behind the Beinn Bhan Plateau and three further small glaciers existed to the north (Figure 5.2).
5.3. Equilibrium Line Altitude (ELA)

Table 5.1 shows the results of ELA calculations using the AAR and AABR methods, as described in section 4.3. The AABR method is considered to be the most reliable (Osmaston, 2005), and as a ratio of 2.0 reflects a mid-latitude, maritime climate and glaciers with small ablation areas (Benn and Gemmell, 1997), ELA 3 is thought to best reflect the YD climate in Wester Ross and is therefore used in the palaeoclimate reconstruction of the Torridon and Applecross ice fields (shown in bold on Table 5.1). The possibility remains that a significant change in the local climate may have enhanced the continentality of this region during the LLS (e.g. Golledge et al., 2009) and an AABR of 1.8 may be more appropriate. However, the range of AABR values in Table 5.1 has a minimal effect on the altitude of the ELA, with the difference between an AABR of 1.8 and 2.0 being ≤ 9 m. When we consider the possible errors encountered due to the uncertainty surrounding the trimline debate and the presence of cold-based ice (section 1.6.2), this difference becomes insignificant. The mean AAR (0.6) is comparable to the mean AABR (1.67) in Torridon, but in Applecross, a difference of 39 m (for the total ice field means) demonstrates the importance of incorporating hypsometry in ELA calculations.

The Torridon reconstruction includes the five main valley glaciers and the Applecross reconstruction includes all glaciers. Mean values are also provided for the four conjoined glaciers of the main ice field in Applecross. Glaciers in Torridon range from 15.1 – 19.4 km², with a total area of 87.1 km². Accounting for the glaciers to the south and east, the ice field is estimated to have covered ~100 km². Using an AABR of 2, ELAs range from 447 – 509 m, and have a mean value of 482 m. Deviation around this mean is 25 m. In Applecross, glaciers included in this reconstruction (see table 5.1), range from 1.2 – 10 km², with a total area of 37.4 km². ELA ranges from 271 – 502 m, with a mean of 361 m and a deviation of 120 m around this value.
Chapter 5. Results II: Glacier Reconstruction

<table>
<thead>
<tr>
<th>Torridon Valley Glaciers</th>
<th>Area (km²)</th>
<th>ELA 1</th>
<th>ELA 2</th>
<th>ELA 3</th>
<th>ELA 4</th>
<th>ELA 5</th>
</tr>
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<td>527</td>
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<td>502</td>
<td>496</td>
<td>570</td>
<td>514</td>
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<td>472</td>
<td>466</td>
<td>502</td>
<td>481</td>
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<td>458</td>
<td>454</td>
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<td>25</td>
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<td>Total Ice field (km²)</td>
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<td></td>
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<table>
<thead>
<tr>
<th>Applecross</th>
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<table>
<thead>
<tr>
<th>Main Valley Glaciers</th>
<th>Area (km²)</th>
<th>ELA 1</th>
<th>ELA 2</th>
<th>ELA 3</th>
<th>ELA 4</th>
<th>ELA 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coire Attadale</td>
<td>12.0</td>
<td>324</td>
<td>320</td>
<td>314</td>
<td>350</td>
<td>324</td>
</tr>
<tr>
<td>Coire nan Arr</td>
<td>8.3</td>
<td>367</td>
<td>360</td>
<td>351</td>
<td>481</td>
<td>399</td>
</tr>
<tr>
<td>Coire nan Cuileag</td>
<td>5.5</td>
<td>448</td>
<td>444</td>
<td>439</td>
<td>454</td>
<td>427</td>
</tr>
<tr>
<td>Coire na Ba</td>
<td>2.9</td>
<td>306</td>
<td>300</td>
<td>291</td>
<td>302</td>
<td>245</td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td>361</td>
<td>356</td>
<td>349</td>
<td>397</td>
<td>349</td>
</tr>
<tr>
<td>Standard deviation</td>
<td></td>
<td>63</td>
<td>64</td>
<td>65</td>
<td>85</td>
<td>82</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Individual Glaciers</th>
</tr>
</thead>
</table>

<table>
<thead>
<tr>
<th>Beinn Bhan Glaciers</th>
<th>ELA 1</th>
<th>ELA 2</th>
<th>ELA 3</th>
<th>ELA 4</th>
<th>ELA 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coire Toll a' Mheine</td>
<td>1.9</td>
<td>448</td>
<td>443</td>
<td>438</td>
<td>397</td>
</tr>
<tr>
<td>Coire Gorm Mor</td>
<td>1.2</td>
<td>509</td>
<td>507</td>
<td>502</td>
<td>516</td>
</tr>
<tr>
<td>Coire Glas</td>
<td>0.2</td>
<td>312</td>
<td>310</td>
<td>307</td>
<td>213</td>
</tr>
<tr>
<td>Coire Muchdaroch</td>
<td>2.6</td>
<td>425</td>
<td>423</td>
<td>420</td>
<td>412</td>
</tr>
<tr>
<td>Meall na Fluid</td>
<td>1.1</td>
<td>307</td>
<td>303</td>
<td>296</td>
<td>280</td>
</tr>
</tbody>
</table>

| Total Ice field mean | 37/2 | 388 | 361 | 363 | 333 |
| Total Ice field st. deviation | 122 | 121 | 120 | 126 | 115 |
| Total Ice field (km²)   | 43.2 |     |     |     |     |

Table 5.1. Area and equilibrium line altitudes (ELAs) of reconstructed glaciers in the Torridon and Applecross ice fields, Wester Ross. The chosen Area-Altitude Balance Ratio (AABR) method of 2.0 (ELA 3) is shown in bold.

The range of values in each ELA estimate is expected, due to the varying contributions of snow blow (mostly from the south-west) and avalanching, depending on the surrounding topography and aspect of glacier surfaces.
5.4. Palaeoclimate reconstruction

Table 5.2. P/T reconstruction based on equations 7.4 and 7.5, derived from a global dataset from Ohmura et al., 1992.

<table>
<thead>
<tr>
<th></th>
<th>$T_3$ (°C)</th>
<th>$P_a$ (mm a⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea level</td>
<td>8.06 ± 0.6</td>
<td>3614 ± 200</td>
</tr>
<tr>
<td>Torridon ELA (482m)</td>
<td>4.9 ± 0.2</td>
<td>2311 ± 266</td>
</tr>
<tr>
<td>Applecross ELA (361m)</td>
<td>5.7 ± 0.2</td>
<td>2656 ± 534</td>
</tr>
</tbody>
</table>

Table 5.3. P/T reconstructions based on equations 7.6 and 7.7, derived from an optimum fit numerical model from Golledge et al., (2008; 2009).

<table>
<thead>
<tr>
<th></th>
<th>$T_3$ (°C)</th>
<th>Precipitation (mm a⁻¹): Seasonality type (S):</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Winter (0.8)</td>
</tr>
<tr>
<td>Torridon ELA (482m)</td>
<td>3.5 ± 0.6</td>
<td>1005 ± 67</td>
</tr>
<tr>
<td>Applecross ELA (361m)</td>
<td>4.2 ± 0.6</td>
<td>1205 ± 233</td>
</tr>
</tbody>
</table>

Table 5.4. Modern P/T Met Office records (1971-2000) for Kinlochewe (57.60°N 5.30°W), approximately 10 km ENE of Torridon village.

<table>
<thead>
<tr>
<th></th>
<th>$T_3$ (°C)</th>
<th>$P_a$ (mm a⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea level</td>
<td>8.4 ± 0.4</td>
<td>2276 ± 163</td>
</tr>
<tr>
<td>Torridon ELA (482m)</td>
<td>1762 ± 230</td>
<td>1886 ± 299</td>
</tr>
</tbody>
</table>

$T_3$ values were firstly calculated according to Benn and Ballantyne, (2005) (equation 4.5), and precipitation values were calculated according to the empirical P/T relationship derived from global values in Ohmura et al., (1992) (equation 4.6). Assuming an adiabatic lapse rate of 0.006-0.007 °C m⁻¹ (Häckel, 1999), $T_3$ at the ELAs in Torridon and Applecross were estimated as 4.9 ± 0.2 and 5.7 ± 0.2 °C respectively (Table 5.2). Palaeoprecipitation has been calculated as 2311 ± 266 and 2625 ± 534 mm a⁻¹ for the Torridon (482 m) and Applecross (361 m) ELAs respectively (Table 5.2).
The results in table 5.3 are derived from a model of the Scottish YD palaeoclimate (Golledge et al., 2009). Temperature and precipitation values for both Torridon and Applecross are lower in all cases (on occasion less than half), than those derived from the Ohmura et al., (1992) global dataset.

In Torridon, a mean summer temperature at the ELA has been calculated as $3.5 \degree C$, and neutral precipitation as $1256 \pm 84 \text{ mm a}^{-1}$ with end members of $1005 \pm 67 \text{ mm a}^{-1}$ and $1758 \pm 118 \text{ mm a}^{-1}$ (for winter and summer dominated precipitation respectively). In Applecross, the mean summer temperature at the mean ELA is estimated to be $4.2 \degree C$, with correspondingly higher precipitation values of $1506 \pm 290 \text{ mm a}^{-1}$ (neutral), $1205 \pm 233 \text{ mm a}^{-1}$ (winter) and $2109 \pm 407 \text{ mm a}^{-1}$ (summer). Modern mean annual precipitation for these ELAs (Pa based on sea level data from Kinlochewe are estimated to be $1762 \pm 230 \text{ mm a}^{-1}$ at the Torridon ELA and $1886 \pm 299 \text{ mm a}^{-1}$ at the Applecross ELA (Table 5.4).
5.5. Basal Shear Stress

The most significant results of basal shear stress calculations (i.e. those provide key information on the controls of glacial dynamics and their relation to climate and topography) are displayed in this section alongside a number of glacier flow line cross sections for Torridon and Applecross. Methods employed in these calculations are outlined in Chapter 4.6, and the results will be discussed in chapter 6.5.

5.5.1. Torridon basal shear stress results

Basal shear stress in the Coire Mhic Nobuil/Coire na Caime glacier longitudinal profile (Figure 5.4) varies between 0.1 and 5.9 bars.
Chapter 5. Results II: Glacier Reconstruction

Figure 5.4. Plan view and cross section of the Coire Mhic Nobuil and Coire na Caime glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.5. Plan view and cross section of the Glen Grudie/Coire Ruadh Staca glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Three obvious peaks can be observed in figure 5.4, starting from the source area in Coire na Caime (8 km from the glacier margin), a minor rise in basal shear stress to 2.3 bars can be seen. The second peak occurs at 5.5 km from the margin, where basal shear stress reaches 4.5 bars. The most significant increase is observed 1 km from the glacier margin where stresses reach 5.9 bars. The glacier varies in thickness, reaching maximum thickness (350 m) between 2.5 and 3 km from the margin. Basal shear stress below the Glen Grudie/Coire Ruadh staca glacier (Figure 5.5) fluctuates frequently between along its longitudinal profile.
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Figure 5.6. Plan view and cross section of the Glen Grudie and Coire Mhic Fhearchair glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.7. Plan view and cross section of the Strath Lungard glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Four peaks in basal shear stress can be observed at 8, 6.5, 4 and 1.5 km from the glacier margin, reaching 6.7, 4.9, 4 and 3 bars respectively. The Glen Grudie glacier reaches its thickest (315 m) at 4.5 km from the glacier margin. The basal shear stress profile of the Coire Mhic Fhearchair stem of the Glen Grudie glacier is more subdued than that of the Coire Ruadh staca stem. According to Figure 5.6, the maximum basal shear stress achieved in the corrie was 2.1 bars.
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Figure 5.8. Plan view and cross section of the Flowerdale Forest glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.9. Plan view and cross section of the Shieldaig Forest glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

The majority of the length of the northward-flowing Strath Lungard (Figure 5.7), Flowerdale forest (Figure 5.8) and Shieldaig forest (Figure 5.9) glaciers are characterised by relatively low basal shear stresses. Basal shear stress below the Strath Lungard glacier (Figure 5.7) ranges between 0 and 2 bars, reaching maximum stresses 2.5 and 5 km from the glacier margin (1.9 and 2 bars respectively). The thickest part (~280 m) of this glacier is located between 5.5 and 6.5 km from the glacier margin. The highest basal shear stress values are observed close to the
Flowerdale forest glacier source area (8-8.5 km from the glacier margin) where a basal shear stress of 2.6 bars is achieved (Figure 5.8), also corresponding to the thickest (~250 m) part of this glacier. The highest basal shear stress (2.4 bars) in the Shieldaig forest glacier can be observed 2 km from the glacier margin (Figure 5.9). The glacier is characterised by a fairly constant thickness between 4 and 8 km, ranging from 275-320 m. From 4 km, ice thickness decreases gradually towards the glacier snout.

5.5.2. Applecross basal shear stress results

Basal shear stress in Coire Attadale is relatively low throughout its longitudinal flowline (Figure 5.11), varying between 0 and 1.8 bars. It is slightly heightened where the glacier is thicker, approximately 7 km from the glacier margin, close to the Bealach nan Arr breach where it reaches 1.8 bars, and 1-2.5 km from the glacier snout, where it reaches 1.6 bars. The Coire Attadale glacier is at its thickest (250 m) approximately 2.5 km from the glacier margin.

According to figure 5.12, the Coire nan Cuileag glacier was particularly thin, reaching a maximum of 95 m between 2.5-3 km from the glacier margin. This is also reflected in the basal shear stress values which are low throughout the longitudinal profile, where basal shear stress varies between 0 and 0.7 bars, achieving maximum stress between 2 and 3 km from the glacier margin.

Basal shear stress under the relatively small Coire na Ba glacier (Figure 5.13) increases with glacier thickness, rising to 3.2 bars, 2.5 km away from the glacier margin where the glacier is 210 m thick. Basal shear stress below the adjacent Coire nan Arr glacier (Figures 5.14 and 5.15) rises slightly to 1.8 bars with distance from the breach at 600 m, where the Coire nan Arr glacier meets the Coire Attadale glacier at altitude. Another noticeable rise in basal shear stress (up to 5.1 bars) occurs in the centre (2.5-3 km for the glacier margin) of the longitudinal profile, where the Coire a’ Chaorchain corrie glacier (Figure 5.15) is confluent with the main stem and glacier thickness reaches approximately 365 m.
The Coire Sgamhadail glacier (Figure 5.16) is characterised by low basal sheer stress throughout much of its longitudinal profile, rising slightly and gradually to a maximum of 1.6 bars towards the glacier snout. The glacier achieves its maximum thickness (~110 m) approximately 1 km from the glacier margin. The Beinn Bhan corrie glaciers (Figure 5.17) are at their thickest approximately 2.25 to 2.5 km from the glacier margin, where the individual stems from the high corries are confluent.
Figure 5.11. Plan view and cross section of the Coire Attadale glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.12. Plan view and cross section of the Coire nan Cuileag glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.13. Plan view and cross section of the Coire na Ba glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.
Figure 5.14. Plan view and cross section of the Coire nan Arr glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.15. Plan view and cross section of the Coire nan Arr and Coire a’ Chaorachain glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.16. Plan view and cross section of the Coire Sgamhadail glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.
Chapter 5. Results II: Glacier Reconstruction

Figure 5.17. Plan view and cross section of the Beinn Bhan corrie glaciers. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.18. Plan view and cross section of the Coire Toll a’ Mheine corrie glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.

Figure 5.19. Plan view and cross section of the Coire Gorm Beag corrie glacier. Basal shear stress is plotted in red alongside the glacier hypsometric profile.
Basal shear stress varies between 0 and 4 bars, peaking approximately 2 km from the glacier margin and again to 2.1, approximately 750 m from the glacier margin.

The Coire Toll a’ Mheine (Figure 5.18) and Coire Gorm Beag (Figure 5.19) glaciers both have thin profiles, with relatively constant and low basal shear stress values, ranging from 0-3.3 bars. The highest (3.3 bars) basal shear stress is observed in the Coire Toll a’ Mheine glacier profile, close to the source area on Beinn Bhan and approximately 1.75 km form the shout.
Chapter 6

Discussion
Chapter 6. Discussion

This discussion chapter addresses the questions initially raised in Chapter 1. Sections 6.1-6.3 focus on the formations and origins of the glacial geomorphological features within the study area. The implications for glacial dynamics and climate will also be addressed and will encompass the wider Wester Ross region (Section 6.1), the mountainous regions of Torridon and Applecross (Section 6.2) and the corries of these regions (Section 6.3), representing the step-wise retreat of ice into the mountainous regions. Section 6.3 also covers the validity of the variety of methods used to map glacial geomorphological features during this research. Section 6.4 addresses the geochronological aspect of this research. The results of cosmogenic isotope dating will be discussed in terms of their implications for the deglaciation of the Wester Ross region and will be placed in context with palaeoclimatic change in the NE Atlantic region. Climatic implications will be investigated further in section 6.5, which addresses the results of the reconstruction of the Torridon and Applecross ice bodies.


Previous studies within the Wester Ross region of NW Scotland have revealed evidence of glacial geomorphological features on scales which reflect the transition from ice sheet to corrie-scale glaciation. This section will briefly highlight the implications of previous studies for glacial re-configuration and the pre-YD climate of Wester Ross. However, much of the focus of this study is placed on the final deglaciation of the LLS ice body present in the mountainous Torridon region during the YD, which will be discussed in the proceeding sections (6.2 and 6.3).

Wester Ross is home to a large variety of glacial geomorphological features, both erosional and depositional, on a wide range of scales and dispersed throughout the mountainous regions (e.g. Torridon and Applecross) and the coastal lowlands. The wide array of post-glacial features throughout the region reflects the change in local
climate over time and a degree of topographic control by the surrounding landscapes.

The presence of offshore features such as Trough-Mouth Fans (TMFs) and prograding wedges at the edge of the continental shelf, and the sequence of submarine moraines observed by Stoker et al., (2006) and Bradwell et al., (2008) all provide evidence of the presence and scale of the BIIS in Wester Ross (Table 1.1). Mega-Scale Glacial Lineations (MSGLs) on the Applecross Peninsula (Bradwell et al., 2005; 2007) indicate that ice was focussed into a fast-flowing stream, a process known to be one of the main regulators of contemporary ice sheets (Clark and Stokes, in Evans, 2003). These features also provide evidence that ice bodies >1,000 km² generally flow regardless of the underlying topography (Figure 1.1).

The climate during the initial stages of deglaciation can be inferred from the glacial geomorphological record. The step-wise retreat of the BIIS (implied by the series of approximately 40 offshore retreat/stillstand moraines; Stoker et al, 2006) suggests that climate did not change in a straightforward manner, and that amelioration was interrupted by intermittent coolings. These coolings could be related to the millennial-scale oscillations outlined in Chapter 1. Occurring approximately every 1,470 years (Rahmstorf, 2003), Dansgaard-Oeschger cycles (and their loosely-associated Heinrich cooling events; Bond and Lotti, 1995) are a likely cause of these readvances. However, given the frequency of these events and the number of retreat moraines (a minimum of 40), the retreat of the BIIS would have taken place over approximately 59,000 years, a much longer time period than the span of the last deglaciation. It is therefore suggested that a number of millennial/sub-millennial North Atlantic climatic cycles were superimposed on one another. When cycles were synchronised and optimum conditions (i.e. the freshening of the Arctic ocean and the consequent slowing/stalling of the Gulf Stream) were met in the North Atlantic, a significant cooling (i.e. Heinrich events) could have resulted, leading to the readvance or stillstand of the retreating ice sheet and the formation of marginal moraines. It has been proposed that most recent Heinrich event (~16 ka; Figure 1.12) could relate to the formation of the WRR moraines (Everest et al., 2006).
However, recently obtained $^{10}$Be isotope dates from the WRR moraine complex (Ballantyne et al., 2009), indicate that these features formed between 14-13 ka.

The WRR moraine complex possibly represents the first onshore position of the BIIS. Until 14 ka (the approximate age of the formation of the WRR moraines according to Ballantyne et al., 2009), sea level was falling rapidly due to glacial unloading (McIntyre and Howe, 2010). Subsequent to this, sea level fell at a steadier state as isostatic uplift became more dominant than the rate of meltwater discharge. As the rate of sea level fall decelerated and isostatic rebound took place, it is possible that the BIIS became topographically pinned to the west coast of Scotland. Ice calving occurs much more frequently in a marine setting (Benn et al., 2007), so it is likely that the BIIS stabilised due to a reduction on ablation. The possibility that the WRR was a stillstand event can be dismissed, due to the presence of multi-directional striae either side of the complex (Figure 1.24; Robinson and Ballantyne 1979) implying that a significant readvance took place. It is possible that following readvance, ice remained at this extent until climate amelioration became the predominant forcing mechanism during the Lateglacial Interstadial.

From observations of the orientation of glacial geomorphological features within Wester Ross, it is apparent that ice gradually became channelled by the surrounding topography during the last deglaciation. Deglacial features such as the WRR moraines represent a time when ice was channelled towards the west by the larger-scale topography of Loch Torridon and Loch Maree (Figure 1.24), overwhelming the mountainous regions of Wester Ross. Topographical controls become increasingly important where isolated mountainous areas (less than ~100 km²) exist (Golledge 2007a). The LLS ice bodies that formed in these regions (ice fields, valley glaciers, small independent glaciers and corrie glaciers) were confined and channelled by the surrounding landscape. In contrast to ice sheets and caps, which generally deposit features irrespective of topography (Golledge, 2007a; 2007b), the glacial features deposited within the mountainous regions were concordant with the local topography.
6.2. Glacial geomorphology in Torridon

This section considers the origins and formation of glacial geomorphological features, focussing on the Coire Mhic Nobuil area and the central region of the Torridon mountains (Figure 3.6). This valley is used here as an analogue for deglaciation in Wester Ross as a whole. The abundance and clarity of the glacial geomorphological features in this valley provide the clearest record of the last deglaciation in the study area. The LLS deglaciation in Wester Ross Torridon will be discussed here in terms of the characteristics of the distinctive glacial geomorphological features (depositional and erosional) and the glacial dynamics associated with each landform. The implications for local climate will be discussed in section 6.4.

The glacial depositional features of Coire Mhic Nobuil and Central Torridon can be divided into three distinct landforms (Figure 3.6). The first of these is retreat moraine, which, as its name suggests, was formed during the retreat and/or readvance of the ice margin in the lower valleys. The origins of this landform are debated and are heavily dependent on internal glacial dynamics, which will be discussed further in section 6.2.1. The second landform is chaotic moraine (section 6.2.2), which is predominant in the central, low-gradient region of Torridon. The origins and formation of this feature are relatively well established but the implications for local climate and topography warrant further discussion. Flutings, which dominate the higher ground and corries (section 6.2.3) provide a contrast to the landscape of the valleys of Torridon (Figure 3.6). The origins of these features are also relatively well established, but the inference for climate and glacial thermal regime in the high corries during the final stages of deglaciation remain unclear.
6.2.1. Depositional landform 1: Ice marginal retreat moraine in the valleys

Retreat (or recessional) moraine (Figure 1.25) dominates the lower valleys of Torridon and is also common in Applecross, defining the extent of individual valley glaciers, and forming a series of conjugate arcs in many valleys, including Coire Mhic Nobuil in Torridon. Retreat moraine is the dominant component of moraine complexes in Scotland (e.g. Bennett and Boulton, 1993a; 1993b; Wilson and Evans, 2000; Lukas, 2005; Benn and Lukas, 2006). In his PhD thesis, Benn (1990) recognised the potential for the use of the transverse linear component of ‘hummocky moraine’ for the reconstruction of retreat sequences. He also highlighted the need for further study of sections through pro-glacially formed moraines, to understand the processes by which sediment is deposited. Several models have been proposed to explain the route of deposition of retreat moraines including an “englacial thrusting” model (e.g. Hambrey et al., 1997) and a model of incremental active retreat (e.g. Lukas and Benn, 2006).

Model I: Englacial thrusting (Figure 3.6): The ‘englacial thrusting’ model (e.g. Hambrey et al., 1997; Bennett et al., 1998) is based on the study of polythermal glaciers in Svalbard. This process is thought to be dominant in polar-region glaciers, where thrusts form due to flow compression at an internal boundary between warm and cold-based ice. This thrust carries sediment to the surface of the glacier close to the snout, where it is stacked in a sequence of pro-glacial ridges. Stresses are propagated into the foreland along a sole thrust (also known as a basal décollement surface) as a glacier advances (Figure 6.1). As the glacier retreats, englacial debris is lowered to form a series of stacked moraines with rectilinear slopes dipping up-glacier. This model has also been suggested for the formation of moraines surrounding LLS glaciers in NW Scotland (Hambrey et al., 2001). This model is based on the morphological characteristics of Scottish retreat moraine (in particular the moraine of Coire a’ Cheud-cnoic, Glen Torridon), and provides little evidence from a sedimentological point of view.
However, this model does not rely on explicit climatic explanations for moraine formation, and implies that multiple and stacked ridges could have formed during a single climatic event. This is worth considering when discussing the retreat sequence in Coire Mhic Nobuil.

**Model II: Incremental active retreat:** This model involves the deposition of sediment at the glacier toe during readvance/stillstand events, thus producing a time-wise pattern of glacial retreat (Figure 6.2). Ice marginal pushing of pro-glacial material is thought to lead to the formation of push moraines during readvances. During stillstands, sediment is dumped in terrestrial pro-glacial fans.
If we consider that the majority of Scottish recessional moraines represent ice-marginal contact fans and/or push moraines, they correspond to successive palaeo-ice fronts (Lukas and Benn, 2006), which can be reconstructed to provide a detailed deglacial history, revealing insights into glacial dynamics and local climate. However, some element of overlap with model I is likely as climate is unlikely to be the only control on where and when sediment is deposited. Sedimentological studies by Lukas (2005; 2007) conclude that much of the retreat moraine in Scotland is formed by dumping of surraglacial sediment at the glacier toe in the form of

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Figure 6.2. Moraine formation according to the active retreat model (Benn, 1990, 1992)
terrestrial ice-contact fans during a winter readvance or stillstand. Evidence of deformation within sedimentary facies implies that a minority of these fans were subsequently deformed by glacio-tectonics/pushing (Figure 6.3).

Moraine sequences such as this are commonly associated with modern temperate glaciers in maritime settings (e.g. Sharp, 1984; Winkler and Matthews, 2009). The oscillatory, seasonal pattern of growth and decay could have been enhanced by the North Atlantic Oscillation (NAO). Taylor et al., (1997) observed a clear correlation between the Gulf Stream and the NAO; that high NAO index values correspond to strong westerlies, favouring a more northerly path of the Gulf Stream a few years later. Nesje et al., (2005) demonstrated that variations in the NAO correlate well with mass balance data from maritime glaciers in south Norway, which are more influenced by winter balance, i.e., an increase in precipitation in the winter. This could account for the oscillatory response of glaciers, and the abundance of retreat moraines in marginal, maritime areas such as Torridon. The dominance of a winter regime could also indicate an increasingly continental climate.

During the reorganisation of North Atlantic circulation at the onset of climate amelioration, it is possible that the oceanic polar front did not migrate as far south as NW Scotland during the winter months, thus leading to a maritime dominated climate moderated by the Gulf Stream and the NAO. This reconfiguration and an associated increase in precipitation could have contributed to the dynamic and prolonged retreat of the Torridon ice field, despite the overall warming trend. The frequency and magnitude of moraine formation during the early stages of LLS deglaciation was regulated by the associated seasonal and sub-annual variations in prevailing climate, perhaps explaining the variation in size of the valley moraines. The influence of local climate and precipitation patterns on glacier dynamics is particularly apparent in Coire Mhic Nobuil where frequent moraine formation can be related to the WSW- ENE trend of the valley, which left the valley open to the effects of the prevailing weather systems from the SW.
Chapter 6. Discussion

Figure 6.3. Formation and deformation of moraines during retreat and readvance of an ice margin. Adapted from Lukas, 2005
Chapter 6. Discussion

**Sedimentary Origins:** The wide range in the angularity of clasts found in moraine sections throughout the NW Scottish Highlands implies that both supraglacial *and* subglacial transport were undertaken prior to the deposition of ice-contact fans (Golledge, 2007a). Studies by Benn and Lukas (2006) suggest that subglacially-transported sediment was elevated to at least the height of the moraine crest prior to deposition. One of the processes behind this elevation is thought to be the thrusting and folding of ice described in model I. This leads to the conclusion that neither model is entirely correct. The poly-genetic nature of moraine formation implies that it varies over space and time, and it is possible that both models are applicable to the processes leading to the formation of LLS retreat moraines in Wester Ross. The scarcity of sedimentological exposures also makes it difficult to judge which (if either) model applies to the formation of the sequence of recessional moraines in Coire Mhic Nobuil. It is however worth discussing the results of the study of the exposure shown in figure 3.14 (section 3.1.4), in the context of the proposed models for moraine formation.

It is difficult to define the stratigraphical history of the exposure shown in Figure 3.14, due to a high degree of ductile deformation throughout the section. However, the presence of fine-grained, well sorted lenses of sediment implies a glacio-lacustrine origin, and suggests that deglaciation in Torridon was characterised by localised, ephemeral pooling. The presence of smaller deformed lenses of sand in unit 1, directly below unit 2; the erosional contact between units 2 and 3; and the existence of a rip-up clast in unit 3, imply that units 1 and 2 were deformed prior to, or during the deposition of the upper units 3, 4 and 5. Observations at this exposure agree with descriptions of ice contact fans in NW Scotland (e.g. Benn and Lukas, 2006; Reinardy and Lukas, 2009). The lower units are likely to be remnant of the first phase of fan development when debris was stacked at a stationary margin, the finer sediments being deposited in an ice-proximal environment, possibly as a subaqueous fan for part of its formation (Evans and Benn, 2004). The evidence of localised pooling implies intermittent climate warming and/or a warm-based regime. This was followed by the retreat and subsequent readvance of the ice margin.
towards or over the original fan, causing glacio-tectonic displacement of the lower units. The addition of the upper units 3, 4 and 5 above the erosion surface, indicates that 1 or more readvances took place. Unit 4 displays evidence of lamination, whereas unit 5 has little evidence of layering. It is therefore possible that unit 5 formed due to slumping of part of unit 4, as a consequence of the removal of ice support.

Glacio-tectonic deformation of the subsurface (Figure 3.14) implies that the Coire Mhic Nobuil Glacier readvanced intermittently during retreat. Glacial readvance was obviously sufficient to deform the sediments within the lower units, but it is unclear whether or not these short-term readvances were substantial enough to override the moraines completely and it is uncertain how often they took place. Sedimentological observations from the region to the south of Glen Torridon by Hodgson, (1986) and also from Skye (Benn, 1992) and Sutherland (Benn and Lukas, 2006), imply a similar, active deglacial regime for these marginal regions. It must also be reiterated that such exposures are rare in Coire Mhic Nobuil and it is therefore difficult to assess whether this particular exposure is representative of other moraines in Torridon. We can however assume that glacial tectonics played a part in deglaciation, however minor.

Combining the topographical control on moraine formation with the effects of regional climate change complicates the retreat pattern in Coire Mhic Nobuil. Evidence of differential retreat of the ice front can be observed in the retreat stages in Coire Mhic Nobuil (Figure 3.7). Where large bifurcated moraines are particularly apparent below the south-west face of Beinn Dearg it is possible that shading of the glacier surface by the imposing north face of Liathach (1055m) may have prevented further retreat until local climate became the dominant controlling factor. Differential retreat could also imply that the part of the glacier toe close to the north-facing slopes of Liathach was insulated by debris cover (Jansson and Fredin, 2002), a plausible scenario, given the oversteepened valley sides and the abundance of debris in Coire Mhic Nobuil.
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Benn and Lukas (2006) noted that one of the most striking features of Scottish moraine (particularly that formed during the LLS) is the large amount of debris it contains compared to moraines in modern environments. Ballantyne (2002b) argues that this resulted from the profusion of unconsolidated paraglacial slope deposits, generated during preceding deglaciations. Paraglacial re-working is evident throughout Torridon in the form of talus cones, alluvial fans and the large rock avalanche under Beinn Alligin. The availability of sediment today implies that it would also have been plentiful during the YD (Ballantyne, 2002b), particularly during the later stages of deglaciation when stress release would have enhanced the development of joints in the surface layer of bedrock.

The area covered by the free faces (steep, exposed rock faces >40°) could have also influenced the amount of debris found in the lateral-terminal moraines on either side of the valley. As outlined in section 2.1.4, Benn (1989) studied within-valley moraine asymmetry in Coire nan Arr, Applecross (Figure 1.3) and other valleys throughout NW Scotland. He observed a correlation between the area covered by free faces and the size of the outermost lateral-terminal moraine, which is also apparent in Coire Mhic Nobuil (Figure 3.13). The most well developed moraines in Coire Mhic Nobuil are found on the same side of the valley as the largest area of free faces. In this instance the most well developed moraines were observed on the south-east facing side of Coire Mhic Nobuil, in contrast to the results of Benn (1989), where larger moraines were observed on the north or east facing side of the valley. Despite the fact that the steepest rock walls were found on the north face of Liathach (Figure 3.13) where the average gradient was calculated as 1, most of the debris seems to have originated from the area below Beinn Alligin, which has a lower average gradient (0.63). Regardless of the lower gradient, the extent of the free faces below Beinn Alligin is the likely cause of the disparity with the results of Benn (1989). The presence of the large sturzstrom in Toll a Mhadaidh Mor (Figure 1.2.8) attests to the availability of sediment and the potential for slope failure on the south-east facing side of Coire Mhic Nobuil. These results indicate the importance of rock jointing and slope failure to sediment supply in Wester Ross.
6.2.2. Depositional landform 2: Chaotic stagnation terrain

The abundance of retreat moraine below 400m contrasts to the abundance of chaotic terrain above (Figure 3.6), which implies that the retreat of ice into the central region also prompted a transition in glacial dynamics during deglaciation. Bennett and Boulton (1993a) suggest that this pattern of active, followed by passive retreat was limited to Skye. However, the contrasting landforms within Torridon suggest otherwise. A vast area of hummocky, disorganised terrain covers much of central Torridon (Figures 3.3 and 3.6). This chaotic landscape is composed of individual mounds, with some circular ridges enclosing hollows. Such vast areas of hummocky moraine are uncommon in most regions of Scotland. However, Benn (1990; 1992) and Benn et al., (1992), describe a similar example of this terrain on Skye as chaotic, with individual mounds and rim-ridges enclosing hollows. The mounds on Skye are composed of subglacial and supraglacial clast, dipping parallel to the slope of the feature close to the hummock surface, but having no obvious structure in the centre of the features. This implies that the mounds were ice-cored during final deglaciation, and suggests a degree of insulation.

Given the relatively large amount of debris deposited in Wester Ross during the LLS, it is possible that stagnating ice was covered by sediment. A thin layer of sediment may speed-up the ablation process due to a localised decrease in albedo, whereas thick, protective cover can prevent ablation due to insulation (Jansson and Fredin, 2002). Figure 6.4 shows the debris-covered toe of the Morteratsch glacier in the Swiss Alps. To the right hand side of the ablation area, ice is preserved by a thick layer of debris. The glacier tongue of the left is actively retreating, whilst the debris-covered area remains stagnant. This debris seems to originate from medial moraine and scree falling onto the surface of the glacier in the ablation area. It is possible that the chaotic terrain observed in central Torridon is the result of stagnant ice preserved under a layer of insulating debris.
Whatever the route of formation, the abundance of chaotic terrain above 400 m altitude indicates that a transition took place during LLS deglaciation. Whether this transition was caused by topographic factors, climatic factors or a mixture of both is difficult to determine. It is possible that the dislocation of accumulation areas in the corries from regions of low gradient could have cut-off ice supply, isolating downwasting ice bodies from their source area (Golledge and Hubbard, 2005). A lack of motion due to a very low longitudinal stress gradient (Willis, 1995), would have cause the remaining ice to become cold based and passive, gradually depositing debris into disorganised heaps. The lack of such an extensive area of hummocky, disorganised moraine in adjacent areas (such as Applecross) implies that the topographic control is significant to the formation of such chaotic landscapes. Small areas of streamlining in this region could indicate a minor, localised increase in basal ice velocity, or could be remnant of a time when ice moved actively across the central region.

In the case of Skye (Benn et al., 1992) the transition in glacial dynamics has been related to changes in local climate, and the initial phase of active retreat is thought to have resulted from a decrease in precipitation. The second stage of retreat can be also identified by the presence of extensive areas of chaotic terrain in Skye, and can
also be linked to final climatic warming. The transitional pattern of decay in Torridon could also be attributed to climate amelioration. It is suggested that the hummocky features in central Torridon formed following a general reduction in precipitation and a rise in temperature. As the local equilibrium line altitude (ELA) raised above the height of the major mountains of Torridon (Bennett and Boulton, 1993a). Ice isolated in the corries would have no-longer supplemented the ice remaining in the central region. Isolated from a source area, it is very likely that this whole area became inactive. A lack of motion due to very low basal sheer stress (Figures 5.7 to 5.9) would have caused the remaining ice to become stagnant, gradually depositing debris into disorganised heaps.

6.2.3. Depositional landform 3: Down-valley radial flutes and ridges

Flutes are prolific in the corries and around the high ground of Torridon and Applecross. Flutes vary in size, ranging from small flutes <50 cm high and <20 m long to much larger-scale flutings, such as the large moraine below Beinn Alligin, which extends down-valley for ~2km and reaches ~100m width in some places (Figure 3.6). Flutes can form during the oscillatory or passive (rapid) retreat of a glacier and are generally orientated parallel to the palaeo-flow of a glacier. They are thought to be formed through a variety of processes, including subglacial moulding by actively retreating glaciers (Hodgson, 1986; Benn, 1992; Wilson and Evans, 2000; Lukas, 2005) and in-situ deposition from an englacial/supraglacial position, during passive decay of cold-based/polythermal glaciers (e.g. Glasser and Hambrey, 2001).

Studies in Polar regions such as Svalbard (e.g. Glasser and Hambrey, 2001) suggest that flutes also form as debris ‘stripes.’ Medial moraines deposited on the surface of polythermal and cold-based glaciers (Figure 6.5) can be traced into the glacier foreland as ‘stripes’ of debris (Figure 6.6), composed of angular cobbles and boulders which are superimposed on the sediment beneath.
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Figure 6.5. Sediment stripes forming from supraglacially-derived material on a Svalbard polythermal glacier. Evidence of longitudinal stratification can be seen on the surface. Picture taken from Glasser and Hambrey, in Evans (2003).

Figure 6.6. Debris stripes can be seen extending onto the foreground of this retreating Svalbard polythermal glacier. Picture taken from Glasser and Hambrey, in Evans (2003).
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Figure 6.7. Fluted ‘stripes’ on the north face of Liathach emerging from Coirreag Dearg. Picture taken looking south. The stripes can be seen emerging from the corrie and diverting to the west in the foreground.

Figure 6.8. A ~2 km long fluted moraine in Coire Mhic Nobuil, Torridon. The red dashed line demarcates the crestline of this feature. Beinn Derag (Figure 1.2) can be seen in the background.

Similar flutings can be observed in many of the high north-facing corries in Torridon, such as Coire Mhic Fhearchair, Coire na Caime, Coire Beag, An coire Mor and Coirreag Dearg (Figure 1.2 and Figure 3.6). Figure 6.7 shows a series of
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stripes emerging from Coirreag Dearg on the north face of Liathach. These features are comparable to those observed emerging from Svalbard polythermal glaciers (Figure 6.6). These features are thought to form due to the stratification of the glacier surface as ice flow converges down-glacier (Glasser and Hambrey, in Evans, (2003)). The origins of the sediment incorporated in these stripes will be discussed in section 6.3.1 in relation to the large example observed in Coire Mhic Fhearchair (Beinn Eighe).

As stagnation was taking place in central Torridon and the ELA rose above the altitude of the Wester Ross mountains, it is likely that the under-nourished corrie glaciers retreated. However, the north-facing aspect of these corries may have prolonged the existence of their glacier counterparts, promoting the formation of features such as the fluted debris stripes observed in the majority of the Torridon corries (Figure 3.6). The diminishing size and thickness of these glaciers may have resulted in a regime change, from a dominantly warm-based regime (indicated by the abundance of erosional features in the corries) to a cold-based or polythermal regime similar to that of the Svalbard glaciers (e.g. Glasser and Hambrey, 2001). This is corroborated by the lack of recessional moraine in the high corries, implying that corrie glaciers did not readvance following final deglaciation. The rapid warming post 11.6 ka, observed in a palaeotemperature proxy from SE Scotland (Brooks and Birks, 2000) implies that climate warming was uni-directional at this point, in accordance with a non-oscillatory final stage of deglaciation.

The superficial form of flutes in the corries contrasts to the substantial flutes observed in the valleys (e.g. Figure 6.8) and implies a contrast in the mode of their formation. Boulton (1976) studied fluted surfaces in deglaciated regions of Svalbard, Iceland, Norway and the Alps, and found that the majority of flutes were formed due to the presence of a rigid structure. Aario (1977) suggests that flutes form due to alternating convergent and divergent subglacial flow. It is likely that the large flute below Beinn Alligin (Figure 6.8) formed as a result of both of the presence of a rigid structure and convergent flow. The presence of this structure provided a source for
material. Located at the junction of two glaciers, one flowing through Bealach a Chomhla and the other originating below Beinn Alligin in Toll a’ Mhadaich Mor, this moraine seems to have been formed where they converged and where flow velocity would have been significantly reduced. A similar moraine can be observed at the junction of Bealach a Chomhla and the main stem of Coire Mhic Nobuil (see Figures 1.2 and 3.6). In contrast to the corrie flutes, these flutes seem to have formed due to internal glacier dynamics, the position of valley junctions and the availability of material at these valley junctions.

6.2.4. Contrasts in erosion and deposition

Within both Torridon and Applecross, erosional features on multiple scales are widely abundant above approximately 400 m altitude and particularly in the high corries. Subglacial erosion occurs when the basal shear stress of a glacier is greater than the bed strength (Benn and Evans, 1998) and can provide information about where glaciers were at their most dynamic and powerful and about the basal thermal conditions. Torridonian Sandstone is generally soft, thickly-bedded and has wide joint spacing. The softness of the rock and the presence of joints in particular lessen the relative strength of the bed (e.g. Dühnforth et al., 2010). The development of widespread subglacial bedforms requires differential movement between the glacier and its bed, and the widespread observation of such features implies a former temperate (warm-based) thermal regime (Boulton, 1974).

The abundance of glacially eroded, exposed bedrock contrasts to the deposition-dominated landscape below 400m. The widespread nature of glacial erosion implies that material from the higher altitudes was transported and deposited in the lower valleys. Without extensive subsurface exposure, it is difficult to say what percentage of sedimentary clasts originated from LLS subglacial erosion, which were supraglacially eroded and which were remnant of earlier glaciation. However, the evidence from $^{10}$Be dating (see section 6.4) implies that at least 2 m of material was eroded from Coire Mhic Nobuil in Torridon and around the Bealach nan Arr
breach in Applecross. This indicates that a significant proportion of glacial till was derived from subglacial erosion and transportation processes during the LLS.

The contrasts between depositional and erosional landforms also have implications for the basal thermal regime of a glacier. Where ice was accumulating in the high-altitude source regions, it would have been thicker and fast flowing and the majority of the erosion would have taken place when the Torridon glaciers were at their most active. Deposition in the valleys could have occurred under warm-based or polythermal regimes via the methods outlined in section 6.2.1. Given the assumption that decay of LLS ice bodies was prompted by the northward migration of the oceanic polar front and the re-introduction of a wetter and warmer climate, it is assumed that the increase in precipitation would have increased mass-turnover. It suggested that the Torridon glaciers became increasingly warm-based and active towards the onset of retreat in the lower valleys, erasing any potential evidence of a cold/polythermal regime. We therefore see a bias towards a warm-based regime in Coire Mhic Nobuil in the glacial geomorphological record.
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6.3. Coire Mhic Fhearchair case study

This section forms a case study based around the glacial geomorphology of Coire Mhic Fhearchair in Central Torridon (Figure 1.2). Firstly, following on from the previous sections, discussion will focus on the origins and formation of the features in this corrie, and secondly, an assessment of the variety of techniques used to map these features will be given.

6.3.1. Glacial geomorphology in Coire Mhic Fhearchair

Coire Mhic Nobuil was chosen as the site for this investigation due to its distinctive glacial geomorphological features and their suitability to test ground-based LiDAR as a potential tool for geomorphological mapping. The corrie floor is mostly obscured by Loch Coire Mhic Fhearchair (shown in the left hand corner of Figure 6.9). Despite this, the abundance of glacially scoured bedrock surrounding its shores implies that the corrie was at one point home to a fast-flowing and highly erosive body of ice. This is also corroborated by the presence of roches moutonnées close to the back wall and at the lip of the corrie, and the results cosmogenic exposure dating, which implies that over 2 m of rock was removed from the roche moutonnée close to the lip during the LLS.

The origins of the most striking features in the corrie, the bright white quartzite stripes (Figure 6.9), are debated. The superficial nature of these features suggests that they were formed during the LLS, and the angularity of the clasts imply that they were supraglacially derived. Tracing the stripes towards the back of the corrie would suggest that their origins lie above ~800 m, where the Cambrian Quartzite overlays the Torridonian Sandstone, also implying that the clasts were not subglacially derived (hence their angularity).
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Figure 6.9. Superficial quartzite fluted ‘stripe’ in Coire Mhic Fhearchair (Figure 1.2), Torridon. The feature overlies bedrock and can be seen stretching towards the lip of the corrie to the north.

It is possible that a fracture in the quartzite was exploited by the LLS glacier that resided in Coire Mhic Fhearchair, resulting in slope failure and the intermittent collapse of fragments onto the glacier surface. The quantity of debris fans shown in figure 3.21 (also composed partially of Cambrian quartzite) which line the walls of Coire Mhic Fhearchair imply that these slopes are highly active. As mentioned in section 6.2.3, Glasser and Hambrey, (2001) studied similar linear features in Svalbard. They were observed to originate from polythermal and cold-based glaciers, implying that towards the end of the LLS, the ice remaining in Coire Mhic Fhearchair (and therefore the other north-facing corries) was thin and slow-moving.

The underlying bedrock also reveals some insight into the glacial history of Coire Mhic Fhearchair. The troughs apparent close to the lip of the corrie (Figure 3.21) underlie and run perpendicular to the quartzite stripes, implying that they formed during a previous glacial episode. The scale of these features (approximately 400 m long and 10 m deep) implies that they formed when ice cover in the Torridon
mountains was much thicker. It is suggested that these features are remnant of the BIIS, during the Dimlington Stadial (~26 - 14.7 ka) when ice extended vertically for up to 900m (or above depending on the presence of cold-based ice). It is difficult to discuss these features further without access to the surface below the thick layer of vegetation that covers this area of the corrie. However, it can be assumed that they are not depositional features, as anything so superficial would have been significantly reworked during the LLS. It can therefore be assumed that they are bedrock groove features, created by fast-flowing ice, under high pressures at the base of the ice sheet. An increase in basal pressure perhaps resulted from the topographic funnelling of ice into this area.

6.3.2. Field mapping

Features (particularly smaller features) were mapped with ease using a GPS (Figure 3.16). Small, streamlined moraines were observed in the central corrie, and a number of striae were noted at a higher altitude close to the back and the lip of the corrie. Large-scale features mapped using a GPS, including the large depositional lineation composed of quartzite boulders. However, towards the corrie lip, this feature becomes unclear where boulder cover becomes sparse and field mapping was longer possible. Field mapping is particularly effective when undertaken in conjunction with remotely sensed imagery. Ground-truthing of features observed in aerial photographs forms the basis upon which to commence field mapping. The ability to map various features in the field is summarised in table 6.2. In theory, all of the observed geomorphological features in Coire Mhic Fhearchair could be mapped in the field, but to map large-scale features would be difficult/dangerous, particularly where features such as debris fans are located on oversteepened corrie slopes. As a stand-alone technique, field mapping can also be a time-consuming and frustrating activity, as the range of view from the ground (particularly in post-glacial terrain, where terrain is undulating) is very limited.
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Table 6.1. Geomorphological features recognised using various mapping techniques.

<table>
<thead>
<tr>
<th></th>
<th>FIELD MAPPING</th>
<th>AERIAL PHOTOGRAPHY</th>
<th>DEM</th>
<th>GROUND-BASED LIDAR</th>
</tr>
</thead>
<tbody>
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<td>Effective maximum spatial resolution:</td>
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<td>~2 m</td>
<td>5 m</td>
<td>1 mm</td>
</tr>
<tr>
<td>Bedrock steps/troughs:</td>
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<td>Y*</td>
<td>Y</td>
<td>Y</td>
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<tr>
<td>Quartzite flutes:</td>
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<td>Y</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>Debris fans:</td>
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<td>N</td>
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<td>Small moraine:</td>
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<td>Y*</td>
</tr>
<tr>
<td>Striae:</td>
<td>Y</td>
<td>N</td>
<td>N</td>
<td>Y*</td>
</tr>
</tbody>
</table>

Table 6.1. Geomorphological features recognised using various mapping techniques.

Y: Feature can easily be observed
N: Feature cannot be observed
Y*: Feature cannot always be seen and/or the technique used does not provide reliable information in this instance.

In this instance, aerial observations of the small moraines in the central corrie did not match up to observations in the field, due to low tonal variation. Debris fans on over-steepened slopes were difficult to map in the field but were easy to define from the aerial photograph, where not obscured by shadow. Features including the quartzite depositional lineation could be mapped with greater ease than on the ground, and numerous other linear depositional features could be observed (mainly due to the contrast between the bright-white quartzite boulders and the underlying vegetation/sandstone). As discussed in section 6.3.2, the concurrent use of aerial photographs and field mapping techniques forms is essential for the accurate mapping and understanding of post-glacial landscapes. Table 6.2 summarises the results of glacial geomorphological mapping from aerial photographs. Features which could not be observed included striae, which can only be seen using very high-resolution imagery or during field-based study.

6.3.4. Digital Elevation Models

During post-processing, the benefits of using analysis tools on a DSM, were evaluated. The DSM in Figure 3.18c was post-processed to achieve a maximum resolution (5 m) and was shaded using an azimuth of 340° to illuminate as much of the corrie as possible, and minimise shadowing around the corrie walls. An
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elevation angle of 30° was chosen following the recommendation of Smith and Clark (2005), to highlight any low-lying features. Many of the features identified in aerial photographs, and during fieldwork, were too subtle in to be seen in the DSM, due to a lack of detail at a lower resolution, and limited tonal variation. However, the slope tool (Figure 3.18b) proved to be extremely useful in highlighting bedrock steps throughout the corrie (Table 6.2), which were not readily observed on foot due to a limited range of view. The curvature tool (Figure 3.18a) did not provide any further geomorphological information.

6.3.5. Ground-based LiDAR

The positive and negative aspects of LiDAR as a glacial-geomorphological mapping tool will be assessed in this section, in terms of the study in Coire Mhic Fhearchair and in terms of its wider applications. In principle, all of the observed glacial geomorphological features in Coire Mhic Fhearchair could be seen using ground-based LiDAR (Figure 3.19c). Undulations in bedrock and topography could be observed and measured by manoeuvring the model on screen in RiSCAN PRO. Fluted topography could easily be mapped due to the contrasts in tone introduced by the integrated digital photographs, as could the debris fans on the steep corrie walls. Striae and small moraines could not be observed in this study due to a lack of time and battery power (Table 6.2).

The sheer bulk and quantity equipment needed to operate a LiDAR for several days in a remote field location requires significant man-power, and clear, dry weather. The intensity and phase of a returning beam can be affected by surface water on vegetation and bedrock, precipitation in the air, and misty conditions. The availability of sufficient time and financial resources are essential for a successful scanning expedition. In this instance, time constraints (due to limited battery power) and poor weather conditions (mist and rain, interrupted by spells of bright sunshine) restricted the quality and quantity of data and imagery that could be acquired.
Financial constraints prevented a repeat expedition during the time-span of this research. However, given the time and the battery power, further ground-based LiDAR investigations within Coire Mhic Fhearchair could have provided more focussed, high-resolution scans of these smaller features.

The battery-powered laptop facilitated the observation of data in the field, so that adjustments could be made at the time of data collection, to minimise problems with shadowing (time constraints limited this on the expedition in July 2007). Scanning resolution could be adjusted to enable the observation and interpretation of features of particular interest, and the ability to scan at up to 1 mm presents the potential to identify small-scale features such as striae in the digital model. The Riegl LMS-Z420i model was able to scan 12,000 xyz points per second within an 800 m range of view. Since the July 2007 trip to collect scan data, LiDAR scanning capabilities have developed significantly, and the potential for the future of mapping (including geomorphological mapping) is high. The integrated system collects data points, gps data and georeferenced, high resolution colour imagery with minimal effort once the relevant location is reached. This information can be used an infinite number of times to study different aspects of a location. Return visits also can be carried out to assess the effects of movement at a study site, providing vital information on the rate and style of Earth surface processes.
6.4. Geochronology

In this section, the deglaciation of Wester Ross during the YD will be considered in terms of the timing of deglaciation and the vertical extent of ice cover in Torridon and Applecross respectively. Mean exposure ages have been used to reconstruct approximate LLS palaeo-ice surfaces for the final stages of deglaciation in the central areas of these marginal ice fields (Figure 6.7). The results for both areas will also be interpreted in terms of marginal ice field dynamics, and the relationship to climate fluctuations, as inferred from proxies and glaciological models. The outlying pre-YD exposure ages from WRC3 and WRC6 will be discussed briefly, in terms of their significance for deglaciation prior to the YD, but given that these dates are stand-alone examples and may have been subject to complex exposure histories, focus will be placed on the six YD exposure ages.

6.4.2. The timing of the LLS in Torridon

The exposure ages for WRC1a and b are statistically indistinguishable from each other (Table 3.1), implying that the quarrying of the bedrock step on the roche moutonnée was part of the same glacial episode that reset both surfaces, in this case, the LLS. The apparent lack of an inheritance signal suggests that a large amount (>2 m) of material was removed from both surfaces by LLS glaciers, by sub-glacial erosion. Extensive bedrock scouring, roches moutoneés and abundant striae within Coire Mhic Fhearchair and throughout the study areas suggest that meltwater-lubricated basal sliding was widespread during the LLS. The average date of exposure (11.8 ± 0.4 ka BP) for the roche moutonnée feature indicates that the corrie lip was first exposed towards the end of the YD. By combining the exposure age for the roche moutonnée with WRC2, a mean age of 11.7 ± 0.4 ka BP (assuming that temperate ice covered all three sites), implies that significant amount of ice still existed in Torridon late in the YD. This ice surface has been reconstructed in cross-section on Figure 6.8 using cosmogenic exposure ages from these three sites.
sites, and provides a plausible estimate of the scenario prior to final deglaciation. This scenario indicates that ice retreat in the corries was simultaneous with the retreat of ice in the centre of Torridon. Ice may have lingered within Coire Mhic Fhearchair for some time following bedrock exposure at the corrie lip, due to its north-facing aspect, but the mean exposure age for WRC1a and WRC1b (11.7 ± 0.4 ka) implies that following this date, ice did not readvance out of the corrie, supporting the geomorphological observations discussed in section 6.2.3.

The exposure age at WRC3 (13.4 ± 0.5 ka) is 1σ different from the exposure age at WRC2, thus implying that the low altitude trimline on Am Beacan (Fig 7a) is a LLS ice limit. The internal uncertainty (0.5 ka) of this exposure age would suggest that this site was exposed within the Lateglacial interstadial, and would make it directly comparable to exposure ages (14-13 ka) from the WRR moraines (Bradwell et al., 2008; Ballantyne et al., 2009; Ballantyne, 2009). However, systematic uncertainty (± 1.3 ka) around this stand-alone date, suggests that the site could also have been exposed during the YD. Given the uncertainty surrounding the origins of trimlines (e.g. Ballantyne, 2010) and the possibility that the site was covered by cold-based, non-erosive ice, it is likely that WRC3 has been subject to a complex exposure history, and must therefore be considered with extreme caution.

6.4.2. The timing of the LLS in Applecross

Applecross is one of the few locations in NW Scotland to have been mechanically isolated from the main LLR ice sheet (Robinson, 1977; Jones, 1998) and wholly self-regulating. Geomorphological mapping by Robinson (1977) suggested that LLR ice breached the high ground between Coire nan Arr and Coire Attadale (Fig 4). This is now corroborated by exposure dating. Dates from WRC 4, 5 and 7 are all statistically similar within 1σ, placing the mean age of bedrock exposure at 11.9 ± 0.5 ka (Fig 7b). These exposure ages are further supported by late-YD cosmogenic ^10^Be dates collected from boulders of Torridonian sandstone in Coire nan Arr (J. Stone, Pers. Comm. 2009). Jones (1998) used a combination of data collected
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during mapping excursions and high-resolution glaciological modelling (Hubbard, 1997), and suggested that the breach and high corrie walls (sampled in this study) were preserved by snow or ice patches during the YD. However, the YD exposure ages for the breaches imply that >2m of erosion took place at this time, thus resetting the cosmogenic clock and removing inheritance. The covering of these surfaces by snow and ice patches can therefore be discounted and Robinson's reconstruction of more extensive ice cover during the LLS (Figure 1.29) is accepted. Combining Jones' mapping/modelling approach with exposure dating demonstrates the added benefits of chronological constraint in glacial reconstruction.

The $2\sigma$ difference between exposure ages for WRC6 and WRC4 in Applecross, and the weathering contrast (supported by Schmidt Hammer readings) between these two sites, implies that the knoll (WRC6) on Applecross is a pre-YD surface. It is possible that this knoll and Beinn Bhan (Figure 6.10) were both exposed as nunataks during the late Dimlington Stadial, but the potential for non-erosive, cold-based ice cover on the knoll during the YD remains. This would result in inheritance of the $^{10}\text{Be}$ signal and could imply that the exposure age for WRC6 ($17.5 \pm 1.2$ ka) is too young and the knoll was actually exposed earlier in the Dimlington Stadial. However, the exposed situation of the Applecross plateau would presumably prevent a significant amount of snow (and therefore ice) from accumulating here, making it unlikely that the site at WRC6 was re-covered by ice during the YD.

6.4.3. A wider perspective

Cosmogenic exposure ages will be considered in terms of cryosphere-climate interactions to gain further insight into the mode of deglaciation in Wester Ross. Temperature proxies ranging from North-Atlantic (NGRIP $\delta^{18}\text{O}$ record) to regional (Scottish) perspectives provide an insight into climate dynamics. Climate-driven glaciological models of Scottish LLS ice bodies and the implications of the marginal, maritime location of the study area will be considered alongside personal observations of local geomorphology from chapter 3.
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It is not clear if deglaciation was complete throughout Scotland, before the onset of the YD. It is debatable whether Lateglacial interstadial ice reached a stillstand just inside the maximum LLS limits, shrank well-within LLS limits and was subsequently rejuvenated or disappeared completely and then re-grew (Ballantyne, 1986; Ballantyne, 2010; Golledge, 2010). It is possible that cooling episodes during the Older Dryas (14.0 - 13.9 ka) and the Intra-Allerød cold phases (13.2 - 12.9 ka) would have promoted the persistence of ice in Wester Ross between 14-13 ka (Ballantyne et al., 2009; Bradwell et al., 2008). Given the uncertainties surrounding the WRC3 exposure age, nothing can be concluded with regard to the situation during the Lateglacial Interstadial. The collection of further samples from the site around WRC3 on Am Beacan could provide an insight into the longevity of ice in source areas during the Lateglacial interstadial, and the extent of its survival into the YD. However, it is clear that during this transitory period, a reconfiguration would have taken place.

Field observations from Coire Mhic Fhearchair, and the results of a LiDAR scan (McCormack et al., 2008) indicate a palimpsest terrain, which is characteristic of much of the NW Scottish Highlands (Wilson and Evans, 2000; Golledge, 2007a) and indicates ice reorganisation from an ice cap dominated landscape to one of a marginal ice field. As discussed in section 6.1, Golledge's (2007b) classification of an ice field (symmetrical valley-side moraine deposition, parallel fluted moraines and heavily eroded valley floors) agrees with field observations of the LLS from Torridon and Applecross. Marginal ice fields such as this responded more sensitively to external changes than the main Scottish LLS ice cap.

In a high-resolution simulation of the LLS in Scotland (Golledge et al., 2008), empirical glacier extents of the main Scottish ice cap (as defined by the BRITICE project by Clark et al., 2004) match the optimum model well. However, Wester Ross is one of the marginal LLS ice field locations, which does not agree with this best-fit scenario.
According to the optimum simulation, the ice remaining in central Torridon and Applecross disappeared at ~12.3 ka. The average exposure age of the six YD sites in central Torridon and from the Applecross breach (11.8 ± 1.1 ka), imply that ice survived could have survived until closer to the onset of the Holocene.

The decay of glaciers on a scale of ~300 years following the LLS maximum, contrasts to the rapid (~40-50 year) transition observed in the Greenland ice core record (e.g. Taylor et al., 1997). A recent optimum simulation model (Golledge et al., 2008) in which the main ice cap responded in-keeping with empirical measurements, also implies that ice disappeared in Torridon rapidly, and well
before the demise of the main ice cap, possibly because it was a small, marginal ice field. However, exposure dating conflicts with the results of this modelling experiment, and implies that a significant amount of ice still existed in central Torridon in the final stages of the YD. As a small ice field during the LLS, it is unlikely that Torridon responded to climate amelioration in the same way as the main ice cap. It is understood that glaciers in maritime locations can survive climatic amelioration as long as winter snowfall is high enough to last through a summer season (Glasser, 1997; Bradwell et al., 2008). A temporally-extensive deglaciation is supported by field observations of conjugate readvance moraines in Coire Mhic Nobuil (Figure 3.6), concurring with Bennett and Boulton’s (1993a) interpretation of LLS an active mode of decay.

The sluggish response time of the LLS glaciers shown by their survival beyond the time shown in glaciological models (e.g. Golledge et al., 2008), in contrast to the rapid transition towards the Holocene observed in the NGRIP curve (Figure 6.2) can mainly be explained by the dynamic response of glaciers to climate change. However, the presence of localised mountain belts may have also influenced the complexity of climatic response during deglaciation (Coope et al., 1998). LLS deglaciation is thought to have taken place from E-W in more northerly locations (Bennett and Boulton, 1993a), likely relating to precipitation anomalies, with high accumulation rates favouring the western regions, where maritime air masses had a high moisture content (Golledge et al., 2008). Sissons (1977) noticed the inland rise in former firn-line altitudes in Scotland, implying that snowfall diminished away from the coast. He states that: “The most massive accumulation was situated between Lochs Torridon and Maree, where some of the highest ground lies close to the west coast.” The preservation of snow and ice is highly dependent on the aspect of a mountain. A high proportion of mountain tops and plateaux aligned E-W in Torridon and Applecross favour the accumulation of snow (through direct precipitation, wind-blow and avalanching) in north-facing corries and on north-facing slopes. It is therefore likely that snow and ice were able to survive in Wester Ross for some time following initial climate warming, explaining the anomalies in
optimum models of LLS deglaciation. The non-dynamic nature of the corrie glaciers prior to the onset of the Holocene implies that climate amelioration eventually became the major driving force behind the final deglaciation of Wester Ross. The glacial response to local climate and topography at the LLS maximum in Wester Ross will be the focus of the proceeding section.
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6.5 Glacial Reconstruction

This section focuses on the reconstruction of palaeoclimatic conditions at the LLS maximum. Firstly, the ELA values (section 6.5.1) for individual glaciers in Torridon and Applecross are analysed in terms of their implication for climate conditions during the YD. Focus is then placed on the basal conditions of these glaciers and the implications for thermal regime and resulting glacial geomorphology. Finally the values of palaeoclimatic precipitation and temperature (P/T) indicators, derived from a global dataset (Ohmura et al., 1992) and a dataset that takes into account climatic anomalies specific to NW Scotland (Golledge et al., 2009), including enhanced seasonality and localised glacier cooling.

6.5.1. Internal variation in ELA within the Torridon and Applecross ice fields

This section focuses on the results of ELA calculations (section 5.3), concentrating on the variation of ELA values within the Torridon and Applecross ice fields. The choice of the AABR of 2 was based on assumptions made about the climate in Wester Ross during the LLS, although it is also possible that an AABR of 1.8 may have also been representative of a stage of the YD when the presence of sea ice led to an increasingly continental climate in Wester Ross. However, the difference in ELA values for Torridon (488-525 m) and Applecross (333-372 m) for all five reconstruction methods is small (Table 5.1), and therefore the effect on palaeoclimate reconstruction is considered to be minimal.

Given that variations in P/T conditions within the Torridon and Applecross ice fields are likely to have been minor, local topographical factors were apparently influential in their formation. On a local scale, aspect and mountain shape are important to the formation and maintenance of such glaciers. High-altitude mountain tops will favour accumulation of snow, but narrow, steep summits will not be able to accommodate the formation of a glacier (Manley, 1955), thus explaining the existence of nunataks.
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The variation around the mean ELA in Torridon and Applecross can be related to the local topographical effect on snow accumulation, aspect, shading of the glacier surface. The dominance of the south-westerly snow bearing winds (Sissons, 1979a; 1980) is of particular importance to the growth of individual glaciers within marginal ice fields. The eastwards and northwards transference of snow and the corresponding decrease in ELA is characteristic of many of the ice fields close to the coast of NW Scotland (e.g. Ballantyne, 1989; 2002a; 2006; and Lukas and Bradwell, 2010).

In Torridon, the dominance of the south-westerly winds is apparent from a number of observations. The highest ELA (509 m) belongs to the Coire Mhic Nobuil glacier in the south, and all of the glaciers to the north are lower, concurring with the northwards transference of snow (e.g. Ballantyne, 1989). The asymmetry of the contouring on the Flowerdale and Strath Lungard glaciers (and to a certain extent the Glen Grudie Glacier) implies that snow accumulated on the eastern flanks of N-S trending ridges, reflecting the eastwards transference of snow. ELA decreases with distance to the east (with the exception of the Glen Grudie glacier which has an ELA of 490 m), declining from 496m on the Shieldaig glacier, to 466m on the Flowerdale Forest glacier and 447m on the Strath Lungard glacier (Figure 7.9). This contrast could represent the transference of snow across ice divides, favouring the accumulation of snow on the northern and eastern lobes of the ice field. This phenomenon has also been observed in the Drummochter Hills (Benn and Ballantyne, 2005); on Skye (Ballantyne, 1989) and on Mull (Ballantyne, 2002a). The increase in ELA on the Glen Grudie Glacier could be a function of hypsometry. Much of the accumulation on this glacier was focussed in the high corries of Beinn Eighe (Figure 5.1), elevating the average altitude of the glacier surface (and therefore the ELA). The relatively high ELA of the Coire Mhic Nobuil Glacier is likely to reflect the greater insolation received by this SW-facing glacier compared to its north-facing counterparts. However, the size of this glacier would indicate that insolation was not a major factor influencing ablation during the LLS (Sutherland,
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1984). This is also supported by the presence of three south-flowing glaciers on the Applecross peninsula (Figure 5.2).

Snow blow source areas, defined as the area upslope and upwind of the glacier surface (in this case to the southwest, between 180-270°), include all glacier-facing slopes, and all plateau-edge slopes <5°, irrespective of orientation (Benn and Ballantyne, 2005). The vast lowland area (covering ~ 100 km²) to the west of the Applecross glaciers would have provided a plentiful source of snow for the ice field to the east, and local ELA variations are likely to result from the presence of N-S ridges within the ice field. The Coire nan Cuileag Glacier had a particularly high ELA, which can be attributed to the fact that it was located in a fairly shallow valley, from which snow would have been easily relocated. Given the relatively low ELA of the Coire Attadale glacier, it is possible that snow from Coire nan Cuileag was blown over, and accumulated behind the N-S trending ridge separating the two valleys (Figure 7.5). Like Coire Attadale, the Coire nan Arr and Coire na Ba valleys to the south are deep and steep-sided, and were obviously suitable for the accumulation and preservation of snow and ice during the LLS.

The east-facing corrie glaciers probably resulted from the same phenomenon, and accumulated behind the high plateau of Beinn Bhan as snow was blown towards the east. The ELA of these corrie glaciers becomes much higher towards the north, with the Beinn Bhan glaciers ELA at 347 m, the Coire Toll a’ Mheine glacier at 438 m and the Coire Gorm Mor glacier at 502 m. This implies that the most southerly glaciers were shaded by the high Beinn Bhan plateau, allowing for the increased accumulation of snow. Further north, it is likely that the smaller corrie glaciers were more open to snow blow from the south west, but were also susceptible to the removal of snow by the same winds. The most compelling evidence for the north-eastward transference of snow by strong SW winds lies in the formation of the small independent ice bodies identified by Robinson (1977), and found in low-lying south-facing hollows. For example, the south-facing Coire Sgamhadail Glacier formed below 400 m altitude and had an ELA of 271, 90 m lower than the mean ELA in
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Applecross. A similar phenomenon has been observed independent glaciers around the Mull ice field (Ballantyne, 2002a). Such low ELAs cannot be explained without the north-eastward transference of snow.

6.5.2. Variation in basal shear stress within the Torridon and Applecross ice fields

The results of basal shear stress and glacier hypsometry (section 5.5) are discussed in this section in relation to topographical control, thermal regime and the resulting glacial geomorphology in the Torridon and Applecross regions of Wester Ross. The accuracy of these reconstructions is dependent on the accuracy of the original reconstruction of the ice surface. Ice surface reconstructions and therefore basal shear stress results are believed to be more accurate in the terminal zone where lateral-terminal moraines have been used as a basis for reconstruction. A lack of control points (i.e. trimlines) in the source areas (particularly in Applecross) means that the area above the ELA may provide a less accurate picture of basal conditions. However they can also be used to corroborate the reconstructions, and provide a profile view of the glacier based on surface contouring.

The contrasts in basal shear stress observed in section 5.5 can be attributed to a number of topographical and glacial characteristics (i.e., the gradient of the bed and the confluence/divergence of glaciers). Glacial dynamics and thermal regime (i.e., warm-based, fast-flowing and erosive or cold-based and passive) seem to be a product of the surrounding topography. These factors combine to define the locations of the erosional and depositional landscapes in Torridon and Applecross.

As a general observation, fluctuations in basal shear stress can be linked to the fluctuations in the ground surface gradient in the valley profiles. Fluctuating shear stresses in Coire Mhic Nobuil and Glen Grudie in Torridon and Coire nan Arr in Applecross can be partially attributed to the variation in the gradient of the ground surface due to the presence of corrie source areas. In contrast, the majority of the length of the northward-flowing Strath Lungard, Flowerdale forest and Shieldaig forest glaciers are characterised by relatively low basal shear stresses (ranging from
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0-2.6 bars). All of these glaciers have gently sloping beds and glacier surfaces within wide valleys. The former accumulation zones of all three glaciers are located in the Central Torridon region, which is itself characterised by gently sloping gradients and chaotic stagnation terrain (Figure 3.6).

The deposition of sediment in the form of chaotic stagnation terrain in the Central region, and in the form of retreat moraine in the valleys can be related to areas of the glacier base characterised by low basal shear stress. Deposition of sediment usually occurs where velocity (and therefore basal shear stress) decreases in the basal zone (Benn and Evans, 1998) and is a result of the decrease in slope gradient of Coire Mhic Nobuil. Comparisons of figure 3.6 and figure 5.4 indicate that recessional moraine has formed where basal shear stress is low in Coire Mhic Nobuil. This is also apparent in Applecross, where an abundance of lateral and terminal moraine complexes (Robinson, 1977; Benn, 1989 and Pers. Obs.) can be observed towards the Coire nan Arr glacier snout (Figure 5.14 and 5.15) where basal shear stress is also much lower (0-1.5 bars).

The importance of glacier thickness (and therefore basal shear stress) to the erosion, transportation and deposition of sediment is particularly apparent in Coire nan Cuileag, a relatively wide and gently sloping valley in Applecross (Figure 5.12). The Coire nan Cuileag glacier was a particularly thin (and possibly fleeting and cold-based) ice body, with very low basal shear stress (0-0.7 bars). Basal shear stress within the longitudinal profile rises slightly where the glacier thickens, approximately 2 km from the margin. Field observations of few, and indistinct glacial geomorphological features in this valley imply that basal shear stress was never high enough to instigate erosion, meaning that subsequent transport and deposition of sediments was unlikely. As discussed in section 6.5.1, it is possible that this valley acted as an accumulation area for snow blown from the south-west. The shallow and open morphology of this valley would have also enabled the relocation of this snow to Coire Attadale. This glacier was probably characterised by
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low velocities, mass turnover and low basal shear stress due to the low gradient of the underlying terrain.

Another characteristic of many of the glacier profiles in section 5.5 (All of the Torridon glaciers and Coire Attadale, Coire na Ba and Coire Sgamhadail), is the slight rise in basal shear stress towards to the snout. It is possible that the rise in basal shear stress can be explained by the thinning and freezing-on of the glacier base (Dowdswell, 1986), a characteristic of modern polythermal glaciers (Glasser and Hambrey, 2001). It is also possible that the high basal shear stress in the marginal area of many of the glaciers relates to the topographic funnelling of the ice towards the snout. This effect would have led to the thickening of the ice and an increase in basal temperature due to the reduction in the conduction of geothermal heat towards the glacier surface (Benn and Evans, 1998). For example, in the case of Coire Mhic Nobuil, ice flowing from Toll a’ Mhadaich Mor, the corrie on the south-facing slopes of Beinn Alligin (Figure 1.2) enters the main stem close to the glacier margin. The increased ice volume and the narrowing of this valley towards the glacier snout would have caused a significant increase in basal shear stress (Thorp, 1991). The convergence of ice from tributaries with the main stem also leads to a significant rise in basal shear stress in Glen Grudie, Torridon (where the ice sourced from Coire Mhic Fhearchair and Coire Ruadh Staca meets the main stem); Coire nan Arr, Applecross (joined by the Coire a' Chaorchain tributary); and the confluence of the Beinn Bhan corrie glaciers also in Applecross. It is also possible that the funnelling of ice through the Bealach nan Arr breach in Applecross would have increased basal temperature and ice velocity. A high basal shear stress would be expected in the breached area as ice accelerated through the narrowing. A heightened basal shear stress can also be observed on both sides of the breach in Coire Attadale (Figure 5.11) and Coire nan Arr (Figure 5.14) to support this.
7.5.2. Contrasts between the Torridon and Applecross ice fields

Despite the contiguous nature of Torridon and Applecross, some contrasts exist, the most obvious being the difference in size of the ice fields. It is probable that topographical factors contributed to the preservation of snow and ice in Torridon, and that the greater build-up of snow in the east could be attributed to the size, quantity and height of the Torridon mountains and the presence of a large accumulation area in central Torridon (Figure 5.1). The link that the Torridon ice cap had to adjacent ice fields to the south and east may have also played a minor role in the growth and preservation of ice in Torridon. It is possible that the proximity to larger ice bodies in the east would have influenced localised atmospheric cooling in Torridon, and that as the ice field grew, it became self-regulated, forming a microclimate cooler than surrounding non-glaciated regions and preventing ablation. Sutherland (1984), states that the relationships between climate, glaciers and topography are scale and time-dependent, in that when a glacier grows, it increasingly modifies its climate. Smaller glaciers are more dependent on climate and topography, whereas larger glaciers increase the area above certain altitudes, changing albedo and having strong local cooling effects.

Given the time-dependence of glacial growth, it is credible that ice survived in Torridon, following the Lateglacial Interstadial. When the Wester Ross Readvance was at its maximum extent, Torridon was located at the junction of two lobes of ice, which filled Lochs Maree and Torridon, whereas (according to previous reconstructions; Figure 1.27) the Applecross Peninsula is believed to have been located on the periphery of the Loch Torridon lobe. Cosmogenic $^{10}$Be dating by Ballantyne et al., (2009) suggests that substantial amounts of ice survived in Wester Ross up to ~14-13 ka. This is tentatively supported by a cosmogenic $^{10}$Be exposure age of 13.4 ± 1.2 ka, from above a ~500m altitude trimline in central Torridon (McCormack et al., 2011). If ice was rejuvenated in regions such as Torridon, it would not have taken as long to establish a coherent ice body before the transition to a colder, drier climate characteristic of the LLS maximum (Hubbard, 1999;
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Golledge et al., 2009), than an ice field that developed from ice free conditions. It is possible that Applecross experienced a period of complete deglaciation prior to the onset of the LLS. This longevity of the Torridon ice field at its maximum extent can be corroborated by the prominence of the geomorphological features which demarcate its extent. Large lateral moraines and distinct weathering limits indicate that the Torridon ice field was well-established for an extended length of time. Geomorphological features in Applecross are less distinguished, and weathering limits more gradual, indicating that the ice field was much more transitory that its Torridon counterpart, possibly relating to a reliance on external climatic controls rather than internal self-regulation.

The ~120m difference in mean ELA between Torridon and Applecross implies a substantial contrast in either climate and/or topographical control on ice build-up over a relatively short distance. Robinson (1977) also calculated mean ELAs for Applecross and the area to the south of Glen Torridon. The results indicated a mean ELA for Applecross was 394m, and 474m for the adjacent area, a difference of 80m over an even shorter distance. This could be a direct result of the growth histories of the ice fields as discussed above, or could be related to the P/T conditions at the time the ice fields were at their maximum extents.

In Torridon, the relatively high ELA reflects either low accumulation or high ablation to maintain mass balance at the LLS maximum. The fact that the Torridon area supported a bigger ice mass despite higher ELA values attests to the importance of topography and altitude to the formation of an ice body. The relatively narrow spread of ELAs (482 ± 25m) of the Torridon ice field, reflects the interdependent nature of its component glaciers and supports the premise that it was self-moderating. The lower mean ELA in Applecross could indicate that accumulation was relatively high, or that ablation was relatively low. Given the presence of small ice tongues in low-lying accumulation areas such as Coire Sgamhadail (Robinson, 1977) it is likely that high accumulation levels were vital to the maintenance of ice
cover in insignificant parts of the Applecross Peninsula, and was a major contributor to the observed contrasts in ELA between the two ice fields.

6.5.3. Palaeoclimate reconstruction based on a global dataset (Ohmura et al., 1992).

Using equations 7.4 and 7.5, palaeoprecipitation estimates of 2311 ± 266 mm a\(^{-1}\) for the mean Torridon ELA and 2656 ± 534 mm a\(^{-1}\) for the mean Applecross ELA, fall within the upper range of the derived contemporary precipitation estimates (1762 ± 230 mm a\(^{-1}\) and 1886 ± 299 mm a\(^{-1}\) for the Torridon and Applecross ELAs), suggesting that mean annual precipitation (Table 5.2) was considerably higher during the Younger Dryas than at present. According to numerical models, at the LLS maximum NW Scotland was characterised by a W-E precipitation gradient of ~40%, and a S-N gradient of 50% relative to today (Hubbard, 1999; Golledge et al., 2009). It is thought that this gradient was a result of efficient scavenging of snow carried on SW air currents, by the main LLS ice cap, centred over the Western Highlands, leaving eastern areas relatively dry (e.g. Benn and Ballantyne, 2005; Finlayson, 2006). In terms of precipitation gradients during the LLS, Applecross can be correlated with the regional S-N and W-E trends, but Torridon is anomalous within this pattern. Sissons, (1979a;1980) demonstrates that the overall ELA pattern in the NW Highlands is one of a marked rise inland from the west coast, where ELAs are typically 400 m or lower, rising to over 1000 m in the eastern mountains such as the Cairngorms (Sissons, 1979b). With a relatively high ELA and low precipitation, this area is more comparable to the drier regions to the east of the main Scottish LLS ice cap, despite its proximity to Applecross.

Taking this into consideration, it is possible that the Applecross ice field (with its lower ELA, higher precipitation value and fragmented appearance) reached its maximum extent when precipitation was at a maximum. The scale of the Torridon ice field suggests that it was much more established than its Applecross counterpart, which was approximately half the size, and much more fragmented. Variations in local topography, and the possibility of the survival of Lateglacial Interstadial ice at
the onset of the LLS, mean that glaciers may not have reached their maximum extent simultaneously and under the same P/T conditions.

6.5.4. Palaeoclimate reconstruction based on a local dataset (Golledge et al., 2009).

The variation of local climate conditions with geographic location and variations in air temperature, wind direction, aspect, topographic shielding can complicate the reconstruction of past climates (Lukas and Bradwell, 2010). It has therefore been suggested that a global dataset (e.g. Ohmura et al., 1992) representing multiple climate regimes, might smooth over any issues involved in the reconstruction of palaeoclimates where modern glaciers no-longer exist. Unfortunately, the use of a global dataset in NW Scotland does not account for the local effects of the reconfiguration of North Atlantic circulation during glacial episodes. Golledge et al., (2009) have accounted for the localised effects of increased continentality at the LLS maximum, caused by enhanced sea ice growth and a shorter ablation season (Denton et al., 2005; Lie and Paasche, 2006).

Glacial growth is also highly dependent on seasonality (Braithwaite, 2008; Hughes and Braithwaite, 2008). Seasonality can vary greatly between glaciated and non-glaciated regions due to variations in annual temperature ranges and precipitation patterns. Denton et al., (2005) linked dominant wintertime seasonality in the North Atlantic during the last glaciation, to a reduction in the strength of the conveyor (section 2.1.2). This enabled winter sea ice to spread across the north Atlantic, enhancing continentality, lowering temperatures and increasing aridity. Given the resultant variation in seasonality between stadials and interstadials and the consequent increase in variation of annual temperature during the YD, it is unlikely that the modern T₃ (the average summer mean temperature for June, July and August used in Ohmura et al., (1992) and annual temperature range reflect those of the YD in Wester Ross.

Ice bodies, once established, can have impacts on local, regional and even global climate. Some large ice sheets are able to buffer themselves from climatic warming,
and lag behind changes in climate in a state of disequilibrium. For example, the high albedo and elevation at the centre of the Antarctic ice sheet, leads to temperatures as low as -70°C (Benn and Evans, 1998). Braithwaite (2003 and 2008) suggests that temperatures adjacent to a glacier surface are significantly lower than the equivalent-altitude temperatures in the free atmosphere. Even valley glaciers can create local temperature anomalies. A glacier of 10-20 km in length may decrease local temperatures by as much as 1.6 – 1.7°C, resulting in a positive mass balance due to the suppression of ablation processes (Golledge et al., 2009).

In addition to this consideration, Golledge et al., (2009) have also accounted for the effects of location of the ice field in question. Despite a high-level of agreement between palaeoenvironmental proxies (Isarin and Renssen, 1999), the direct comparison of data from ice-free and glaciated regions during the YD in Scotland does not account for the effects of localised glacial cooling. As discussed earlier, an increase in albedo, the generation of a micro-climate and associated katabatic effects can lead to a significant decrease in local temperature (Golledge et al., 2009). Contrasts in palaeoclimate between the ice-free SE Scotland and glaciated NW Scotland during the YD are not reflected in $T_3$ estimates, and therefore precipitation estimates calculated from equation 7.2 are likely to be erroneous. A $T_3$ estimate covering the summer months of June, July and August may also not reflect the contrasts in seasonality suggested by Denton et al., 2005 and Lie and Paasche, 2006. This reconstruction introduces the proximity of the palaeoenvironmental proxy in SE Scotland to the study site in NW Scotland, which should lower the temperature (and therefore the precipitation estimate) by accounting for latitude. This is reflected in calculations, where palaeotemperatures are estimated to be 1.4 °C and 1.5 °C lower than those calculated using the Ohmura et al., (1992) method (section 6.5.3) for the Torridon and Applecross ELAs respectively, reflecting the influence of distance from the proxy site in SE Scotland and the influence of local glacier cooling. Palaeoprecipitation values at the Torridon and Applecross ELAs (Table 5.3), are significantly lower than those based on the global dataset (2311 ± 266 mm a$^{-1}$ and 2656 ± 534 mm a$^{-1}$ respectively) in neutral (1256 ± 84 mm a$^{-1}$ in Torridon;
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1506 ± 290 mm a\(^{-1}\) in Applecross) and winter-dominated (1005 ± 67 mm a\(^{-1}\) in Torridon; 1205 ± 233 mm a\(^{-1}\) in Applecross) precipitation scenarios, and at the lower end of the range in the summer-dominated (1758 ± 118 mm a\(^{-1}\) in Torridon; 2109 ± 407 mm a\(^{-1}\) in Applecross) precipitation scenarios. Based on the reconstruction methods of Golledge et al., (2009), summer seasonality palaeoprecipitation totals closely match the ranges calculated for contemporary ELA precipitation totals (Table 5.4; 1762 ± 230 mm a\(^{-1}\) in Torridon; 1886 ± mm a\(^{-1}\) in Applecross), implying that summer type precipitation levels during the YD were similar to today. However, given the effects of increased continentality and the shortened ablation (summer) season in Wester Ross during the YD (Denton et al., 2005; Lie and Paasche, 2006), focus should be directed towards the winter and neutral-type precipitation scenarios. The results for neutral-precipitation and winter-type precipitation scenarios imply that precipitation was actually significantly lower during the YD, highlighting the importance of seasonality and local climate to palaeoclimate reconstruction.

Much of the evidence from this research suggests that the LLS glaciers in Wester Ross were formed within a maritime, temperate, warm-based setting. However, by taking into account the local anomalies on the north-east Atlantic seaboard (increased seasonality, the consequent shortening of the ablation season and the increase in continentality) during the YD, the possibility of a partially cold-based (polythermal) regime cannot be dismissed. It is possible that as the climate warmed towards the end of the YD and the climate progressed towards Holocene conditions (increasingly maritime), the basal thermal regime became dominantly warm, erasing any evidence of a cold/polythermal regime. This is not out of the question, given that retreat moraine not dissimilar to that observed in Torridon, and has been observed in both temperate and polar settings (Evans, 2003).
6.6. Major themes and wider implications

Although the last ice age was forced by long-term orbital (Milankovitch) oscillations, this study has highlighted the fact that short-term internal feedback loops in the North Atlantic played a pivotal role in the last deglaciation. Fluctuations in the latitude and strength of the Gulf Stream were particularly influential on the climate of the NE Atlantic seaboard during deglaciation (Broecker et al., 1989; Stocker et al., 1992). The position of the oceanic polar front and associated sea ice was intimately associated with the migration of the Gulf Stream (Ruddiman et al., 1977), affecting the precipitation and temperature conditions of the coastline of NW Scotland.

The IPCC report (Alley et al., 2007) stated “…some exceptional results indicate the complexity of both regional to local-scale climate and respective glacier regimes.” The complexity of glacial response to these climate conditions is particularly apparent in the Wester Ross region of NW Scotland, where adjacent ice bodies demonstrate contrasting responses to the climate of the transition back to glacial conditions during the Younger Dryas. The response of the Torridon and Applecross ice fields to climate is also complicated by local boundary conditions, including the surrounding topography, aspect and the effects of glacially-induced cooling.

The glacial response to variables in climate and boundary conditions is of particular interest at a time when many mid-latitude mountain range glaciers are in retreat. The major drive of this retreat is the increase in global temperature associated with contemporary climate change, but the relationship between glacier and temperature is complicated by variations in glacier geometry and aspect (Oerlemans, 2005). Studies of formerly glaciated landscapes can provide a basis for the prediction of the effects of climate amelioration on the deglaciation of ice bodies in a variety of settings. In this instance, the investigation of ice fields in a location which varied between continental and maritime, according to the position of sea ice on the NE Atlantic seaboard, is analogous to glaciers in both high and mid-latitude settings.
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6.7. Recommendations for future research

As the techniques and technologies used for the mapping of formerly glaciated landscapes advance and improve, reconstructions of past glacial climates will become increasingly accurate. In particular, remote sensing imagery such as Google Earth and NEXTMap have made the study of vast areas of post-glacial terrain a much more straightforward process, providing a wealth of information instantaneously.

Advances in current ground-based LiDAR technology would make terrestrial LiDAR particularly useful within the scope of glacial geomorphological mapping. The LiDAR model in this study provided detailed information regarding most of the glacial-geomorphological features in Coire Mhic Fhearchair. What is lacking in resolution and clarity will be improved upon as LiDAR technology is upgraded. Developments in scanning equipment are leading to the extension of the acquisition range from several hundred metres to several kilometres. Software is also being developed to enable easy post-processing, manipulation and display of digital models. In remote areas, it is essential that the issue of energy supply is addressed. A three day expedition to Coire Mhic Fhearchair required four car batteries to charge the scanner and laptop, requiring extra man-power and an awareness of the limits of the power source. Advances in the current technology would make terrestrial LiDAR particularly useful within the scope of glacial geomorphological mapping over areas >2km² and for extended period of time. LiDAR does not however entirely replace the need for field visits, as the investigation of the subsurface is of vital importance to the interpretation of the origins of glacial-depositional features.

Refinements in the processes involved in the chemical processing of rock samples, for cosmogenic analysis should improve the accuracy of dating techniques. The CRONUS calculator is constantly being updated to account for variations in cosmogenic nuclide production rate according altitude and temporal fluctuations.
Chapter 6. Discussion

(Balco, 2008). Reductions in required sample size and the ability to isolate $^{10}$Be from an increasing array of rock types (e.g. Kong et al., 1999) mean that exposure dating techniques are continuously being developed to be applicable to an expanding range of scenarios (Gosse and Phillips, 2001).

Of particular interest are the issues surrounding glacial erosion and in particular the debate surrounding the origins of glacial trimlines and corresponding weathering contrasts, specifically whether trimlines represent absolute ice extents or an englacial boundary. Exposure histories are complicated by the potential of the existence of cold-based, passive ice above trimlines, which does not ‘reset’ the cosmogenic clock, thus providing anomalously old exposure ages. As a result, it is uncertain whether ice-surface reconstructions based on trimline evidence, and therefore Equilibrium-Line Altitude (ELA) and palaeoclimatic reconstructions are reliable. Given more time and funding, this research would benefit from the collection of more samples transecting the weathering limits in Torridon and Applecross. Currently, the stand-alone, pre-Younger Dryas exposure ages indicate contrasting exposure histories, but cannot provide unequivocal evidence for exposure during a particular episode of the last deglaciation, due to the uncertainties surrounding the origins of trimlines. This would reveal insights into the survival of ice in the Wester Ross region during deglaciation, and could further understanding of the debate surrounding the origins of trimlines and the extent of warm vs. cold-based ice during the LLS. Further investigations into the lateral extent of the LLS (through the collection of samples from boulders on or adjacent to terminal moraine complexes in Torridon and Applecross) would provide an additional constraint for the timing of deglaciation in this region.
Chapter 6. Discussion
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Conclusions
Conclusions are summarised below for each question posed in section 1.9.

Question 1: What can the glacial geomorphological signatures of Wester Ross tell us about the last deglaciation in NW Scotland?

At the onset of deglaciation it is highly plausible that the continental climate of the LLS maximum (caused by the southward expansion of sea ice along the NE Atlantic seaboard) was replaced with a relatively wet and warm maritime climate, as the oceanic polar front migrated northwards away from this region of NW Scotland (Benn, 1997). Glacial geomorphological and sedimentological studies of a ~45 km$^2$ area imply that the deglaciation of the Torridon ice field took place in three distinct phases.

The first phase of deglaciation was characterised by a highly dynamic glacial environment in the lower valleys of Torridon, balanced by winter accumulation and summer ablation and maintaining a position close to climatic equilibrium. The lateral and terminal moraines formed during this stage are composed of sediment excavated during sequential glacial episodes. This sediment was sourced subglacially from the high corries (which were subject to extensive erosion during the LLS), and paraglacially, from the free faces that line the walls of the valleys. Within the maximum extent of the LLS, a series of 84 prominent retreat moraines formed in Coire Mhic Nobuil. These moraines probably formed as a response to variations in precipitation-driven winter time accumulation. The presence of large and bifurcated moraine ridges indicate that the ice front readvanced periodically, pushing and over riding previously deposited moraines. This is supported by evidence of subsurface glacio-tectonic displacement (e.g. Benn, 1992; Reinardy and Lukas, 2009), and supports the theory of oscillatory decay in the lower valleys.

This phase of retreat took place over ~300 years and lagged behind the rapid ~40 year climatic amelioration observed in the NGRIP data (Taylor et al., 1997), reaching its final stages later than optimal high-resolution models of the main
Chapter 7. Conclusions

Scottish ice cap (Golledge et al., 2008) suggest. This can be attributed to the retreat of sea ice to the north and the exposure of the NW Scottish seaboard to Gulf-Stream dominated maritime conditions (Benn, 1997).

As climate amelioration accelerated after ~11.6 ka (Brooks and Birks, 2000), the local equilibrium-line gradually rose. During this second phase of deglaciation, ice remaining in the central area downwasted in-situ to create a chaotic landscape (Benn et al., 1997). Finally, as the snow line achieved an altitude above the highest mountains in Torridon (Bennett and Boulton, 1993a), cosmogenic $^{10}$Be exposure ages indicate that ice retreated into the high north-facing corries after approximately 11.7 ka. Corrie glaciers retreated passively, depositing a series of fluted moraines onto the exposed bedrock below. These glaciers were probably polythermal or cold-based despite rapid climate warming, due to their decreased thickness and the north-facing aspect of their corrie locations.

Question 2: Can terrestrial Light Detection and Ranging (LiDAR) be applied to the mapping of glacial-geomorphological features, and how does it compare to relatively traditional mapping methods?

A variety of mapping techniques were used to compile the final geomorphological map (appendix 2), including field mapping, aerial photography and DEMs. The potential for each technique to promote feature recognition was assessed against LiDAR, with LiDAR producing similar results to field mapping. In both instances, all of the glacial-geomorphological features could be observed, but in some cases it was not always possible to collect accurate data. For example, when using LiDAR, it can be difficult to see low-relief features such as moraines, where there is no obvious colour contrast and in the field, over-steepened corrie slopes make the mapping of debris fans a logistical impossibility. DEMs can be post-processed to highlight variations in slope, and are therefore useful for collecting information on larger-scale features such as steps in bedrock. Aerial photographs provide clear contrasts where shadow does not obscure features. These contrasts are particularly
apparent in Coire Mhic Fhearchair, where the bright white Cambrian quartzite contrasts to the underlying Torridonian sandstone.

Terrestrial LiDAR scanning of Coire Mhic Fhearchair, Torridon has led to the production of a detailed 3D model. The ‘virtual reality’ nature of 3D models such as this enables the visualisation of features from a variety of perspectives, and the accurate measurement of vertical and horizontal distances, as well as dip and strike measurements of the surrounding bedrock. Data can be stored indefinitely, reducing the need for regular return field visits to remote and inaccessible regions for reassessment or error checking. The integration of the scanned image with co-located high-resolution, digital colour photography provides clear, high-quality imagery, negating the need for the complicated georeferencing and draping processes required when using DEMs and aerial photography (chapter 4.2). Problems regarding shadowing in aerial photography (e.g. Mitchell, 2006; Smith and Wise, 2007), and azimuth biases encountered in DEMs (e.g. Smith and Clark, 2005), can be minimised through the collection of data from multiple locations. The ability to adapt the scanning resolution to suit the surface from which information is being collected enables detailed coverage of glacial-geomorphological features at all scales, up to a resolution of ~1mm (Bellian, 2005). This allows the interpretation of features in the context of their surroundings, and can provide insights into the glacial-geomorphological history of the area.

Despite the advantages of using ground-based LiDAR as a mapping technique, glacial geomorphological mapping should be undertaken using multiple techniques. By combining traditional and contemporary techniques, the advantages of each can be exploited to produce an accurate and informed final map. In the instance of Coire Mhic Fhearchair, the map of glacial-geomorphological features compiled using all four methods (Figure 3.21) provides a detailed glacial history of the corrie, and can be used to infer a chronology for central Torridon. A palimpsest surface has been revealed, providing evidence of glacial reorganisation during the Lateglacial. Field mapping and LiDAR imagery from the north-eastern edge of the corrie have
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revealed undulations in the bedrock which indicate that the bedrock was eroded by ice flowing from east to west. This is likely to have occurred when local topography was overwhelmed by the movement of ice from the main Scottish ice-shed in the east, towards the lowlands in the west during the Dimlington Stadial (Ballantyne and Sutherland, 1987; Ballantyne et al., 1998; Bradwell et al., 2007). Superimposed on this landscape is a series of debris stripes, similar to those observed in polar regions (Glasser and Hambrey, 2001). These fluted features are composed solely of angular boulders of quartzite, and indicate a period of corrie glaciation during the late LLS, when ice spilled out of Coire Mhic Fhearchair towards the valleys in the east. A significant lack of retreat moraine suggests that this corrie was not a prominent site of active decay at the end of the LLS, and that glaciers did not readvance following final LLS deglaciation.

Question 3: What insights can cosmogenic $^{10}$Be isotope dating provide into the deglacial history of Wester Ross?

Eight cosmogenic $^{10}$Be exposure ages for the central Torridon and Applecross regions yield six mean Younger Dryas ages and two outliers. The six YD ages (samples WRC 1a, 1b, and 2 in Torridon, and 4, 5 and 7 in Applecross) were collected from glacially-scoured bedrock. The two outlying samples were collected from relatively weathered rock surfaces, and yielded a mean Dimlington Stadial exposure age within Applecross (WRC6), and a mean Lateglacial Interstadial age in Torridon (WRC3).

The $2\sigma$ difference between exposure ages for WRC6 and WRC4 in Applecross, and the weathering contrast (supported by Schmidt hammer readings) between these two sites, implies that the knoll (WRC6) on Applecross could be a Dimlington Stadial surface. It is therefore possible that this knoll and Beinn Bhan (Figure 6.8) were both exposed as nunataks during the late Dimlington Stadial. The exposure age at WRC3 in Torridon is $1\sigma$ different from the exposure age at WRC2, thus implying that the low altitude trimline on Am Beacan (Appendix 2) is a LLS ice
limit. The internal uncertainty (0.5 ka) of this exposure age would suggest that this site was exposed within the Lateglacial interstadial, and would make it directly comparable to exposure ages from the WRR moraine (Bradwell et al., 2008; Ballantyne et al., 2009; Ballantyne, 2010). However, systematic uncertainty (± 1.3 ka) around this stand-alone date, suggests that the site could also have been exposed during the YD. Large statistical uncertainties surrounding these stand-alone samples and the distinct possibility of cold, passive ice cover (Balco et al., 2008), means that both outliers should be considered with caution. Further sampling is required to achieve any definitive conclusions.

The YD exposure ages indicate that >2m of erosion occurred during the LLS glaciation of these marginal regions, thus implying a warm-based regime and high basal meltwater velocities. The mean of the six YD exposure ages (11.8 ± 1.1ka) suggests that a significant amount of ice was present in both regions towards the very end of the LLS, despite a significant and rapid (~40 year) warming observed in the NGRIP record at the LLS-Holocene transition, and the non-correlation with optimum Scottish climate models for this time (e.g. Golledge et al., 2009; Lukas and Bradwell, 2010). YD cosmogenic exposure ages indicate that the Torridon and Applecross ice fields fit the more vertically extensive LLS reconstructions of Boulton and Bennett (1993) and Robinson (1977) respectively. It is concluded that a sensitive ice-field response to the maritime-dominated (i.e. increased precipitation) climate of Wester Ross promoted extensive ice cover and survival in Torridon and Applecross during the LLS.

*Question 4: What was the climate like in Wester Ross during the most recent glacial advance during the Younger Dryas (~12.9 – 11.5 ka)?*

Palaeo-ice surfaces have been reconstructed for the Torridon and Applecross ice fields, using a combination of field mapping, aerial photography, DEMs and previously mapped ice extents. The Torridon ice field (~ 100 km²) was radial in form and was composed of five prominent valley glaciers (15 – 19 km²). The
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Applecross ice field (~43 km$^2$) was composed of four main valley glaciers (~3 – 15 km$^2$) joined by breaches, and seven separate valley and corrie glaciers (~0.2 – 5 km$^2$). The greater extent of the Torridon ice field relate to the height and scale of the mountains in this region, the presence of a large central accumulation area, the link with ice fields to the south and east and possibly the existence of ice at the onset of Younger Dryas cooling, allowing a coherent and established ice field to form by the time the LLS maximum was achieved. The size of the ice field and the proximity of Torridon to the main ice cap suggest that it developed a self-regulating microclimate, favourable to the preservation of ice.

In contrast, ice surface reconstructions and geomorphological observations imply that the Applecross ice field was much more responsive to local climate. The relatively small size of this ice field and its component glaciers and lack of valley connectivity produced a wider-spread of ELA values, and indicate that self-regulation would have been minimal. The relatively sparse distribution of glacial depositional features indicate that this ice field was more fleeting, and probably disappeared quickly as rapid climate warming took place at the end of the LLS. Basal shear stress calculations also indicate that parts of the Applecross ice field (in particular the Coire nan Cuileag glacier) were thin, fleeting and had little impact on the landscape. Low basal shear stresses are also apparent in the north-flowing Torridon glaciers which had gently-sloping beds. Higher basal shear stress values were observed in both Torridon and Applecross, where gradients increased and tributary glaciers were confluent with the main stems, causing a thickening of the glacier profile.

The ELAs for both ice fields have been calculated using an Area-Altitude Balance Ratio of 2, yielding mean ELAs of 482 ± 25m and 361 ± 120m for Torridon and Applecross respectively. Inter-ice field variation in ELA and glacier morphology can be attributed to topographical disparities and the localised influence of dominant south-westerly snow blow (e.g. Ballantyne, 1989; 2005; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010). The accumulation of snow behind N-S ridges is
supported by the asymmetrical pattern of the glaciers, which developed in these valleys. The formation of independent low-lying glaciers in Applecross was particularly dependent on the north-eastwards transference of snow from the extensive (~100km²) area of low-lying snow-blow area to the west, and is probably the cause of the relatively low ELA (Sissons, 1979a; 1980).

Using empirical values from a global dataset (Ohmura et al., 1992), average annual Younger Dryas palaeotemperature and palaeoprecipitation values were estimated for the Torridon and Applecross ice fields. Temperature at the ELA (derived from an equation in Benn and Ballantyne, 2005) was estimated as 4.9 ± 0.2ºC and 5.7 ± 0.2 ºC respectively. Precipitation was found to be 2311 ± 266 and 2565 ± 534 mm a⁻¹ respectively, suggesting a cooler and wetter climate than today, concurrent with reconstructions from other maritime regions of NW Scotland (e.g. Ballantyne, 1989; 2002; 2006; 2007). The palaeoprecipitation conditions of the Applecross ice field concur with regional W-E and S-N precipitation gradients reconstructed using the same dataset and the northwards decline in ELA in NW Scotland (Sissons, 1979b). In contrast, Torridon is anomalous within these regional trends. It is possible that ice in Torridon and Applecross did not reach their maximum extents simultaneously, and each reflect a response to different climatic episodes of the LLS.

However, these reconstructions do not consider the effects of variations restricted to Scotland during the Younger Drays. The proximity of the Wester Ross region to the NE Atlantic seaboard introduces a variable in the form of the North Atlantic circulation, which reorganises itself between stadials and interstadials, thus influencing the position of the Gulf Stream, the extent of sea ice and therefore the predominant climate (Ruddiman et al., 1977; Ruddiman and McIntyre, 1981). Glaciological models (Hubbard, 1999; Gollledge et al., 2008) indicate that the LLS maximum was characterised by a drier climate in NW Scotland as winter sea ice expanded along the NE Atlantic margin. An enhanced seasonality (Denton et al., 2005; Lie and Paasche, 2006) and a resultant short ablation season, led to an
increase in continentality. The prominence of glacial geomorphological features (particularly in Torridon) demarcating the outer extent of the LLS, implies that ice was maintained by this cold and dry climate at its maximum limit for an extended amount of time. Golledge et al., (2009) based a new set of palaeoclimate equations on these assumptions. These equations also take into account the local effects of glacier cooling (Khodakov, 1975), unaccounted for in the global dataset. Palaeotemperature and palaeoprecipitation have also been calculated based on these equations, yielding lower values of $3.5 \pm 0.6^\circ$C and between $1005 \pm 67$ mm a$^{-1}$ and $1758 \pm 118$ mm a$^{-1}$ for the Torridon ELA and $4.2 \pm 0.6$ °C and $1205 \pm 233$ mm a$^{-1}$ to $2109 \pm 407$ mm a$^{-1}$ for the Applecross ELA, reflecting a more continental climate than observed in the Ohmura’s (1992) global model.
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Appendices
### Appendix 1. Quaternary Geochronology

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<th>Age (ka BP)</th>
<th>Oxygen Isotope Stage</th>
<th>Quaternary Chronostratigraphy</th>
<th>Britain</th>
<th>NW Europe</th>
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<td>24</td>
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<td>Dimlington Stadial</td>
<td></td>
<td>Peniglacial</td>
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<td>15</td>
<td>Devonian</td>
<td>Middle Devonian</td>
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<td>12</td>
<td>Devonian</td>
<td>Late Devonian</td>
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<tr>
<td>10</td>
<td>Devonian</td>
<td>Late Glacial</td>
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<tr>
<td>5a, 5b, 5c, 5d</td>
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<td>Brimpton Interstadial</td>
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<td>Ipswichian</td>
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<td>Eemian</td>
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<td>Devonian</td>
<td>Sourlie interstadial (Upton Warren)</td>
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Silts & clays: Particles of <0.063mm

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